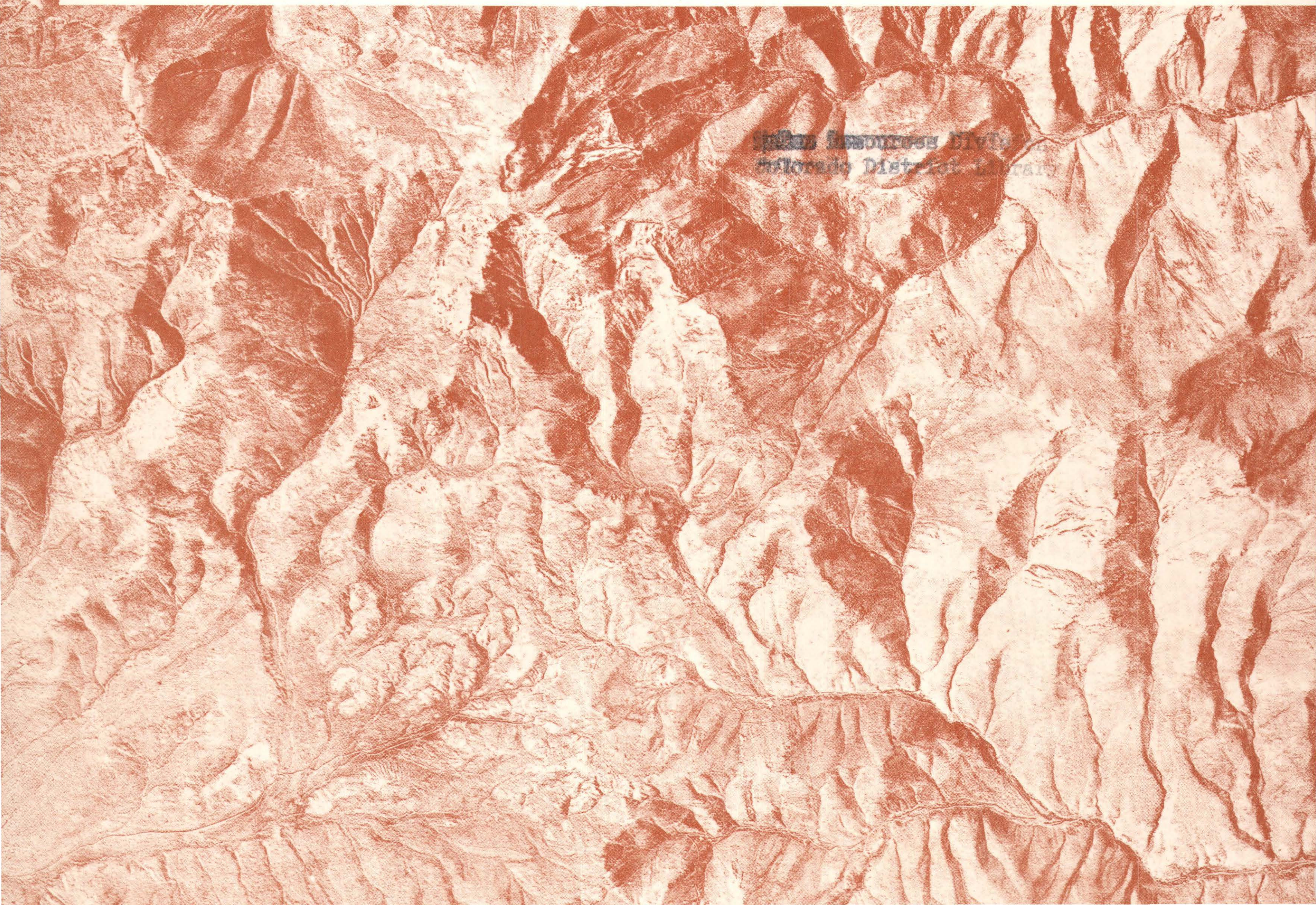
TECTONIC AND IGNEOUS GEOLOGY OF THE NORTHERN SHOSHONE RANGE, NEVADA

GEOLOGICAL SURVEY PROFESSIONAL PAPER 465



Tectonic and Igneous Geology of the Northern Shoshone Range Nevada

By JAMES GILLULY *and* OLCOTT GATES

With sections on

GRAVITY IN CRESCENT VALLEY, *by* DONALD PLOUFF
and ECONOMIC GEOLOGY, *by* KEITH B. KETNER

GEOLOGICAL SURVEY PROFESSIONAL PAPER 465

Prepared in cooperation with the
Nevada Bureau of Mines

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1965

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

The U.S. Geological Survey Library catalog card for this publication appears after index.

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402

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By James Gilluly and Olcott Gates

ABSTRACT

Part of the northern Shoshone Range and its bordering valleys are shown in the Mount Lewis and Crescent Valley quadrangles of the U.S. Geological Survey's Topographic Atlas. A very small part of the Cortez Mountains extends into the Crescent Valley quadrangle and is also described.

As a representative part of the Basin and Range physiographic province, much of the area is masked by Quaternary alluvial and playa deposits; exposed rock is confined to the mountains, which rise from 3,000 to 5,000 feet above the adjacent valleys.

The paramount feature of the geology is the Roberts thrust, a major structural element of the Cordillera, which here has its westernmost known exposures in this latitude. This thrust brings into contact thick sections of rock of early and middle Paleozoic age which contrast dramatically in lithology. One sequence, here considered allochthonous (transported), includes rocks of Ordovician, Silurian, and Devonian age, and is almost entirely siliceous. It consists of sandstone, quartzite, shale, greenstone, chert, and almost negligible lenses of limestone. This sequence has been carried by the thrust over a second sequence of rocks of Cambrian, Ordovician, Silurian, Devonian, and possible Mississippian age of notably contrasting—almost wholly carbonate—composition. The carbonate sequence greatly resembles comparable sections at Eureka and elsewhere to the east in the Great Basin, and is considered to be practically in its original area of deposition (autochthonous and parautochthonous). The strongly contrasted sequence of siliceous rocks obviously has been brought into the area by movement on the Roberts thrust; the area of original deposition must surely have lain far to the west of the site of the present Shoshone Range.

The carbonate sequence, considered as the westernmost known representative of the autochthonous rocks, includes the following Cambrian formations: the Prospect Mountain Quartzite of presumed Early Cambrian age (perhaps 300 ft exposed), a possible thin representative of the Pioche Shale of Early and Middle Cambrian age, the Eldorado Dolomite of Middle Cambrian age (about 500 ft exposed), and a newly recognized formation—the Shwin—of limestone, shale and greenstone, also of Middle Cambrian age and estimated to have an exposed thickness of 1,500–2,000 feet. Fossil evidence of ages is confined to the Shwin Formation; the other formations are correlated with the typical sections in the Eureka district wholly on lithologic criteria.

Ordovician rocks of the autochthonous sequence include the Eureka Quartzite (perhaps 150–200 ft exposed) and Hanson Creek Formation (about 500–700 ft thick) of Middle and Late Ordovician ages, respectively. No fossils have been found in the Eureka but a representative collection from the Hanson Creek permits confident correlation with the typical exposures in the Roberts Creek Mountains 25 miles to the southeast.

Silurian rocks in the autochthonous sequence are represented only by the Roberts Mountains Limestone, though in the westernmost window this formation is notably silty, a fact that perhaps suggests transition to the siliceous western facies. About 600 to 1,000 feet of beds referred to this formation are exposed.

Devonian limestones several hundreds of feet thick as exposed cannot be confidently assigned to any of the formations recognized to the east. They are exposed in the Crescent Valley quadrangle, near Gold Acres, together with several hundred feet of pink-weathering shale which is considered to represent the Pilot Shale of more easterly localities. Farther east, the Pilot Shale includes rocks of both Devonian and Mississippian ages, but the only fossil here found is referred to the Devonian.

The siliceous facies (allochthonous) includes the Harmony Formation, a remarkable arkose of Late Cambrian age, here represented in only thin thrust slices. Belonging to a different depositional sequence is the Valmy Formation of Early, Middle, and Late Ordovician age. This formation is here several times as thick as in the type area, 15 miles to the northwest, and must be at least 12,000, perhaps 30,000 feet thick. As in the type area, it includes much sandstone, siltstone, shale, quartzite, and chert, as well as considerable thicknesses of pillow lavas, fragmental volcanics, and other greenstones.

Recognized Silurian rocks of the siliceous facies are here referred to a new formation, the Elder Sandstone, which consists largely of feldspathic sandstone and minor amounts of thin-bedded chert and shale. It is 2,000 to 4,000 feet thick. A few graptolites enable a Silurian age assignment. The graptolites of the Valmy and Elder Formations are reviewed in a separate section by R. J. Ross, Jr., and W. B. N. Berry.

The last formation of the siliceous facies is a thick chert, here named the Slaven Chert. It is almost wholly very dark gray or even black, about 4,000 feet thick, and contains a sparse fauna, largely of ostracodes, that suggests correlation with the Middle Devonian part of the Traverse Group, of Michigan.

Stratigraphic and structural evidence suggests that the Roberts thrust took place in Early Mississippian time. Both here and in the Battle Mountain area to the northwest, the Battle Conglomerate rests in depositional contact on Ordovician rocks of the western facies in such a way as to demonstrate deep erosion of strongly deformed rocks. This deformation, called the Antler orogeny by Roberts, is definitely earlier than Middle Pennsylvanian.

Several isolated bodies—the thickest about 400 feet thick—are referred to the Battle Conglomerate of Middle Pennsylvanian age because of their marked resemblance to the characteristic rocks of the type section across the Reese River Valley to the northwest. In one or two localities the Battle is conformably overlain by the Antler Peak Limestone, just as in the type locality. The Antler Peak Limestone, of Late Pennyslvanian

and Early Permian age, is about 800 feet thick. A second formation of Middle Pennsylvanian to Early Permian age, the Havallah, is represented by several small thrust masses. As at Antler Peak, the Havallah is in fault contact with the Antler Peak and presumably was not part of the same depositional sequence.

In the Cortez Mountains a second facies of Upper Paleozoic rocks rests in fault contact on the Valmy Formation. Displacement on the fault is probably not great, so the rocks are probably nearly in place. They include dolomite about 200 feet thick, overlain by about 400 feet of conglomerate and sandstone, which is in turn overlain by several hundred feet of interbedded sandstone and limestone. Fossils of Late Pennsylvanian age were collected from the basal dolomite and upper limestone. These rocks are assigned to a new formation, the Brock Canyon.

A conglomerate unconformably overlies the Antler Peak Limestone in the Shoshone Range, is several hundred feet thick, and contains boulders derived from the Antler Peak. It has yielded a few indigenous molluscan remains as well as the fossils from boulders, but they are not diagnostic of its age. The tentative reference here made to the China Mountain Formation of Early Triassic age is based wholly on stratigraphic position and lithologic resemblance.

No rocks of Jurassic or Cretaceous age have been recognized.

The Cenozoic is represented by intrusive rocks of two cycles, though both are of similar composition (quartz monzonitic to granodioritic). The earlier stocks, possibly of Eocene or Oligocene age, were deroofed prior to the volcanic eruptions associated with the later cycle. The later cycle, of Miocene and Pliocene age as determined from vertebrate fossils in related tuffs, involved eruption of large volumes of quartz latite tuffs and related rhyolites. Several volcanic vents of this cycle have been recognized; the rocks filling them closely resemble the tuffs petrographically. These vents could have supplied the tuffs, but the abundant north-south dikes could also have served as feeding fissures at levels now removed by erosion. A few beds of gravel are interbedded with the tuffs and with a few flows of hornblende andesite and basalt.

The volcanic vents are largely filled by intrusive breccia, abundantly charged with accidental inclusions of wallrocks and cognate inclusions of related volcanics. Several vents contain downdropped blocks of sedimentary rocks; this proves that the magma column rose and fell alternately during eruption. Presumably such a "pumping" action was a major factor in producing the breccia fill.

Unconformably overlying the quartz latite tuffs and interbedded with their associated gravels on the east flank of the Shoshone Range are large flows of basaltic andesite. These cover much of the east side of the range for a long distance within and to the north of the map area. The eastward tilt of these rocks at angles of 5° to 10° makes a conspicuous dip slope, conditioned by Pliocene and Pleistocene faulting. Across Crescent Valley, the steep scarp of the Cortez Mountains exposes many dikes of dolerite which could have fed the lava flows, for similar lavas also form a dip slope on the southeast flank of the Cortez Mountains and extend for many miles to the east and north.

Little attention was given the Quaternary rocks of the area and they are mapped under a single rubric. They include colluvium on the steep mountain slopes, terrace and alluvial gravels along most streams, alluvial fans fronting the ranges, and minor playa and dune deposits.

The main general geologic interest lies in the tectonics of the area. As indicated by the stratigraphy and by the associations of the different rock units, two major orogenies, both involving thrust faulting, have preceded the Basin Range faulting to which the present relief of the area is primarily due.

The earlier episode was the one during which the Roberts thrust was formed. It is inferred from regional geology that this fault probably originated during Early Mississippian time. The Roberts thrust is exposed in windows, five of which are in the map area, over a belt at least 50 miles wide. The fault separates facies of comparable age and strongly contrasting lithology throughout the belt, so that the upper plate must have traveled more than 50 miles. The rocks both below and above the fault are highly disturbed, though very little metamorphosed. Conglomerates indicative of surface travel of the thrust have not been recognized either in this or nearby areas, and the great deformation, both of the upper and lower plates and of the thrust surface itself, suggests that, at least in this area, the thrust was wholly subsurface and at a depth of many thousand feet.

The fault is highly folded and has a present relief of at least 10,000 feet. Relations between the curvature of the fault and its branches into the hanging wall strongly suggest that much of this warping of the thrust surface took place during the thrust movement. Detailed sections drawn normal to and parallel to the strike enable reasonable inferences as to subsurface structure, though of course only as to major elements in such a complex. A generalized structure-contour map of the Roberts thrust is presented.

An overturned segment of the thrust is separated from an openly folded segment to the south by a remarkable transcurrent fault about 10 miles long. The structures flanking this fault are quite independent throughout this distance.

The rocks of the upper plate are highly sheared and deformed, so that several minor thrust sheets within that plate are folded isoclinally and some are recumbent. The lower plate is also much faulted but its exposures are so patchy that little system can be recognized in the arrangement of the blocks. Metamorphism has been slight; a few phyllite exposures represent the most severe.

Faulting of comparable complexity but probably less tectonic travel involved rocks as young as the supposed China Mountain(?) Formation of Early and Middle(?) Triassic age. This orogeny is here named the Lewis orogeny. A single thrust sheet of this episode extends across the exposed edges of four or five lower thrust sheets (attributed to the Roberts thrusting) and leaves fault outliers over an area several miles beyond the limit of the main post-China Mountain(?) thrust sheet. Though faults of this episode are somewhat folded, none are as greatly contorted as the Roberts fault. No date more precise than post-Early Triassic(?) and prior to the intrusion of the early Tertiary stocks can be given to this younger thrusting episode. It might be Late Triassic, Early or Late Jurassic, Early or Late Cretaceous or even Eocene in age, for closely bracketed orogenic episodes of each of the listed ages are known within a 200-mile radius. Because of the scant data available, it seems fruitless to speculate at present.

Although considerable structural disturbance must have accompanied each of the igneous cycles, only a few faults, peripheral to the vents and stocks, are recognizable. The country was deeply eroded after the later thrusting before the volcanic episode began.

After the eruption of the basaltic andesite flows of the east flank of the Shoshone Range, movement on normal faults began to block out the range. The northwest-facing fronts of both the Shoshone Range and the Cortez Mountains are fault scarps that record many thousand feet of normal fault movement. Smaller faults of like trend dissect the east slope of the Shoshone Range internally. Much of the present relief of the ranges is due to uplift along these faults. This conclusion from the geologic relations is strongly reinforced by the gravity survey of Crescent Valley discussed by Plouff.

Plouff reports that 47 stations in and bordering Crescent Valley between Indian Creek (Shoshone Range) and Four Mile Creek (Cortez Mountains) were occupied by gravimeter. The Bouguer anomalies computed from these observations suggest a deep fill of low-density material in Crescent Valley, somewhat thicker to the east of the valley axis, precisely as was inferred from projecting the surface geology. Different values are derived for the thickness of the fill depending on the values chosen for the density contrast between valley fill and the bedrock of the adjoining ranges. The density difference of 0.6 used in Plouff's example is probably conservative if the probable increase in consolidation of the valley fill at depths of several thousand feet is considered. This conservative assumption yields an estimated thickness of valley fill of 5,600 feet. A second, perhaps more reasonable, solution, which assumes increasing compaction with depth, gives 7,000 feet. Geologic evidence suggests at least 6,000 and perhaps as much as 10,000 feet of valley fill.

The economic geology of the area involves the currently active gold mines at Gold Acres and Granite Mountain, the sporadically active barite mines of the Shoshone Range, and the now inactive antimonial silver mines of the Kimberly (Hilltop), Dean, Campbell, and Bullion districts. Although most of the inactive mines are no longer accessible, an attempt has been made to summarize available information both on the mines and on the mineralogy of their ores. The barite deposits of commercial size seem to be confined to the Slaven Chert, perhaps because of that formation's susceptibility to replacement.

INTRODUCTION

GEOLOGIC INTEREST OF THE AREA

Merriam and Anderson's discovery (1942) of the marked facies contrast across the Roberts thrust in the Roberts Creek Mountains between Cortez and Eureka (see fig. 1) suggested that this fault is a major orogenic feature. At that time the age of the fault was assumed to be Laramide. A few years later, however, it became clear from the work of Ferguson, Muller, and Roberts (Ferguson and Muller, 1949; Ferguson, Muller, and Roberts, 1951a, 1951b; Roberts, 1951; Muller, Ferguson, and Roberts, 1951) that more than one orogeny was recorded in the country to the northwest of the Roberts Creek Mountains and that at least one of them was of late Paleozoic age. It became, therefore, a matter of compelling interest and of pressing importance to an understanding of the geologic history of this part of the Cordillera to determine the extent of each of the thrust systems recognized. The northern Shoshone Range, whose geology is the subject of the present report, is strategically located to supply information on this subject. The U.S. Geological Survey therefore authorized and undertook the study, to which we were fortunate enough to be assigned.

LOCATION AND ACCESSIBILITY

The Mount Lewis and Crescent Valley quadrangles lie in northern Lander County, Nev., in the north-central part of the State. They comprise an area of about 455 square miles between the meridians 116°30′ and 117°00′ W. and the parallels 40°15′ and 40°30′ N. (See fig. 1.)

As seen in figure 2 the only paved road in the project area is Nevada Highway 21 from Beowawe to Gold Acres. Gravel roads and unimproved tracks, however, lead into most of the major canyons of the Shoshone Range and four of them cross passes in the divide. A car can be driven to within 2 or 3 miles of all points in the area, except for places in Trout and Crippen Canyons on the west and Fire Canyon on the east. The north and west slopes of the Shoshone Range and Cortez Mountains are generally extremely steep and rugged and are practically accessible only on foot, but the east and south slopes are mostly smoothly rounded and can be reached by four-wheel drive vehicles.

PHYSICAL FEATURES

The Mount Lewis and Crescent Valley quadrangles embrace a topographically representative part of the Basin and Range province. The area includes segments of the Shoshone Range and Cortez Mountains, of the intervening depression of Crescent Valley, of Carico Lake Valley to the south, and of the Reese River Valley to the west. (See fig. 2 and pl. 1.)

The Shoshone Range is one of the longest ranges in Nevada, extending from the valley of the Humboldt River near Dunphy for nearly 200 miles to the south-southwest. The area here described is near the north end of the range. The Cortez Mountains, though extending about 30 miles only along a north-northeast trend, form a link in a linear uplift that extends, with only short interruptions, south through the Simpson Park and Toquima Ranges almost to the latitude of Tonopah. Only a very small part of the northwest-facing scarp of this range lies within the area.

Crescent Valley is an undrained depression separated by a low alluvial divide from the Humboldt River, a few miles to the north. A southern spur of the Shoshone Range and an extreme northward-reaching spur of the Toiyabe Range separate the south end of Crescent Valley from Carico Lake Valley to the west; Carico Lake (playa) overflows to Crescent Valley during wet weather only. Only a few square miles of the area drain to Carico Lake. The west and north slopes of the Shoshone Range here drain to the Humboldt River, either directly or by way of the Reese River.

Altitudes in the map area range from 9,680 feet at the summit of Mount Lewis to a few feet less than 4,600 feet in the Reese River Valley. Both the Shoshone Range and Cortez Mountains are bounded on their northwest sides by steep, relatively smoothly faceted scarps. These smoothly linear scarps are demonstra-

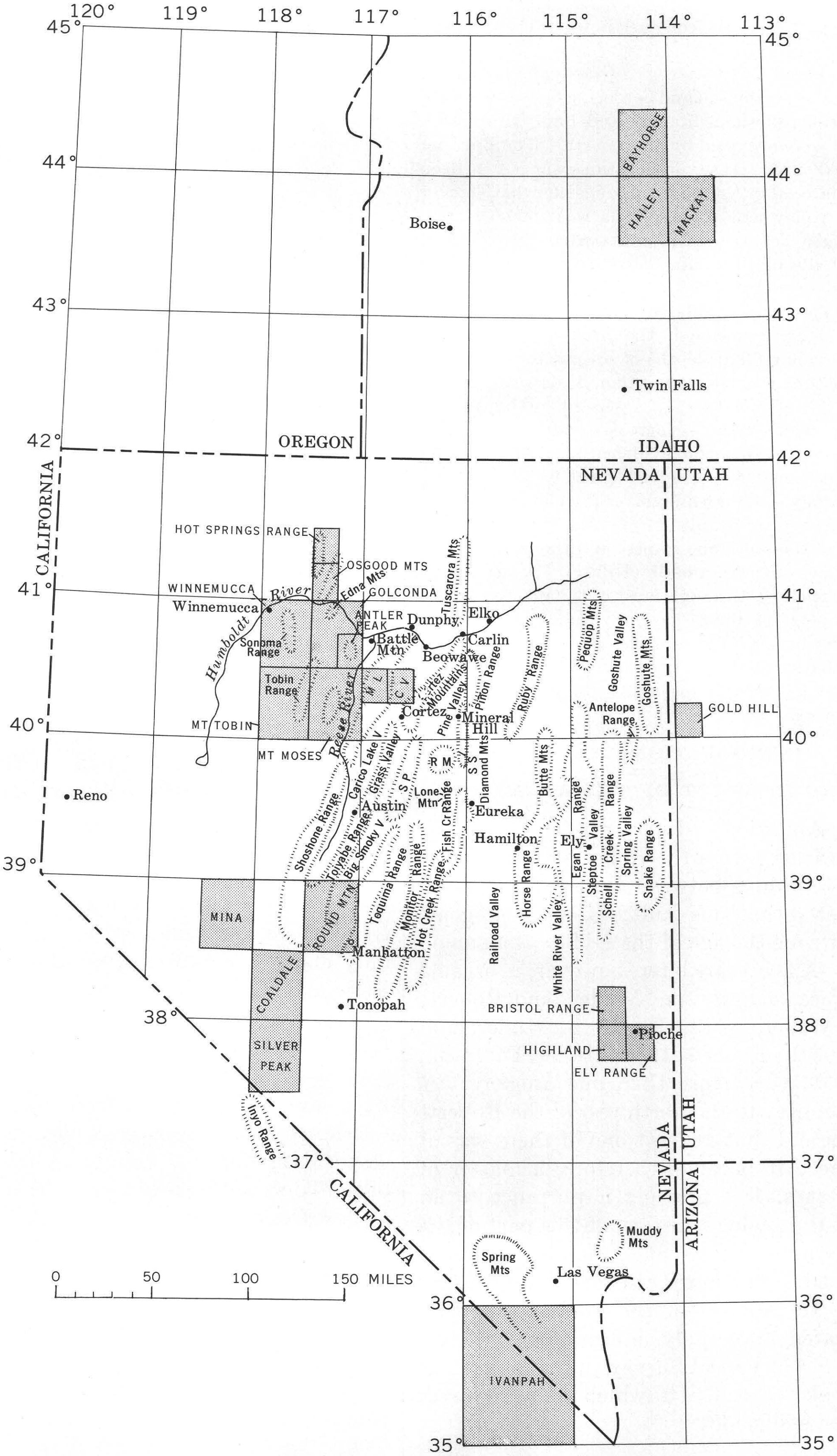

FIGURE 1.—Index map, showing location of Mount Lewis (ML) and Crescent Valley (CV) quadrangles, Nevada, and of other areas referred to in this report. Areas for which the U.S. Geological Survey has published geologic maps are shown shaded. RM, Roberts Creek Mountains; SP, Simpson Park Range; SS, Sulphur Springs Range. Other topographic features referred to in the text are also shown.

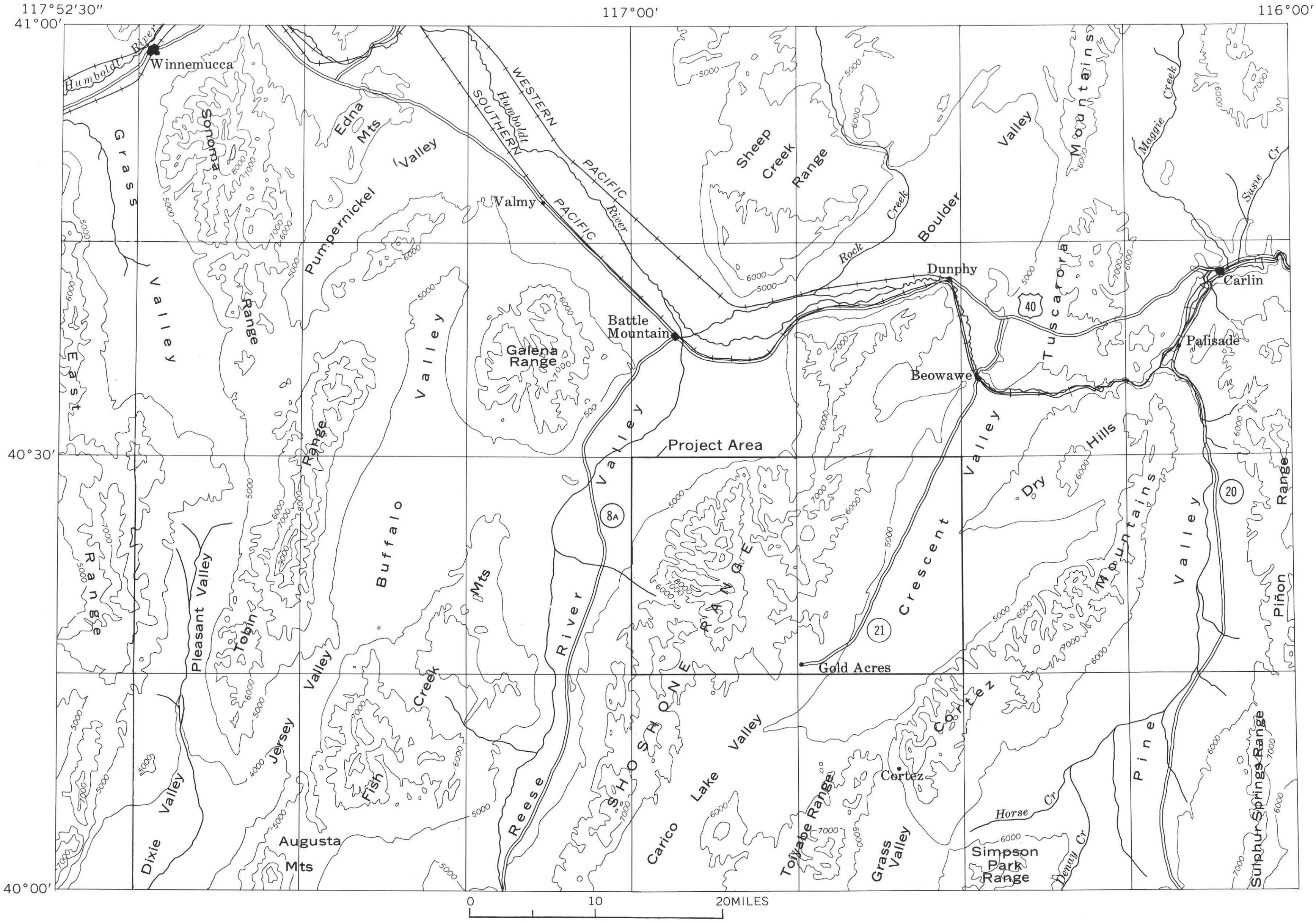

FIGURE 2.—Index map showing relief features and the main transportation routes in and near the northern Shoshone Range.

bly fault scarps. Locally the relief is as much as 3,500 feet in a square mile, but in general the mountains are only moderately steep and somewhat rounded, rising above much flatter alluvial slopes. The southeast slopes of the Cortez Mountains are outside the map area, but are much gentler where they descend to Pine Valley; similarly, the south and east slopes of the Shoshone Range fall off more gently to the alluvial aprons of Crescent and Carico Lake Valleys.

Crescent Valley is asymmetrical, with long, gently sloping alluvial fans on its northwest side extending far out from the Shoshone Range. The fans off the Cortez Mountains are short and much steeper; the playa surface thus lies near—almost at the foot of—the Cortez Mountains. This asymmetry is clearly to be referred to the downfaulting of the valley block along the normal fault bounding the Cortez Mountains (p.127).

Except after storms, few of the canyons carry water· Short segments of Mill, Trout, and Rock Creeks, and Crippen and Lewis Canyons on the northwest slopes of the Shoshone Range are perennial, as are segments of Indian Creek, Mud Spring Gulch, and Fire Canyon on the southeast side. But the only streams whose flow suffices to justify keeping irrigation canals in repair are Mill, Trout, Rock, and Indian Creeks. Each of these supplies water during the spring and early summer for hayfields in the Reese River Valley or Crescent Valley.

CLIMATE AND VEGETATION

As in most of the Basin and Range province, the land is arid. The mean annual precipitation at Battle Mountain and Beowawe (fig. 2) is 6.4 inches. Both these towns are at altitudes below 5,000 feet. At higher altitudes the precipitation is somewhat greater—at Austin, 60 miles to the south and at about 6,800 feet in the Toiyabe Range, the mean annual precipitation is 11.83 inches. Probably the summit areas of the northern Shoshone Range receive between 12 and 15 inches annually. Though the precipitation increases notably with altitude, the increase is probably less here than it is generally in the province, because of the high Tobin and Sonoma Ranges to the west (fig. 2) which effectively strengthen the broad rain shadow of the Sierra Nevada. At any rate, the pinyon and juniper that are widespread at altitudes above 6,000 feet in ranges both to the south and north are sparse here, even hundreds of feet higher, though they do occur, expecially on north-facing slopes. In general, the vegetation is that of the lower part of the Upper Sonoran association, with greasewood, sage, rabbit brush, and sparse grasses dominant. Along streams, box elder and willow are common, and on northern exposures, mountain mahogany, juniper, pinyon, and a few types of shrubs are abundant.

PREVIOUS WORK

Prior geologic work in this area has been almost all reconnaissance. The earliest, by the 40th Parallel Survey (Hague, 1877), is of only historic interest. A glance at the map (pl. 1) will show why the generally excellent reconnaissance of Arnold Hague and S. F. Emmons was not here so fruitful as in most other areas. These men recognized the intrusive bodies of Mount Lewis and O'Haras Peak, the Tertiary volcanics east of the crest of Shoshone Range, and the fact that most of the range is composed of Paleozoic rocks; their assignment of these rocks to a "Lower Coal Measures" age was, however, without support. W. H. Emmons (1910, p. 113–126) visited the mining areas of Tenabo, Lander, Mud Spring, Hilltop, and Lewis Canyon, and described the workings active in 1908. His discussion of the general geology largely followed the early interpretations of Hague. Schrader (1934) briefly visited Horse Canyon. Waring's description of the water resources of the Reese River Valley (1919) made incidental mention of part of the area. Brief descriptions of some of the mines, in connection with discussions of mineral production, have been made by Raymond (1875), Whitehill (1877, p. 76; 1879, p. 71), Stuart (1909, p. 127), Martin (1910), Carpenter (1910), Hill (1912, p. 215–216), Lincoln (1923), Smith and Vanderburg (1932, p. 59–60), Schrader (1934), Vanderburg (1939), and Gianella (1941, p. 295–299), and in the annual volumes of "Mineral Resources of the United States."

FIELDWORK AND AUTHORSHIP

The map (pl. 1) represents about 58 man-months of fieldwork by 12 geologists, beginning in July 1950 and closing in September 1959. Geologists whose fieldwork is represented, with the approximate time they devoted to it, include James Gilluly (22 mo.), Olcott Gates (12 mo.), M. R. Mudge (5½ mo.), and Frank Barnett, A. J. Boucot, Harald Drewes, H. R. Gould, K. B. Ketner, H. E. Malde, R. J. Ross, Jr., E. M. Shoemaker, and G. L. Snyder, all of whom worked from 1½ to 4 months. Ketner has also examined all the mines that were being actively worked in 1955, as well as the accessible parts of the inactive mines. He is responsible for the "Economic geology" section of the report. Olcott Gates is principally responsible for the sections dealing with the igneous rocks, James Gilluly for the other sections, though both have collaborated throughout.

ACKNOWLEDGMENTS

The cooperation of many colleagues in the Geological Survey greatly facilitated the work here reported. C. W. Merriam, H. G. Ferguson, R. J. Roberts, and

T. B. Nolan, all long familiar with problems of Nevada stratigraphy, gave important help during early parts of the study. Jean Berdan, Helen Duncan, A. R. Palmer, Mackenzie Gordon, Jr., R. J. Ross, Jr., and P. E. Cloud, Jr. all visited the area and aided in collecting fossils and in correlations. The keen observations of M. R. Mudge were responsible for a disproportionately large number of fossil collections; they furnish an important element in our stratigraphic foundation. F. B. Van Houten, of Princeton University, kindly added to our collection of Tertiary vertebrates. The courtesy of George Shwin and Dan Fillippini, both ranchers in the area, is also gratefully acknowledged. Aid in preparing the report for publication was given by Harold Masursky, G. C. Cone, and Roberta Dixon. The report has greatly benefited from the critical reviews of C. B. Hunt and J. Fred Smith, Jr.

GENERAL GEOLOGY

PRINCIPAL FEATURES

The rocks of the mountains include Paleozoic sedimentary and volcanic rocks, Mesozoic(?) sedimentary rocks, and Tertiary igneous and sedimentary rocks. The intermontane valleys are occupied by Quaternary alluvium and playa deposits.

The rocks of early and middle Paleozoic age include strata of two distinct sequences. One sequence, which embraces rocks of Cambrian to Early Mississippian age, consists largely of carbonate rocks and broadly resembles the classic stratigraphic section at Eureka, Nev. (pl. 3; Hague, 1883, 1892; Merriam and Anderson, 1942, p. 1680–1693; Nolan, Merriam, and Williams, 1956). The second sequence, though of roughly equivalent age—Late Cambrian to Middle Devonian—contains almost no limestone and consists almost wholly of chert, siliceous clastics, and volcanic rocks. The two sequences are everywhere in fault contact. They are respectively referred to herein as the "carbonate" or "eastern" facies and the "siliceous" or "western" facies of the rocks of early and middle Paleozoic age.

The carbonate facies includes minor quartzite, shale, some greenstone (suggesting transition to the clastic facies), and predominant dolomite and limestone of Early to Middle Cambrian age, about 3,000 feet thick; quartzite, dolomite, and limestone of Ordovician age aggregating about 1,000 feet; shaly limestone, limestone, and dolomite of Silurian age, perhaps 1,200 feet; Devonian limestone about 1,000 feet; and Devonian and Mississippian shale about 500 feet. None of these thicknesses is accurate, as almost every section is cut by faults.

The siliceous facies lacks definite lithologic marker beds and is poorer in fossils than the carbonate facies. Consequently, only rough estimates of thickness can be made. The Cambrian includes only a few score feet of arkosic grit. The Ordovician includes at least 12,000 feet—perhaps 30,000—of quartzite, chert, arkose, graywacke, pillow lava and tuff, and minor limestone; the Silurian, 2,000 to perhaps 4,000 feet of feldspathic sandstone; and the Devonian at least 2,000 feet—more likely 4,000—of chert, and subordinate shale, sandstone, and limestone.

The emphatic contrast in lithology between these two sequences compels the conclusion that they represent quite different depositional environments, and indeed that a gap of at least a few miles must have intervened between the area of deposition of any one bed of either sequence and that of its correlative in the other, as now seen. As discussed in detail in the pertinent sections on stratigraphy, the very slight tendency toward convergence of facies in the Cambrian and Silurian rocks of the two sequences is far too insignificant to vitiate this conclusion. The present juxtaposition of the two facies is demonstrably due to a major thrust—undoubtedly a western extension of the thrust first recognized by Merriam and Anderson (1942) in the Roberts Mountains, 50 miles to the southeast, and here referred to as the Roberts thrust.

Two distinct facies of upper Paleozoic rocks are also represented in the Shoshone Range. One, here referred to as the Antler Peak facies, consists of a conglomerate of Middle Pennsylvanian age a few hundred feet thick, and limestones of Early Permian age about 800 feet thick. Although allochthonous here, this facies is virtually identical with the sequence at Antler Peak in the Galena Range, across the Reese River Valley (fig. 2), which is there autochthonous. Some thrust slivers in the Mount Lewis area show the conglomerate in unambiguously depositional contact on the siliceous facies of the Ordovician, a relation precisely like that at Antler Peak. The lower Paleozoic rocks had therefore been deformed and deeply eroded prior to Middle Pennsylvanian time, as demonstrated by Roberts (1951).

A second facies of upper Paleozoic rocks is represented at Antler Peak by two clastic formations—the Pumpernickel and Havallah—both of which are there allochthonous but are autochthonous in the Sonoma and Tobin Ranges farther west (Ferguson, Muller, and Roberts, 1951a; Ferguson, Roberts, and Muller, 1952; Roberts, Hotz, Gilluly, and Ferguson, 1958, p. 2824). This facies of fine-grained, generally siliceous rocks, here called the Sonoma Range facies, is represented near Mount Lewis in this area by a few small thrust blocks of argillite a few hundred feet across. Associated with it is a thrust slice of Cambrian arkose, also of more westerly provenance. These were carried into the range during the Lewis orogeny.

A third sequence of upper Paleozoic rocks, here called the Cortez Mountain sequence, somewhat resembles the Antler Peak facies. It consists of a basal dolomite about 200 feet thick, overlain by conglomerate and sandstone about 400 feet thick, and these, in turn, by interbedded limestone and sandstone about 200 feet thick. A considerable additional section many hundred feet thick overlies these rocks farther east, outside the map area. These rocks have yielded fossils of Late Pennsylvanian age. This sequence is confined to the Cortez Mountains; it rests on a basal fault of uncertain significance but is probably almost in its site of deposition. It belongs to the same general facies as the Antler Peak sequence but contrasts strongly with the Sonoma Range facies.

Tertiary rocks include predominantly volcanic rocks, some plutonic bodies, and some stream gravels and conglomerate. They seem clearly to belong to three distinct episodes, separated by erosional intervals.

The earliest episode is represented by a granodiorite stock, whose age is unknown but perhaps Eocene. This mass was deroofed by erosion, and a relief of several hundred feet was carved across it and the surrounding country rocks before the whole area was buried beneath quartz latitic pyroclastic rocks and lava. Three breccia pipes near the crest of the Shoshone Range and two domes on the east side of Mount Lewis were probably active vents of this episode. Most of the eruptive rocks are quartz latite, with some rhyolite and dacite; the associated intrusives include granodiorite and quartz monzonite. Interbedded gravels contain fossils of an age near the Miocene and Pliocene transition.

Unconformably on the Miocene and Pliocene volcanics in the eastern flank of the Shoshone Range are cuesta-forming flows of basaltic andesite, presumably of late Pliocene or early Pleistocene age. Similar flows form the dip slope of the Cortez Mountains east of Crescent Valley. The numerous dolerite dikes cutting the western slopes of the Cortez Mountains may represent some of the fissures through which these lava flows were fed. These flows are southern extensions of the great lava plateau of northern Nevada and southern Idaho and are probably of Pliocene age. In the Cortez quadrangle to the southeast, they are associated with quartz latite tuffs that resemble the fossiliferous beds of the Mount Lewis area. The youngest deposits are the widespread alluvial and fan gravels, and subordinate playa sediments that occupy most of the Reese River and Crescent Valleys. These sediments, of probably Pleistocene and Recent age, were not studied in detail.

The structure of the area is dominated by thrusting. During the early thrusting, the Antler orogeny, the rocks were strongly contorted and the thrust surfaces locally folded, even overturned. Progressively less deformation of successively higher branch thrusts shows that at least some of this folding of the thrusts was concurrent with their active horizontal displacement and did not result from later folding of a quiescent thrust blanket.

The youngest observed strata of the lower plate of the Roberts thrust are of Early Mississippian age. But the upper plate is composite and composed of a thick pile of discrete thrust sheets. The structurally highest thrust plates involve rocks of Pennsylvanian, Permian, and Triassic(?) age, are apparently less complexly deformed than most of those beneath, and probably record an epoch of thrusting, the Lewis orogeny, long subsequent to the emplacement of the Roberts thrust. Two epochs of thrusting, one pre-Pennsylvanian and one Mesozoic, have been demonstrated in the adjoining Antler Peak quadrangle (Roberts, 1951). The latest thrust movement recognized in the Shoshone Range involved the Lower Triassic(?) strata and is perhaps the same as the post-Permian episode at Antler Peak. It was clearly preintrusive and prevolcanic, as dikes cut the thrust faults without offset and volcanic rocks lie on thrust sheets of widely different levels.

Younger than all the lavas is the large-scale normal faulting responsible for the major part of the present relief. The displacement on these Basin Range faults is measured in thousands of feet, as suggested both by the physiography and the results of the gravity measurements by Donald Plouff and Samuel Stewart, of the U.S. Geological Survey, in Crescent Valley and the Cortez Mountains.

Precious metal lodes, both of silver and gold, have been worked intermittently at least since 1869, if not for a few earlier years, and aggregate production has been more than $11 million. The silver deposits were largely antimonial silver ores in the Lewis, Kimberly (Hilltop), and Bullion districts. None were being continuously mined at the time of this study. Minor quantities of copper and lead have also been recovered from the silver deposits. Gold deposits at Gold Acres, south of Tenabo, were not discovered until 1935 but during the time of this survey were the major sources of ore in the area. Barite deposits were being intermittently worked at several localities: in the foothills west of Slaven Canyon, in Bateman Canyon, and on the Elder Creek-Cooks Creek divide. All these are bedded replacement bodies in chert and limestone of the Slaven Chert.

ROCKS OF THE AREA

GENERAL FEATURES

Table 1 summarizes the stratigraphy of the pre-

Tertiary rocks of the area. Details are presented in immediately succeeding pages. The locations of the fossil collections upon which the various age assignments are based are presented on plate 4.

TABLE 1.—*Pre-Tertiary rocks of the Mount Lewis and Crescent Valley quadrangles, Nevada*

Rocks cut by thrust faults (probably Mesozoic) not demonstrably related to the Roberts thrust

		Sonoma Range sequence	Antler Peak sequence	Cortez Mountains sequence		
Triassic(?)	Middle(?) and Lower	China Mountain(?) Formation 1,200± ft			Middle(?) and Lower	Triassic
Permian	Upper				Upper	Permian
	Lower	Havallah Formation[1] 500± ft	Antler Peak Limestone[2] 700± ft	*Brock Canyon Formation[4] 700± ft	Lower	
Pennsylvanian	Upper				Upper	Pennsylvanian
	Middle	Fault contact; depositional relations unknown	Battle Conglomerate[3] 700± ft		Middle	
	Lower			Depositional relations uncertain-faulted	Lower	

Rocks demonstrably cut by the Roberts thrust (probably upper Paleozoic)

			Siliceous (western) facies	Carbonate (eastern) facies		
Mississippian	Middle				Middle	Mississippian
	Lower			Major thrust fault	Lower	
Devonian	Upper		Fault contact	Pilot Shale 300(?) ft	Upper	Devonian
	Middle		*Slaven Chert[5] Fault contact (depositional relations unknown)	Limestone[6] (undifferentiated) 1,000(?) ft	Middle	
	Lower		Fault contact	Fault contact	Lower	
Silurian			*Elder Sandstone 2,000± ft			Silurian
			Fault contact (depositional relations unknown)	Roberts Mountain Limestone[7] 1,000(?) ft		
Ordovician	Upper		Valmy Formation 20,000± ft	Hanson Creek Formation 600(?) ft	Upper	Ordovician
	Middle			Eureka Quartzite[8] 200 ft Fault and unconformity	Middle	
	Lower		Fault contact		Lower	
Cambrian	Upper	Harmony Formation 200± ft			Upper	Cambrian
	Middle			*Shwin Formation[9] 1,500± ft Fault contact Eldorado Dolomite 500 ft Fault contact	Middle	
	Lower			Pioche(?) Shale and Prospect Mountain Quartzite[10] 200± ft	Lower	

[1] Present only as small thrust slivers of a few hundred feet of strata; thickness in Antler Peak quadrangle about 4,600 ft and in the Sonoma Range about 10,000 ft.

[2] Apparently conformable on Battle Conglomerate, but with no faunal suggestion of any strata of Middle Pennsylvanian age equivalent to the Highway Limestone of the Edna Mountains area (Ferguson, Roberts, and Muller, 1952). This possible disconformity is evidently present at the type locality in the Antler Peak area also.

[3] Most bodies are thrust slivers with mechanical contacts only, but two rest unconformably on bodies of Valmy Formation; the large composite blocks so composed are themselves thrust slices.

[4] Base of formation between Brock and Mule Canyons, Cortez Range, is a fault, and a second fault separates the basal dolomite from the overlying conglomerate. Probably both faults are small and the formation is parautochthonous.

[5] Both base and top of the Slaven Chert are mechanical; there are no clues as to whether the formation embraces more than the Middle Devonian to which all fossils thus far found in it have been referred.

[6] The only fossils thus far found in this unit in this area, north of Gold Acres, are of Late Devonian age, suggesting relation to the type Devils Gate (Merriam, 1940). The original definition was based on faunal zones not here recognizable because of metamorphism, and as the lithology of the Devils Gate Limestone is essentially indistinguishable from that of the Nevada Limestone, where neither is dolomitic, the possibility that Nevada Limestone may be present cannot be excluded.

[7] The Lone Mountain Dolomite, of Middle and Late Silurian age, is not exposed anywhere northwest of the Roberts Mountains; its absence here may be due either to faulting, nondeposition, or facies change into Roberts Mountain lithology.

[8] The only exposed body of Eureka Quartzite in this area (Goat Ridge, head of Hancock Canyon, Mount Lewis quadrangle) has fault contacts. At Cortez, however, the formation rests unconformably on the Hamburg Dolomite of Late Cambrian age.

[9] The Shwin Formation strongly resembles the Secret Canyon and Geddes Limestones of Middle Cambrian age, as these are developed near Eureka, except for intimate association (interpreted as interbedding) with metavolcanic rocks.

[10] Thinly laminated schistose beds a few feet thick may represent the faulted-out remnant of some Pioche Shale; the base of the quartzite concealed.

EXPLANATION

* New formation defined in this report

No evidence of strata of age indicated: hiatus may be nondepositional, erosional, or tectonic

Known erosional unconformity either locally or within radius of 30 miles

STRATA OF THE LOWER PLATE OF THE ROBERTS THRUST (CARBONATE OR EASTERN FACIES)

CAMBRIAN SYSTEM

LOWER CAMBRIAN SERIES

PROSPECT MOUNTAIN QUARTZITE

The only rock body here referred (with some doubt) to the Prospect Mountain Quartzite is a block about 700 feet long and 250 feet wide that forms a prominent cliff on the southwest spur of the ridge between Hancock Canyon and Mill Creek, about 2¾ miles west-northwest of Goat Peak.

All contacts of the block are faults. Obscure cross-bedding suggests that the nearly vertical beds, which strike west, face northward. About 300 feet of highly sheared strata cemented by veins of white quartz is exposed. Surfaces of unsheared masses weather brown, but so much of the rock is sheared that the general aspect of the cliffs is almost white. The rock is composed of fairly well rounded quartz grains, some feldspar, and locally a few flakes of muscovite. Although there are some granules as much as 3 or 4 mm across, most of the grains are considerably finer—perhaps averaging half a millimeter or less. Quartz predominates but some beds contain considerable feldspar and some euhedral pyrite, or limonitic pseudomorphs after it.

No fossils have been found in the Prospect Mountain Quartzite either here or elsewhere. At Eureka, the type locality, 70 miles to the southeast (fig. 1), the quartzite passes by transition into the overlying Pioche Shale, from which fossils diagnostic of Early Cambrian age have been obtained (Nolan, Merriam, and Williams, 1956, p. 7). The reference of this local body to the Prospect Mountain Quartzite is based upon its resemblance to the Prospect Mountain of the type locality both in general lithology and in the association of similar shaly beds that suggest transition perhaps to the Pioche Shale.

The quartzite also resembles the Osgood Mountain Quartzite of the Osgood Mountains (fig. 1) and neighboring parts of the Golconda quadrangle, about 50 miles to the northwest (Ferguson, Roberts, and Muller, 1952). We prefer to refer the rocks in the Shoshone Range to the Prospect Mountain, however, because the associated strata here resemble the lower Paleozoic rocks of Eureka much more closely than they do the Preble and Comus Formations, which are the associates of the Osgood Mountain Quartzite.

Slivers, a few feet thick, of sheared and silicified shale, sandstone, and sandy shale intervene between the more massive quartzite and the Roberts Mountains Limestone to the north. These slivers closely resemble the Pioche Shale, though no fossils have been found in them. The whole assemblage of rocks is reminiscent of the Lower Cambrian strata at Eureka (T. B. Nolan and C. W. Merriam, oral commun., 1953).

The only other quartzite of the eastern facies with which these rocks might be confused is the Eureka Quartzite. The Eureka, however, lacks comparable shaly and feldspathic interbeds, and furthermore the upper part of the Eureka generally weathers white, in contrast to the rusty brown of the unsheared quartzite here referred to the Prospect Mountain.

LOWER AND MIDDLE CAMBRIAN SERIES

PIOCHE SHALE

As just noted, a few beds suggesting transition between the Prospect Mountain Quartzite and the Pioche Shale intervene between the massive quartzite and the limestone to the north. No fossils have been found to confirm an assignment of these rocks to the Lower or Middle Cambrian. The beds do, however, resemble the Pioche Shale. The outcrops are much too small to be shown on the scale of plate 1. The Pioche Shale, therefore, is not included in that map's explanation; it is mentioned here merely to put on record the occurrence of rocks resembling parts of the Pioche, from which a more discerning or fortunate collector might derive fossil evidence for assignment to the Pioche.

MIDDLE CAMBRIAN SERIES

ELDORADO DOLOMITE

Distribution and topographic expression

The rocks here referred to the Eldorado Dolomite crop out in the steep cliffs and slopes on the mountain front on both sides of Hancock Canyon in the NE¼ sec. 19, T. 29 N., R. 45 E. The area of outcrop is perhaps 40 acres, much of it in steep cliffs.

Lithology and thickness

The formation is composed almost wholly of highly broken and irregularly bedded dolomite. The vestigial bedding surfaces are spaced from 1 to 8 feet apart. The dolomite is coarsely granular and largely recrystallized, as is shown by coarsening alongside many of the fissures. Locally, especially toward the north, a dedolomitized matrix holds sprays and fibers of tremolite. The average grain size of the dolomite is probably 1 mm. Most of the discernible bedding lies nearly flat, suggesting a simple structure, but all the contacts, except possibly one, are obviously faults and the innumerable fissures and breccia zones suggest that the true structure may be considerably more complex than is apparent. If the apparent structure may be trusted, however, the thickness is roughly 500 feet, with no base exposed. This thickness is well below the 2,000-foot thickness of the Eldorado Dolomite in the Eureka district (Wheeler and Lemmon, 1939, p. 19; Nolan, Merriam, and Williams, 1956, p. 11). Owing to the structural complexities and especially to concealment of the base, this difference is not significant.

Age and correlation

This dolomite resembles both the Eldorado Dolomite and the Hamburg Dolomite of the Eureka district. Several lines of evidence seem to favor its correlation with the Eldorado, though none is conclusive.

On the north spur of Hancock Canyon at the mountain face, the dolomite seems to be conformably overlain by mottled limestone of the Shwin Formation, which resembles some in the Geddes Limestone and Secret Canyon Shale at Eureka. Agnostid trilobites from the limestone of the Shwin Formation at Hancock Canyon are like those of the Secret Canyon, of Middle Cambrian age (A. R. Palmer, oral commun., 1953; this report, p. 14). Thus, if the contact is truly conformable, the underlying dolomite should be Eldorado rather than Hamburg, for the Hamburg overlies the Secret Canyon Shale at Eureka. There is a real question, however, as to whether the contact is conformable. It is much sheared and no volcanics are associated with the limestone here, whereas elsewhere it seems that volcanics are especially abundant at lower stratigraphic levels in the Shwin Formation. Perhaps, then, the contact between dolomite and the overlying limestone is really a fault, despite the apparent conformity of bedding on either side.

There is perhaps more persuasive indirect evidence for correlating the dolomite at Hancock Canyon with the Eldorado rather than the Hamburg. At Eureka, the Eureka Quartzite rests on rocks of the Pogonip Group, but as it is followed toward the northwest the Eureka rests on older and older beds. At Pete Hanson Creek in the Roberts Creek Mountains, it rests on beds low in the Pogonip (Kirk, 1933, p. 31), and at Western Peak on the north face of these mountains it rests on still older beds, almost at the base of the Pogonip, as noted by R. J. Ross, Jr., Harold Masursky, and James Gilluly, who visited the locality together in 1958. Still farther northwest, at Cortez, 15 miles southeast of Hancock Canyon, the Eureka Quartzite rests on a dolomite considered by Merriam (Merriam and Anderson, 1942, p. 1684) to be older than the Pogonip.

The only thick dolomites older than Pogonip in the general region are the Hamburg and Eldorado, neither of which is fossiliferous so far as known. If the rock at Cortez is Hamburg, many hundred feet of Pogonip beds has been removed by erosion between Eureka and Cortez. It would not be surprising to find the unconformity to have cut still lower stratigraphically in the distance between Cortez and Hancock Canyon. No other thick dolomite, older than the Eureka, is known in the area of the Goat window in the Roberts thrust. It thus seems best to correlate this dolomite with the Eldorado.

SHWIN FORMATION

Distribution and topographic expression

The rocks here designated as the Shwin Formation are confined to the area of the Goat window in the Roberts thrust (fig. 29). The principal exposures are (1) on and east of Shwin Ranch, from which the formation is named, on the south slope of the ridge between Goat Canyon and the North Fork of Mill Creek; (2) on the south wall of Trout Creek Canyon just above its mouth; (3) on the spur north of Hancock Canyon, just west of the steep mountain slope.

Stratigraphy

The principal constituents of the formation are chloritic argillite, metadolerite, greenstone, chloritic phyllite, black limy slate, mottled shaly limestone, limy mudstone, and calc-phyllite. The stratigraphic succession is uncertain because of complex folding and faulting. The chloritic phyllite and green argillite appear to be interlayered with the limy phyllite through a thickness of at least 300 feet, both on the spur south of Goat Peak and on the one between Hancock and Trout Creeks. Locally, the formation is complexly—even isoclinally—folded, the beds are commonly sheared out, and the interlayering of metavolcanics and limy mudstone may indeed be wholly mechanical rather than depositional. Nevertheless, the association is intimate over wide-enough areas to make it seem probably depositional, and in any event the several lithologic components are not separately mappable even on a scale far larger than that of the geologic map (pl. 1).

All the contacts of the formation with others are faults (with the possible exception of one north of Hancock Canyon already mentioned in connection with the Eldorado Dolomite, and even this is sheared); neither base nor top is identifiable as such. For what it is worth, our impression is that the rocks derived from the more limy and muddy sediments—the limestone, mudstone, calc-phyllite, slate, and local argillite—are generally higher stratigraphically than are the greenstone, metadiabase, and chlorite phyllites.

The metadolerite and chlorite phyllites, interpreted as making up much of the lower part of the formation, are best exposed in Trout Creek Canyon, for half a mile above its mouth. The more massive metadolerite is difficultly distinguished from some similar greenstones of the Valmy Formation. The chloritic phyllite is at least partly demonstrable (through the presence of all transitional stages) as derived from the shearing out and recrystallization of the massive dolerite. It is distinguished by a sheen on the somewhat crinkly cleavage surfaces that has not been observed in other formations of the area. A similar phyllitic aspect has also been noted in the associated limestone and mudstone, which develop facies of calc-phyllite and slate. The metamorphism does not appear to be directly attributable to deformation along the major Roberts thrust, however, for it is not observed in greenstones of the Valmy Formation in contact with the trust, nor is the phyllite in the Shwin Formation restricted to the neighborhood of the major fault.

The limy parts of the formation greatly resemble the beds of the Secret Canyon Shale as exposed in the Eureka district (Wheeler and Lemmon, 1939, p. 20; Nolan, Merriam, and Williams, 1956, p. 13–15). They include the gray limy mudstone and mottled shaly limestone that weathers to tan or light-brown hues, shale-pebble conglomerate layers, minor edgewise conglomerate and, locally, thin-bedded limy shale. A few 10-foot beds of more massive gray limestone, reminiscent of the Geddes Limestone at Eureka (Wheeler and Lemmon, 1939, p. 20–22), are interfolded with chloritic phyllite on the south slope of Goat Peak.

Thickness

The intense folding, smearing out of beds, and abundant faults prevent measurement of stratigraphic thickness. Only an estimate can be made: perhaps 600 to 1,000 feet of dominantly metavolcanic rock and 500 to 1,000 feet of dominantly limy and fine clastic. As a guess, 1,500 to 2,000 feet for the total exposed thickness may not be far from correct.

Petrography

The least sheared doleritic rocks preserve relict ophitic texture, together with chloritic pseudomorphs after olivine. These rocks were obviously olivine dolerites originally, though they are now composed almost wholly of albite, chlorite, sericite, calcite, local epidote, and accessory apatite, leucoxene, and magnetite. Some specimens from near the mouth of Trout Creek Canyon contain porphyroblasts of biotite and brown hornblende, locally biotite and chlorite, presumably contact effects of a subjacent intrusive mass. Specimen 405, table 2, is such a rock, notably rich in ilmenite. The ilmenite appears to be of late introduction.

Most of the metavolcanic rocks, however, show no igneous textures; they have been so intimately sheared out that their present textures are wholly lepidoblastic. They consist chiefly of chlorite and muscovite, with subordinate albite, calcite, epidote, and locally quartz, which perhaps represent added clastic material or intermixed tuffaceous and detrital sediments. Rutile and ilmenite are abundant. Some of the volcanic rocks, such as specimen 408 of table 2, are interlaminated, even on the scale of a thin section, with calcite.

TABLE 2.—*Chemical analyses of Paleozoic sedimentary and metavolcanic rocks, Mount Lewis and Crescent Valley quadrangles, Nevada*

[Analysts: L. N. Tarrant, specimens 2–5, 12; V. C. Smith, other 10 specimens]

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Field No	272	403	404	408	405	1	4	245	19	225	15	488	608	264	393
SiO_2	86.26	66.73	61.44	33.08	51.74	98.42	97.45	98.17	87.32	91.70	98.96	40.09	86.98	86.78	72.97
Al_2O_3	6.52	13.91	17.56	11.74	16.83	.24	.43	.37	.59	4.14	.23	13.18	6.30	6.16	11.41
Fe_2O_3	.31	.99	.57	1.43	1.63	.07	.27	.39	.15	.20	.16	.81	.89	.51	.84
FeO	1.17	3.93	4.94	7.01	9.09	.11	.13	.09	.09	.18	.18	10.64	.01	.29	2.98
MgO	.30	5.09	4.99	4.72	4.84	.00	.00	.10	2.39	.17	.00	13.93	.37	.50	4.14
CaO	.62	.97	.33	18.73	1.72	.00	.00	.00	3.52	.12	.00	6.10	.15	.18	.00
Na_2O	2.02	.22	3.19	.31	5.11	.00	.00	.00	.03	.04	.01	.00	.05	.94	.46
K_2O	1.11	2.68	2.02	1.79	1.27	.00	.07	.02	.13	.48	.02	.00	3.07	2.72	3.14
H_2O	.09	.24	.13	.12	.10	.01	.03	.06	.04	.26	.01	.18	.12	.28	.21
H_2O+	.67	3.91	3.75	3.96	2.97	.09	.11	.15	.16	1.43	.09	7.49	.78	.68	2.95
TiO_2	.23	.57	.67	2.24	3.43	.03	.03	.02	.05	.21	.01	2.70	.47	.26	.56
CO_2	.49	.23	.00	14.33	.11	.01	.02	.02	5.30	.02	.02	4.41	.09	.02	.03
P_2O_5	.03.	.14	.06	.46	.66	.02	.06	.02	.01	.04	.01	.35	.06	.05	.00
SO_3						[1].15	[1].28								
S		.00			.02	[1].05	[1].05			.06					
MnO	.02	.05	.04	.16	.14	.00	.01	.02	.04	.01	.00	.11	.06	.01	.02
BaO		.05			.06	.28	.51			.07			.07		
						99.48	99.45			99.13					
Less O					.01	.03	.03			.03					
Total	99.84	99.71	99.69	100.08	99.71	99.45	99.42	99.43	99.82	[2] 99.10	99.70	99.99	99.47	[2] 99.38	99.71

[1] Calculated from total S. The amount of S necessary for barite was calculated to SO_3; the remainder of the sulfur to sulfide.

[2] High carbon and organic matter.

DESCRIPTION AND LOCATION OF SPECIMENS

1. Arkose, Harmony Formation, N. ridge of Mount Lewis, sec. 1, T. 29 N., R. 45 E.
2. Argillite, Shwin Formation, S. side Trout Creek, sec. 18, T. 29 N., R. 45 E.
3. Slate, Shwin Formation, S. side Trout Creek, sec. 18, T. 29 N., R. 45 E.
4. Greenstone, Shwin Formation, N. side Trout Creek, sec. 17, T. 29 N., R. 45 E.
5. Albite dolerite, Shwin Formation, S. side Trout Crock, sec. 18, T 29. N., R. 45 E.
6. Quartzite, Valmy Formation, N. of Nelson mine, sec. 28, T. 30 N., R. 46 E.
7. Quartzite, Valmy Formation, SE. of Nelson mine, sec. 33, T. 30 N., R. 46 E.
8. Quartzite, Valmy Formation, on front E. of mouth of Crum Canyon, sec. 4, T. 30 N., R. 46 E.
9. Dolomitic sandstone, Valmy Formation, E. side Crum Canyon, sec. 21, T. 30 N., R. 46 E.
10. Chert arenite, Valmy Formation, Maysville Summit, sec. 4, T. 29 N., R. 46 E.
11. Chert, Valmy Formation, W. side Bateman Canyon, sec. 21, T. 30 N., R. 46 E.
12. Greenstone, Valmy Formation, S. side North Fork of Mill Creek, sec. 35, T. 29 N., R. 45 E.
13. Feldspathic siltstone, Elder sandstone, ½ mile SE. of Utah Mine camp, T. 28 N., R. 46 E.
14. Feldspathic siltstone, Slaven Chert, E. side Slaven Canyon, sec. 24, T. 30 N., R. 46 E.
15. Shale, Slaven chert, S. side Crippen Canyon, sec. 8, T. 29 N., R. 45 E.

The rocks of the formation that are not obviously of volcanic derivation are chiefly shale, laminated sericitic siltstone, coarse quartz siltstone, phyllitic slate, argillite, and limestone. They are composed of muscovite, chlorite, quartz, subordinate plagioclase (probably albite), and calcite that ranges in abundance from a trace to overwhelming dominance. Organic matter and authigenic pyrite are abundant. Some specimens contain porphyroblasts of andalusite, doubtless products of thermal metamorphism near the quartz diorite masses on the ridge west of Goat Peak.

A specimen of quartz siltstone with incipient foliation is illustrated in the photomicrograph of figure 3, and an outcrop of crumpled phyllite is shown in figure 4.

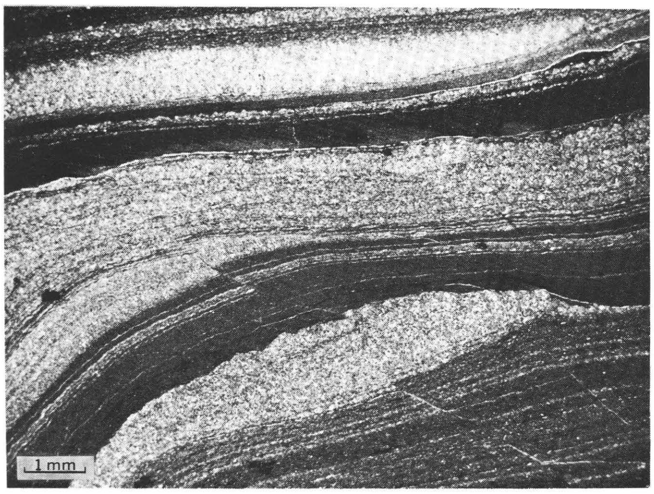

FIGURE 3.—Photomicrograph of quartz-sericite siltstone of the Shwin Formation. Composed of quartz silt (45 percent), ranging from very fine (0.005 mm) in some layers to coarse (0.035 mm) in others, sericite (50 percent), in flakes as much as 0.03 mm long, pyrite (chiefly altered to iron oxides), and organic matter (5 percent). Minor sphene, calcite, and apatite are present. Incipient foliation, defined by bent, rotated, and newly grown sericite, crosses the bedding from upper left to lower right in the photograph. Where the foliation crosses the dark laminae, it simulates foreset bedding. There has been considerable flowage of the coarse silt (light bands) into the fold apexes as shown in the lower and upper center parts of the photograph. Crossed nicols.

Chemical composition

Table 2 contains analyses of four specimens of the Shwin Formation, selected to show the most diverse lithologic varieties that are abundant in the area, except for some nearly pure limestones. Table 3 contains semiquantitative spectrographic analyses for many of the rocks of the quadrangles.

Age and correlation

Intensive deformation and stretching of some of the fossils have rendered them unidentifiable. Half a dozen localities, however, have yielded at least some organic remains. The collections were submitted to A. R. Palmer, of the U.S. Geological Survey, who reported as follows:

F–49 (USGS colln. 1966–CO)—600 ft N., 1,000 ft W. of SE cor. sec. 18, T. 29 N., R. 45 E. Alt 6,800 ft, on nose of ridge S. of Trout Creek.
Peronopsis? sp.
Ptychagnostus? sp.

F–50 (USGS colln. 1967–CO)—1,700 ft N., 2,200 ft W. of SE cor., sec. 17, T. 29 N., R. 45 E. Alt 6,480 ft on nose N. of Trout Creek.
Peronopsis sp.

F–51 (USGS colln. 1968–CO)—2,000 ft W., 4,700 ft N. of SE cor., sec. 19, T. 29 N., R. 45 E.; 600 ft NE. of road crossing at mouth of Hancock Canyon. Alt 6,280 ft.
Unidentifiable trilobites, one suggestive of an eodiscid.

F–53 (USGS coll. 1969–CO)—3,600 ft W., 4,400 ft N. of SE cor., sec. 20, T. 29 N., R. 45 E. Alt 7,520 ft on S. side of spur.
Hypagnostus sp.
Ptychagnostus sp.
Unidentifiable trilobite fragments

F–56 (USGS colln. 1970–CO)—2,250 ft W., 1,500 ft N. of SE cor., sec. 28, T. 29 N., R. 45 E. Alt 7,960 ft on spur SW. of Goat Peak.
Inarticulate brachiopods; some of *Pegmatreta* type.

All the collections appear to be Middle Cambrian in age. The agnostid trilobites cited above, although badly distorted, are similar to those found in the Secret Canyon shale at Eureka.

Collection F–56 contains a distinctive kind of inarticulate brachiopod known at present only from Middle Cambrian rocks. It is not possible to determine from the fossils whether this collection is older or younger than the others.

These are important occurrences, worthy of further collecting.

Further search failed to find additional fossils at these localities.

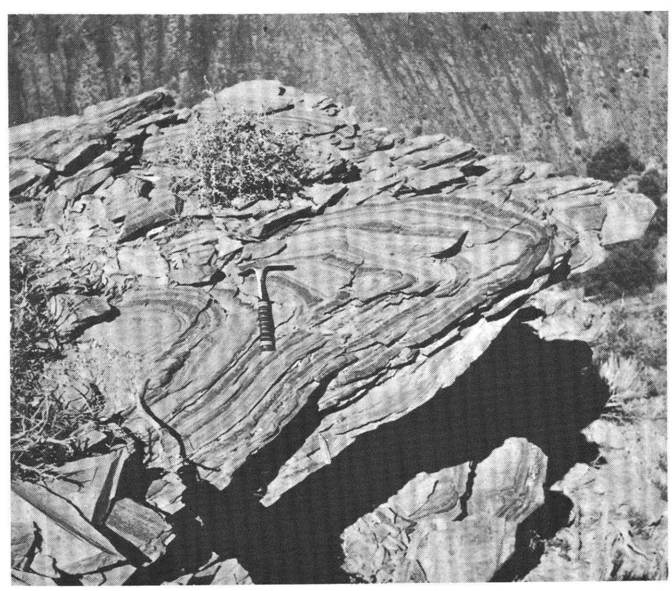

FIGURE 4.—Small outcrops of phyllitic Shwin Formation on the south slope of Trout Creek canyon, showing crumpling and incipient fissility to the left of hammer. This mass lay probably 100 to 200 feet in the footwall of the Roberts thrust, judging from projections from nearby klippes.

TABLE 3.—*Semiquantitative spectrographic analyses, in grams per metric ton, of Paleozoic sedimentary rocks, Mount Lewis and Crescent Valley quadrangles, Nevada*

[0, looked for but not found. Analyst: P. R. Barnett. See table 2 for description and location of specimens]

	1	6	7	8	9	10	11	13	14	15
Field No	272	1	4	245	19	225	15	608	264	393
B	30	0	15	0	30	30	15	70	30	70
Ba	300	3,000	7,000	30	30	700	70	700	1,500	700
Co	0	0	0	0	0	0	0	0	0	7
Cr	15	1.5	1.5	1.5	1.5	15	1.5	15	15	30
Cu	1.5	15	15	15	7	70	7	150	15	70
Ga	1.5	0	0	0	0	0	0	1.5	1.5	7
La	0	0	0	0	0	0	0	0	0	30
Nb	0	0	0	0	0	0	0	15	0	15
Ni	7	0	0	15	0	15	3	0	15	30
Pb	15	0	0	0	0	0	0	0	7	0
Sc	0	0	0	0	0	0	0	7	0	15
Sn	0	0	0	0	0	0	0	7	0	0
Sr	70	7	15	0	3	30	0	30	70	15
V	7	7	15	0	0	70	0	15	15	30
Y	7	0	0	0	0	7	0	15	7	15
Yb	1.5	0	0	0	0	1.5	0	3	1.5	3
Zr	300	150	70	70	70	70	0	1,500	700	150

Looked for but not detected: Ag, As, Au, Be, Bi, Cd, Ce, Dy, Er, Eu, Gd, Ge, Hf, Hg, Ho, In, Ir, Li, Lu, Mo, Nd, Os, Pd, Pr, Pt, Rb, Re, Rh, Ru, Sb, Sm, Ta, Tb, Te, Th, Tl, Tm, V, W, Zn.

Though, as noted by Palmer, the fossils are rather poorly preserved, their similarity to those from the Secret Canyon Shale at Eureka, combined with the similarity of the calcareous mudstone and shaly limestone of the Shwin Formation to those of the Secret Canyon, makes correlation of the Shwin and Secret Canyon very probable. If the metavolcanic rocks actually are interbedded with the fossiliferous beds, we may infer that the Shwin Formation is transitional between the eastern facies Secret Canyon Shale and a truly siliceous and volcanic (western facies) deposit.

Volcanic rocks of Early Cambrian age abound in the cherty Scott Canyon Formation in the Antler Peak quadrangle immediately northwest of the map area, but have not yet been found in rocks of Middle Cambrian age farther west in this region (Roberts and others, 1958, p. 2826–2830).

The question may well be raised, in view of the abundant volcanics in the Shwin Formation, as to why it is referred to the eastern rather than western facies. If our assignment is correct and the formation is autochthonous or parautochthonous, it is obvious that the transition between carbonate and siliceous facies took place farther east in Middle Cambrian time than in either earlier or later parts of the Paleozoic. There is nothing inherently less probable about this hypothesis than about the alternative: that the Shwin is of true western facies and wholly allochthonous in the Shoshone Range. And there are cogent reasons in its favor.

In favor of assigning the Shwin Formation to the eastern facies are:

1. The marked similarity in lithology of much of the Schwin to the Secret Canyon and Geddes.
2. The intimate structural association of the Shwin with such undoubted representatives of the eastern facies as the Eldorado Dolomite, Roberts Mountains Limestone, Hanson Creek Dolomite, and Eureka Quartzite.
3. The absence of the Shwin from the many intercalations of thrust sheets of undoubted western facies such as those of Valmy with Slaven Chert and Elder Sandstone. This is a most significant point, in our opinion. It is easy to find a single thrust that divides the lower and middle Paleozoic rocks into two suites, the one of virtually all siliceous rocks, the other of virtually all carbonate rocks (except for the Shwin). The Shwin Formation is everywhere on the same side of this thrust with the other carbonate-rich formations. We therefore consider the Shwin, despite its important volcanic content, as part of the eastern facies. It is here parautochthonous; though doubtless dislodged, there is no evidence whatever to suggest long thrust transport from the site of its deposition.

UPPER CAMBRIAN-LOWER ORDOVICIAN HIATUS

No rocks of Late Cambrian or Early Ordovician age are known in the map area. In the classic Eureka section the Secret Canyon Shale—the probable equivalent of the Shwin Formation of this area—is followed by the Hamburg Dolomite (about 1,000 ft), Dunderberg Shale (265 ft), Windfall Formation (650 ft) and Pogonip Group (about 1,600–1,950 ft), beneath the Eureka Quartzite (Nolan, Merriam, and Williams, 1956, p. 16–25). No correlatives of any of these pre-Eureka rocks have been recognized in the Shoshone Range beneath the Roberts thrust. The Eureka Quartzite is the oldest Ordovician formation of the eastern facies in this part of the Shoshone Range. Thus more than 3,000 feet of eastern facies beds in the Eureka district, representing all Late Cambrian and nearly all Early Ordovician time, are here absent.

As all the upper contacts of the Shwin and the basal contact of the Eureka are faults, it is possible to explain the local absence of the Upper Cambrian and Lower Ordovician rocks as due to faulting. The regional relations, however, suggest that the hiatus is more probably due to unconformity.

The Eureka Quartzite rests on progressively older rocks for at least 40 miles northwestward from the vicinity of Eureka. In Antelope Valley it rests on the Copenhagen Formation, a unit younger than the Antelope Valley Limestone (Nolan, Merriam, and Williams, 1956, p. 28); at Eureka, on various horizons of the Antelope Valley Limestone; in the Roberts Creek Mountains, on the Goodwin Limestone; and at Cortez, on the probable Hamburg Dolomite (p. 30). (See fig. 1 and pl. 3 of this report.) In view of the northward overlap thus demonstrated, it would not be surprising to find that the Eureka in the area of the Goat window was originally deposited on Middle Cambrian rocks, perhaps on the Shwin Formation.

The Eureka Quartzite, though widespread in the Great Basin, is absent at Gold Hill (Nolan, 1935, p. 16) and in the Stansbury and Lakeside Mountains, Utah (Kirk, 1933, p. 38). In the southern Ruby Mountains, Nev., it has been reported as absent (Sharp, 1942, p. 659, 680) but it is present west of Sherman Peak in that range (R. J. Ross, Jr., 1961, p. 333). Whether its absence in the localities mentioned is due to later pre-Hanson Creek erosion or to nondeposition because of considerable pre-Eureka relief is uncertain. But the indisputable evidence of Eureka overlap between the Monitor Range and Cortez shows that the base of the Eureka rests on a notable regional unconformity. No angular discordance has been reported at the base of the Eureka Quartzite anywhere in the region, but except in the Eureka district proper, the scale and detail of mapping have not been sufficient adequately to test conformability in areas of less complex structure than that of the Goat window.

The unconformity may represent either a single post-Copenhagen uplift, planed off by the encroaching Eureka sea, or the coalescence of several smaller unconformities. Cooper (1956, p. 125–128 and chart 1) has interpreted the faunas of the strata since named the Copenhagen Limestone in the Antelope Valley area to indicate a considerable intraformational hiatus, equivalent to most of his Porterfield and Wilderness Stages. He also infers a nondepositional interval beneath the lower part of the Copenhagen and above the Antelope Valley Formation, amounting to the whole of his Marmor and much of his Ashby Stage.

Local evidence cannot indicate how much of the hiatus in the more northerly exposures is due to nondeposition during times corresponding to the pre- and inter-Copenhagen disconformities of Cooper and how much to erosion immediately prior to Eureka deposition. Possibly the intra- and pre-Copenhagen gaps inferred by Cooper widen northwestward; equally possibly a significant break began in this area only immediately before Eureka time.

ORDOVICIAN SYSTEM

MIDDLE ORDOVICIAN SERIES

EUREKA QUARTZITE

Name, distribution, and topographic expression

The Eureka Quartzite was named by Hague (1883, p. 253, 262) from exposures in the Eureka district. In the northern Shoshone Range a single mass of Eureka Quartzite crops out in the ridge that divides the steep head of Hancock Canyon, in sec. 20, T. 29 N., R. 45 E. The mass is about 700 feet long and 150 feet wide, and forms a straight, rugged ridge crest.

Stratigraphy

The quartzite stands nearly vertical and faces south. The basal contact is a high-angle fault against the Roberts Mountains Limestone to the north of the ridge. Presumably a considerable thickness of the lower part of the quartzite has been eliminated by the fault because there are here no brown-stained or pyritic members such as compose the lower third of the Eureka over wide areas to the east and south.

The formation is remarkably homogeneous both along and across the bedding. It is massive bedded, in ledges 4 to 8 feet thick, gray on weathered surfaces but vitreous white on fresh fracture. Grains roughly half a millimeter across are readily visible, but the rock is a true quartzite, and breaks across them.

The upper, south contact of the formation is locally apparently conformable with the overlying dolomite of the Hanson Creek, as is the rule in areas to the east and south of the northern Shoshone Range. This appearance may, however, be deceptive, as even in this short mass, as much as 80 feet of beds can be seen to cut out against an oblique, near-bedding fault.

The exposed thickness of the Eureka Quartzite is about 150 to 200 feet. Perhaps the original thickness was half again as great, as it is in nearby ranges to the southeast.

Petrography

Most of the Eureka Quartzite is a dense white rock that shows no obvious minerals other than quartz. It is a moderately well-sorted medium sand-sized rock containing a little coarse sand and some silt-sized grains. Although it breaks across the grains with a clean conchoidal fracture, many of the original grains can be recognized. (See fig. 5.) They are generally well rounded, though overgrowths render the texture granoblastic. Many rocks have been sheared so that undulatory extinction is common and many grain boundaries are sutured or show microstylolites. No feldspar has been noted, either in hand specimen or

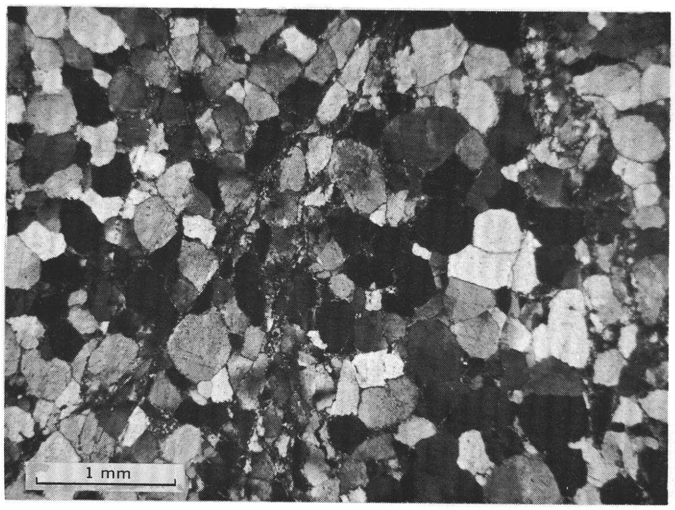

FIGURE 5.—Photomicrograph of quartzite of the Eureka Quartzite, Hancock Canyon Shoshone Range. Crossed nicols. Note rehealed breccia zones and mosaic texture.

under the microscope. Aside from a little tremolite and diopside, which are doubtless not detrital but metamorphic minerals formed in the aureole of the small masses of quartz diorite on the ridge west of Goat Peak, the only accessory minerals seen are magnetite, staurolite, tourmaline, kyanite, apatite, zircon, brown hornblende, and biotite; all together make up less than 1 percent of the rock.

In comparison with the pure quartzites of the Valmy Formation, some of which are of virtually the same age, the Eureka Quartzite has very similar mineral composition and grain shape but is better sorted. Both formations probably had the same source, but the sands of the Eureka were better winnowed, perhaps on a shallower marine platform. Corals in the Eureka of the Cortez area suggest a relatively shallow sea, as does the sweeping crossbedding common in the lower part of the Eureka in the general region.

Age

No fossils have been found in the Eureka Quartzite of the northern Shoshone Range. The age of the formation is therefore determinable only by bracketing between fossiliferous formations. A review of the question in the Eureka area indicates an age between late Chazyan and Richmondian, thus about Middle Ordovician (Nolan, Merriam, and Williams, 1956, p. 31–32).

MIDDLE AND UPPER ORDOVICIAN SERIES

HANSON CREEK FORMATION

Name, distribution, and topographic expression

The Hanson Creek Formation was named by Merriam (1940, p. 10–13) from exposures on Pete Hanson Creek in the Roberts Creek Mountains. In the northern Shoshone Range, the Hanson Creek Formation is confined to the two westernmost windows through the Roberts thrust: the Horse Mountain and the Goat windows (fig. 29). In both it is conspicuous by reason of its massive, thick beds, which are highly resistant to erosion and form prominent cliffs and benches. In the Horse Mountain window there are also large areas of dark-gray limestone forming smooth, more gentle slopes.

Stratigraphy

The lower contact of the Hanson Creek Formation, where not faulted, appears to be conformable—perhaps even transitional through a few inches of sandy dolomite—with the underlying Eureka Quartzite in the Goat window. In the Horse Mountain window the dolomite beds of the Hanson Creek form horses along a flat and gently dipping fault that separates the limestones of the formation from the Roberts Mountains Limestone. Here the original stratigraphic relations are impossible to decipher and it is uncertain whether the dolomite or the limestone belongs lower in the formation. The base is here concealed.

The most readily notable features of the Hanson Creek Formation are the very dark or very light gray colors of its dolomite beds—nearly every bed is either conspicuously dark gray or very pale gray. This color banding of the generally conspicuous ledges sets the formation apart from any other of the carbonate formations. The limestone beds are less conspicuous, as they are normally light blue gray and in beds a few inches thick.

The dolomites of the formation are chiefly rather coarse, of about a millimeter grain size, and some appear sandy on weathered surfaces. Some beds, especially toward the base, do indeed contain a few percent of rounded quartz grains but most of the "sandy" appearance is due to the coarse granularity of the carbonate composing the rocks.

These beds generally range from 3 to 10 feet in thickness, and the bedding partings are inconspicuous except for the color differences on the two sides.

The limestones of the formation, which have not been recognized in the Goat window but are abundant in that of Horse Mountain, are light blue gray on weathered surfaces but dark gray, almost black, on fresh fracture. They are aphanitic, in beds that range from 1 to perhaps 6 inches in thickness and are separated by thin interbeds of shaly limestone ⅛ to ¼ inch thick that weather to buff surfaces along the parting planes. The mottled weathered surfaces and the absence of sooty carbon on the bedding planes suffice to distinguish the limestones of the Hanson Creek from those of the Roberts Mountains Limestone. The limestones of the Hanson Creek are also abundantly fossiliferous, with many trilobites and some gastropod fragments.

The formation is overlain by the Roberts Mountains Limestone in apparent conformity in the Goat window; the contact is a fault at Horse Mountain.

Thickness

Faults mark either or both the top and bottom of the Hanson Creek Formation everywhere in this area. It is therefore impossible to measure the thickness of the formation, but in several sections in the Goat window at least 300 feet of dolomite referable to it are exposed. No limestone has been recognized in the formation here. As an estimate, a thickness of 500 to 600 feet seems not unreasonable—about the same as at the type locality. In the Horse Mountain window the thickness is also difficult to measure because of faulting and shearing; an estimate is 600 or 700 feet, of which most is limestone.

Age and correlation

Although a few fossils were seen in the dolomite of the Hanson Creek Formation, they were difficult to free from the matrix without fracture. Among the recognizable organic remains, cryptolithid trilobite fragments, indeterminable coralline debris, a conularid, and an orthoconic cephalopod were found, but none were identifiable even generically. The limestone beds of the Horse Mountain window are, however, abundantly fossiliferous.

Trilobites collected from the southernmost hill spur in the Horse Mountain window (F–60 and F–66) were referred to Prof. H. B. Whittington of Harvard University, who reported (letter of Feb. 27, 1956, to R. J. Ross, Jr.) that they have characters strongly suggesting *Cryptolithus* sp., comparable to *C. carinatus* or *C. convexus*; they do not seem to be *Cryptolithoides*. They resemble *Cryptolithus* from the Copenhagen of the Roberts Mountains quadrange (old sheet, 1: 250,000; the same area is now mapped on a scale of 1: 62,500 on the Horse Heaven Hills quadrangle).

Other fossils (F–60, F–62, and F–66) identified from this locality by R. J. Ross, Jr., of the U.S. Geological Survey, include *Bumastus* of similar age significance. A cephalopod from a nearby outcrop (F–64) could only be identified as an immature orthoconic variety. A. R. Palmer identified some inarticulate brachiopods from a point in the next gulch to the north, (F–61) as of acrotretid type, significant only of an early Paleozoic age.

The "Copenhagen formation" referred to by Professor Whittington has been described by Merriam (1963) from the Monitor Range-Antelope Valley region. These beds were long ago considered by Kirk (1933, p. 28, 30, 32–34) and Merriam (Merriam and Anderson, 1942. p. 1684–1685) as a lateral equivalent of the lower part of the Eureka Quartzite in areas farther east. Webb (1958, p. 2339–2340) has described an estimated 600 feet of sandstone, limy shale, siltstone, and fine and coarse limestones beneath the Eureka Quartzite in Copenhagen Canyon in the Monitor Range which he refers to the Copenhagen Formation, and like Kirk and Merriam, considers equivalent to the lower part of the Eureka. It is with the fossils of these beds that Whittington compares the *Cryptolithus* collected in the Horse Mountain window.

It should, however, be pointed out that *Cryptolithus* has long been known from the Hanson Creek Formation in the Roberts Creek Mountains, where it is unquestionably stratigraphically above the Eureka Quartzite (Nolan, Merriam, and Williams, 1956, p. 33; Webb, 1958, p. 2344). It is also in similar position near Wood Cone, in the Eureka district whence Hague (1892, p. 59) reported "*Trinucleus*," which Merriam (op. cit.) thinks is probably a *Cryptolithus* sp.

There seems little doubt of the correlation of the formation near Horse Mountain and in the Goat window, here mapped as Hanson Creek, with the rocks of the type section in the Roberts Creek Mountains. The few fossil fragments seen resemble, in assemblage, the identifiable fossils of the more easterly locality. But much more compelling is the sequence of conspicuously banded dolomite virtually conformably above the Eureka Quartzite in the Goat window and below the platy beds of the Roberts Mountains Limestone in both localities. The limestone of the Horse Mountain window is indistinguishable from that in the middle of the Hanson Creek at Cortez, where that formation rests directly on the Eureka. As noted by Merriam (Nolan, Merriam, and Williams, 1956, p. 33–34), the characteristic lithology of the Hanson Creek dolomite is largely like that of the equivalent Ely Springs and Fish Haven Dolomites of the southern and northeastern Great Basin, respectively; it resembles also the dolomite beds of the Saturday Mountain Formation of central Idaho, though it lacks the shale associated with the thicker sections of that formation (Ross, C. P., 1934, p. 952–956; 1937, p. 18–22). Here in the northern Shoshone Range we have still another characteristic section of the formation. If the trilobites permit a pre-Eureka age, as Professor Whittington suggested, we have here and in the Roberts Creek Mountains and probably also at Wood Cone evidence for extending their range to include both Eureka and Hanson Creek times.

SILURIAN SYSTEM

ROBERTS MOUNTAINS LIMESTONE

The Roberts Mountains Formation was named by Merriam (1940, p. 11–13) from the Roberts Creek

Mountains where it lies between the Hanson Creek Formation and the Lone Mountain Dolomite. In the area of this report the formation consists dominantly of limestone, so that this name is preferred for formal usage.

Distribution and topographic expression

The Roberts Mountains Limestone crops out in each of the four largest windows through the Roberts thrust (fig. 29). In the Horse Mountain window it forms a crescentic outcrop on the slopes above the Hanson Creek Formation. The entire area of the small window southwest of the forks of Mill Creek is occupied by Roberts Mountains Limestone, in an unusually shaly facies. The topographic eminence of this window can only be explained by its geologically very brief exposure to erosion, for the rock composing it is very soft and weak. Fully a third of the Goat window is underlain by this formation, which forms more rolling and smoothly sculptured slopes than the associated Hanson Creek Formation. Although most contacts are faults, it is perhaps significant that the main mass of Roberts Mountains Limestone in the Goat window apparently underlies most of the Hanson Creek exposed there—the whole section is apparently upside down. Perhaps the inversion is due to thrusting within the lower plate of the Roberts thrust but it may be a completely overturned fold—exposures are too discontinuous for certainty. In Gold Acres window the formation crops out on both walls of the gulch that runs southeast from the mine.

Stratigraphy

The lower contact of the Roberts Mountains Limestone is everywhere a fault except perhaps in the canyons near the Shwin Ranch, where contacts possibly concordant with the Hanson Creek dolomite are poorly exposed. No beds analogous to the basal chert described by Merriam (1940, p. 12; Nolan, Merriam and Williams, 1956, p. 36–37) from the type locality have been recognized in this area, but such beds are also lacking at Cortez where the apparently conformable base is well exposed.

The Roberts Mountains Limestone exhibits notable though not drastic facies differences in the several areas of outcrop. All the rocks, like most of those in the type locality, tend to weather into thin, fissile plates.

In the Gold Acres window, the rock, though platy, is nearly all carbonate and carbon. Much of it is nearly black on fresh fracture, with enough carbon along the bedding planes to smear the fingers. Most of the rock is thin bedded, though there are a few beds 1 to 3 feet thick.

In the Goat window, near the Shwin Ranch, and northwest of it, the Roberts Mountains Limestone is much the same but contains relatively more beds 2 to 4 feet thick. These thicker beds are generally light gray on weathered surfaces but are almost black on fresh fracture. They, too, contain almost as much carbon as the thinner platy beds, and like them tend to break down to shinglelike chips. Both of these facies of the formation are readily matched by many beds at the type locality.

In the small Mill Creek window, southwest of the forks of Mill Creek, and in the Horse Mountain window farther southwest the formation is notably different. Though still highly calcitic, and conspicuously thinly laminated, in neither locality is the rock black on fresh fracture nor does it contain much carbon. The rocks of both areas weather yellowish red or reddish brown, rather than gray. Doubtless in both they contain considerably more clastic material than elsewhere in the range, though not more than is generally connoted by the term "shaly limestone"; neither is a true shale or siltstone. The contrast in lithology of this formation between the exposures in the window southwest of the forks of Mill Creek and those in the Goat window, less than a mile away, is striking, though not enough to cast doubt on the virtual equivalence of the two sequences. It constitutes part of the argument for a fault of significant displacement within the lower plate of the Roberts thrust (p. 110).

Thickness

Structural complications are such as to prevent a satisfactory measurement of the Roberts Mountains Limestone in this area. Faults everywhere limit either top or bottom or both. An estimate of 600 to 1,000 feet of strata in the Horse Mountain and Goat windows seems reasonable; at the other localities, less. This thickness is about half that at the type locality (1,900 ft). Perhaps the absence of dolomite from the Roberts Mountains Limestone in this area is due to the absence of the upper part, which is the dolomitic part in the Roberts Creek Mountains, but the general tendency of the Ordovician, Silurian, and Devonian rocks of the carbonate facies to become more calcitic westward (Nolan, Merriam, and Williams, 1956, p. 40) is so consistent that it seems equally or more plausible that the dolomite of the eastern localities has graded westward into limestone.

Age and correlation

Fossils are not abundant in the Roberts Mountains Limestone in this area, but enough have been found to establish the equivalence of these beds with the more fossiliferous section at the type locality. On the south side of the Goat window at a locality 600 feet south, 1,600 feet east of the NW cor. sec. 33, T. 29 N., R. 45 E., and an altitude of 7,030 feet, a collection (F54) was obtained and submitted to Jean Berdan of the U.S. Geological Survey, who reported as follows.

Favosites sp., large corallites
Favosites sp., smaller corallites
Heliolites sp.
Thamnopora? sp.
Tryplasma sp.
Columnariid coral
Cystiphylloid coral
Miscellaneous unidentifiable horn corals
Stromatoporoids?

This fauna is of Niagaran age, probably roughly equivalent to the Laketown Dolomite of Utah (Nolan, Merriam, and Williams, 1956, p. 37), though it is possible that the lower part of the Roberts Mountains Limestone may be slightly older. Perhaps it correlates in part with the Trail Creek Formation of central Idaho (C. P. Ross, 1934, p. 956–957; 1937, p. 22) which locally underlies the rocks there correlated with the Laketown.

R. J. Ross, Jr., examined two collections of graptolites and reported as follows:

F–67=D18SD, from platy sandy limestone at alt 7,400 ft on spur of Horse Mountain 1,150 ft N., 200 ft W. of SW cor. sec. 19, T. 28 N., R. 45 E. Collected by James Gilluly.

Large specimens of *Monograptus*. The thecal features are very poorly preserved, despite profusion of specimens. Probably equals *M. pandus* Lapworth, which ranges through zones 22–24 of the British section. Tarranon or high lower Silurian.

F–79=D20SD from platy shaly limestone near hilltop at alt 6,820 ft, 600 ft N., 2,550 ft W. of SE cor. sec. 4, T. 28 N., R. 45 E. Collected by C. W. Merriam.

Two species of *Monograptus*, not identifiable subgenerically: Silurian.

Other specimens of badly smeared monograptids(?) were collected from dark carbonaceous platy limestone in the canyon about 1,500 feet upstream from the Shwin ranchhouse, but these were not even generically identifiable.

HIATUS BETWEEN SILURIAN AND DEVONIAN SYSTEMS

As noted above, everywhere in the northern Shoshone Range the Roberts Mountains Limestone is cut off by faults at the top. In the Roberts Creek Mountains to the southeast (Merriam, 1940, p. 13–14, Merriam and Anderson, 1942, p. 1687–1688), not only is the Roberts Mountains Formation thicker than it is here—1,900 feet as against at most 1,000 feet—but it is separated from the Devonian limestones by 2,200 feet of dolomite referred to the Lone Mountain Dolomite of Silurian age. In the northern Shoshone Range we know of no rocks resembling either the dolomitic upper part of the Roberts Mountains Formation of the type locality (Merriam and Anderson, 1942, p. 1687) nor the Lone Mountain Dolomite.

The Gold Acres window exposes some limestones that are lithologically so distinctive that their reference to the Roberts Mountains Limestone is unquestionable. Other limestones in this window are lithologically much like unnamed but known Devonian rocks of the Cortez Mountains and have yielded fossils of doubtful late Middle or Late Devonian age. But there remain considerable areas underlain by limestones of uncertain age, mapped as Devonian (pl. 1). The scarcity of fossils in these rocks is attributed to the widespread brecciation and low-grade metamorphism and not to original conditions of deposition or to dolomitization. The Devonian age of these rocks is highly probable on grounds of regional similarity, but they could belong to any part of the system.

The absence here of formations that elsewhere intervene between the lower part of the Roberts Mountains Limestone and the Devonian may be due either to unconformity, facies change, or simply to elision by faulting. Faults abound and the structural complexity is such that they could easily account for the apparent stratigraphic gap. Neither of the other possibilities can be eliminated, however.

Facies changes from dolomite in the east to limestone in the west have been noted elsewhere in the section in this general region (Nolan, Merriam, and Williams, 1956, p. 33, 37, 39, 41). Indeed Murphy and Winterer have observed a lateral change of the Lone Mountain Dolomite into limestones which closely resemble those of the Roberts Mountains within a few thousand feet of the type locality of the Roberts Mountains Formation (Winterer and Murphy, 1960). The absence, then, of both the dolomitic upper beds of the type Roberts Mountains and of the overlying Lone Mountain Dolomite—as well as the ostensible absence of Lower Devonian beds—may be wholly or partly due to facies change.

There is, however, widespread evidence in the Great Basin of an unconformity at the base of the Devonian (Nolan, 1935, p. 18; Nolan, 1943, p. 143; Calkins and Butler, 1943, p. 18–19; Gilluly, 1932, p. 18; Hazzard, 1938, p. 327–328; Nolan and others, 1956, p. 41). But in neither the Monitor Range, the Roberts Mountains, nor the Sulphur Springs-Pinon Range areas has any unconformity been recognized at this horizon (Merriam, 1940, p. 14; 1954; Carlisle, Murphy, Nelson, and Winterer, 1957, p. 2180). Accordingly the probability that a facies change rather than unconformity accounts for the local hiatus is enhanced as these areas are nearer the northern Shoshone Range than those from which unconformity has been reported. Only further detailed work in this and nearby areas will allow a decision as to these three possibilities.

DEVONIAN SYSTEM

LIMESTONE UNDIFFERENTIATED

Name

The name Nevada Limestone was formerly applied to the entire section of Devonian rocks of east-central Nevada (Hague, 1892, p. 70–84; Walcott, 1884, p. 4–8; Merriam, 1940, p. 14). Merriam restricted the name in the type area, near Eureka, to the beds between the top of the Lone Mountain Dolomite and the top of the *Stringocephalus* zone, and referred the higher Devonian beds to a new formation, the Devils Gate (Merriam, 1940, p. 14–16).

This restriction has created a problem in nomenclature. Where such faunal subdivisions as the *Stringocephalus* beds are not recognizable because of structural complexities or metamorphism the division cannot be made. Perhaps biofacies changes alone prevent recognizing this boundary, as seems to be the case in the Cortez area. In effect, Nolan, Merriam, and Williams (1956, p. 47) revised the boundary in the Diamond Mountains from the faunal one originally designated by Merriam to a dolomite-limestone contact. This revised dolomite-limestone boundary cannot be recognized in either the Cortez area or in this one. Carlisle, Murphy, Nelson, and Winterer (1957, p. 2177, 2187–2188) believe that the lithologic boundary lies at different faunal horizons at different places, as would perhaps be expected. Many of the lithologic members recognized by Nolan, Merriam, and Williams (1956, p. 42) in the restricted Nevada east of Eureka appear to be local and not generally recognizable even close at hand, as is stated in their paper (p. 43—Oxyoke Canyon Sandstone Member; p. 44—Sentinel Mountain Dolomite Member; p. 44—Woodpecker Limestone Member; p. 45—Bay State Dolomite Member). The lithologic divisions feasible in the Sulphur Springs-Pinon Range area 40 miles north of Eureka seem fewer and also differ in lithology (Carlisle and others, 1957, p. 2181). It now appears that the formation boundary proposed between Nevada (as restricted) and Devils Gate also is usable only near Eureka. All formational boundaries are of course similarly restricted in some degree and this statement is not a criticism of the useful local classification of these authors. The limitation does, however, create a nomenclatural problem in the surrounding area of which this is a part. The restriction of the name Nevada to a part of the strata formerly included under it has left us with no name to apply to these thick sections of Devonian limestone in which the distinctions recognized near Eureka and the Roberts Creek Mountains cannot be made. Youngquist (1949, p. 276) has commented in similar vein with respect to the areas from the Pancake Range to the Utah line and has therefore persisted in using Nevada Limestone in the wide sense of Hague. This is also the usage of Humphrey (1960). It now appears unfortunate that Nevada was not retained as a group name when the subdivisions recognized near Eureka were established. Such a group term will long be useful and one will doubtless eventually be recognized. The raising of Nevada to group status, with Devils Gate one of its included formations, has also been advocated by Carlisle, Murphy, Nelson, and Winterer (1957, p. 2189).

Obviously the highly faulted and even brecciated area of the Gold Acres window is hardly appropriate for establishing the type section of a new formation or group. The Devonian rocks are far less disturbed and altered in the Cortez area than here; a new classification for this area should be deferred until that nearby section has been adequately studied. As the use of the term "Nevada" in the former wide sense would at present be confusing, and the assignment of all the local strata to either of the restricted successor formations recognized near Eureka seems impossible, or at least premature, the limestones of Devonian age in the eport area are not given a formation name. They are herein referred to simply as "limestone of Devonian age."

Distribution

Almost all the hill in secs. 30 and 31, T. 28 N., R. 47 E., east of the Gold Acres mine, is underlain by linestone of Devonian age. These rocks also occur on the low hills across the valley to the northeast in secs. 30 and 19, and on the south border of the map area, near the southwest corner of the Crescent Valley quadrangle.

Stratigraphy

All the contacts of this body of rock are faults, and innumerable unmapped faults also dissect the mass. The rocks in this block—just beneath the Roberts thrust—have such diverse attitudes and their mutual relations are so obscure that neither base nor top can be identified, nor can any statement confidently be made regarding their sequence. As far as tested the rocks are all limestone, and almost no siliceous constituents or dolomite are megascopically recognizable. There is considerable variation in bedding thickness. Bedding is generally thicker—from 4 inches to 4 feet—than in the Roberts Mountains Limestone, and the outcrops weather to lighter gray tones. On fresh fracture the rocks are dark gray. Most are aphanitic or fine grained, but a few beds are coarsely crystalline. Fossils are sparse.

Thickness

The thickness exposed is impossible to measure, owing to the confused structure. There are several single sections, each a few hundred feet thick, but the extent to which they represent duplication is completely unknown. As a guess, 1,000 feet of strata may be exposed here, but, whatever the thickness, it surely does not represent the original depositional total.

Age and correlation

The lithologic identity of these rocks with the abundantly fossiliferous Devonian rocks of the Cortez Mountains, and the differences between them and any rocks of the region known to be of different age leaves little doubt as to their general age assignment.

One poor collection of badly broken fossils (F–128) from a point 500 feet northwest of the bunkhouse of the Gold Acres mine in sec. 31, T. 28 N., R. 47 E., was referred to Jean Berdan, who identified only echinoderm debris and a few fragments of spiriferoid brachiopods and bryozoa. This fossil evidence indicates a post-Ordovician age—probably Silurian or Devonian.

Another collection (F–129) made by R. J. Ross, Jr., from a point 1,600 feet north, 250 feet east of the road intersection whose indicated altitude is 5,221 feet, in sec. 30, T. 28 N., R. 47 E., was referred to the late W. H. Hass, of the U.S. Geological Survey. He identified:

Ancyrodella
Bryantodus
Icriodus
Ligonodina
Nothognathella
Palmatolepis
Polygnathus cf *P. pennata* Hinde
Prioniodus cf *P. alatus* Hinde
Spathognathodus

These fossils, all conodonts, in the opinion of Mr. Hass indicate an early Late Devonian age.

As emphasized above, it would be premature to assign all the limestones in this particular structural block to like age; they may well include both older and younger strata, though all are probably Devonian.

DEVONIAN AND MISSISSIPPIAN SYSTEMS

PILOT SHALE

Name

The Pilot Shale was named by A. C. Spencer (1917, p. 26) from the Ely district. It is generally considered equivalent to the lower beds of the White Pine Shale as that formation was defined by Hague (1892, p. 68–69). Although it has been classed as a member of the White Pine Shale as recently as 1953 (Strat. Comm., Eastern Nevada Geol. Assoc., 1953, p. 147, 149), its lithological distinction from the rest of that formation is clear enough (even where the Joana Limestone is absent) to warrant recognizing it as a separate formation, as proposed by A. C. Spencer and reiterated by Nolan, Merriam, and Williams (1956, p. 56–58).

Distribution

In the area of this report the Pilot Shale is confined to the Gold Acres window. Here it occupies an area of about 50 acres along the west side of the stream that drains southward through the centers of secs. 18, 19, and 30, T. 28 N., R. 47 E., to join the drainage from the Gold Acres mine. Its outcrops form a rolling topography in contrast to the bolder features carved on the nearby limestones of Devonian age and on the Valmy, Elder, and Slaven Formations.

Stratigraphy

Although both base and top of the formation are faulted, it seems probable that not much of the lower part is missing. This inference is based not only on the fact that the thickness represented here—perhaps 300 to 400 feet—is virtually as great as that commonly found near Eureka (Nolan, Merriam, and Williams, 1956, p. 52), but also on the fact that the only fossil found is a Late Devonian form, here found about 50 feet stratigraphically above the basal fault. Inasmuch as the Mississippian-Devonian boundary in the more easterly sections lies within this shale unit (op. cit., p. 53), as was indeed recognized long ago by Walcott (1884, p. 5), the local occurrence of fossils of definite Devonian age suggests a horizon low in the formation. The possibility of lateral thickening in the long distance between Eureka and Gold Acres of course prevents certainty in this inference.

The rocks of the formation are nearly all limy shale, dark gray on fresh fracture, but weathering to pinkish and light yellowish brown or purplish brown. The formation contains a few thin limestones, and micaceous parting planes are conspicuous. Its resemblance to the Pilot Shale as exposed in the Diamond Range northeast of Eureka is striking, in view of the long intervening distance.

Age

The only fossil (F-141) found in the Pilot Shale during this survey was a single conodont fragment identified by W. H. Hass as *Palmatolepis subrecta* Miller and Youngquist. The locality is at altitude 5,360 feet, 2,500 feet north, 400 feet west of the road intersection (alt 5,221 ft) in sec. 30, T. 28 N., R. 47 E. This genus has also been found in the Pilot Shale in the Eureka area (Nolan, Merriam, and Williams, 1956, p. 53). Mr. Hass stated:

> I consider this species to indicate an Upper Devonian age, though it, as well as the genus to which it belongs, has been found associated with Lower Mississippian conodonts in some collections from the Llano region of Texas. I regard its associa-

tion with Lower Mississippian conodonts to have resulted through reworking, though others would extend the stratigraphic range of the genus to include beds of Lower Mississippian age.

Palmatolepis subrecta is present in a faunal zone of the Chattanooga shale of central Tennessee and adjacent States. It is also present in the Arkansas Novaculite at Caddo Gap, Montgomery County, Ark., where it is located 184 feet below the top of the middle division of the Arkansas Novaculite. This portion of the novaculite as well as the Chattanooga shale of central Tennessee and adjacent States is classified as Upper Devonian by the U.S. Geological Survey. *Palmatolepis subrecta* or a very closely related species is present in the basal beds of the Dunkirk shale member of the Perrysburg formation of western New York; the species may also be conspecific with *Palmatolepis flabelliformis* Stauffer from the Olentangy shale of Ohio. The Olentangy shale and the Dunkirk shale are classified as Upper Devonian by the Survey. *Palmatolepis subrecta* Miller and Youngquist was first described on material obtained at the type locality of the Sweetland Creek shale near Muscatine, Iowa. The Survey classifies the Sweetland Creek shale as Upper Devonian or Mississippian but it is my opinion that the beds from which Miller and Youngquist's conodonts came are Upper Devonian.

If these inferences and correlations are correct, it may be that the Mississippian boundary is not represented in the local area, though it evidently is present in somewhat thicker sections farther east.

STRATA OF THE UPPER PLATE OF THE ROBERTS THRUST (SILICEOUS OR WESTERN FACIES)

CAMBRIAN SYSTEM

UPPER CAMBRIAN SERIES

HARMONY FORMATION

Name and distribution

The Harmony Formation was named from Harmony Canyon, in the northern Sonoma Range, 35 miles to the west of this area (Ferguson, Muller, and Roberts, 1951). Only one small body referable to this formation was recognized in the northern Shoshone Range. This is a thin thrust sheet high on the northwest slopes of Mount Lewis, just east of the head of the west fork of Lewis Canyon and traceable for about half a mile along the strike. This thrust sheet lies on the Whisky Canyon fault, of probable Mesozoic age, and thus although it is now in the hanging wall of the Roberts thrust, it was not transported to its present position by that thrust. Nevertheless, it clearly belongs to the siliceous facies, whose major representatives in this area have been transported by the Roberts thrust, and therefore is most conveniently described here.

Petrography

The Harmony Formation is composed of most remarkable arkose and arkosic siltstone. Grain sizes range from fine conglomerate to silt and include abundant coarse sand. Grain shapes are angular to subrounded, sphericity is low, sorting is very poor, and fine-grained material abounds even among the coarse beds. The coarser sandstone consists of 50 to 70 percent quartz of a notably blue hue, 10 to 20 percent potassium feldspar, 10 to 15 percent albite, as much as 2 percent of partially chloritized biotite, a little muscovite and as much as 3 percent of calcite and clay matrix. The heavy minerals are apatite, sphene, zircon, tourmaline, and rutile.

The chemical composition of the arkose (specimen 272) is shown in table 2, the minor element content in table 3, and a photomicrograph of the same sample in figure 6.

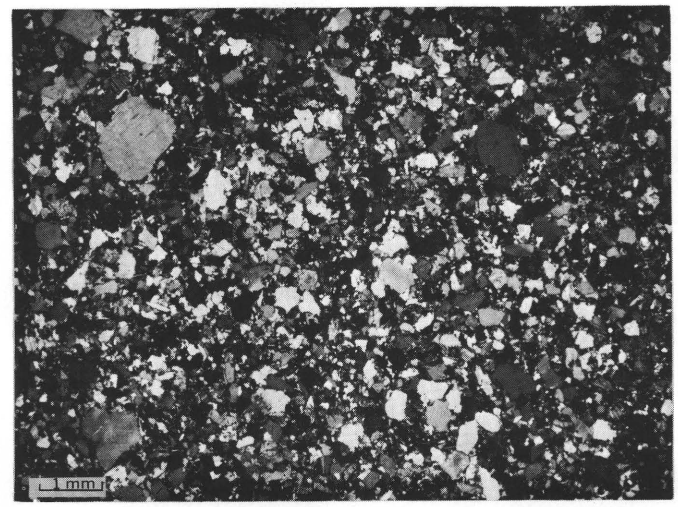

FIGURE 6.—Photomicrograph of silty arkose of the Harmony Formation, northwest spur of Mount Lewis. Note poor sorting and angularity of many grains. Crossed nicols.

Conditions of deposition

The angularity of the clasts, poor mineralogic and size sorting, and the presence of clastic grains of only slightly chloritized biotite all testify to rapid erosion, short transport, and considerable relief of the source area. The source rocks were silicic plutonic rocks, perhaps an albitized biotite-quartz monzonite the principal one. It is significant that the formation is allochthonous in this area, for this shows that deroofed granitic plutons existed some distance to the west, not far from the locus of deposition of this rock. Thus, not all the formations of the western facies are eugeosynclinal, in the sense of association with ophiolitic volcanics, though they are all siliceous as contrasted with carbonate rocks.

Age

No fossils have been found in the Harmony Formation, either here or in the type locality. The formation was first thought likely to be of Mississippian(?) age (Ferguson, Muller, and Roberts, 1951a). In the Hot Springs area (fig. 1), 40 miles to the northwest of Mount Lewis, P. E. Hotz, Jr., has found in the formation fossils that were determined by A. R. Palmer of the Geological Survey as of Late Cambrian age (Roberts and others, 1958, p. 2827). The highly unusual mineral composition of the formation in all the places mentioned seems an adequate ground for correlation.

Structural significance

Though the Harmony Formation is not widespread in this area, the presence of even small masses here is noteworthy. Its significance lies in the fact that, in the adjoining Antler Peak quadrangle, it occurs in overriding thrust plates of western provenance, unconformably overlain by (and therefore recording thrust movement older than) the Battle Formation. (See pl. 3.) The Havallah Formation in the Antler Peak area is confined to thrust plates which, though also of western derivation, override the autochthonous Battle Formation and Antler Peak Limestone. Here, on the western and northwestern spurs of Mount Lewis all four of these formations—Harmony, Havallah, Antler Peak, and Battle—lie in a jumble of closely associated fault blocks. The association of fault slivers from so many tectonic horizons suggests that we are here dealing with materials along the sole of a fault younger than the Antler orogeny of probable Early Mississippian age (Roberts, 1951). When account is taken of the occurrence of blocks of China Mountain(?) Formation—itself younger than the Permian and Triassic Koipato Formation (Ferguson, Muller, and Roberts, 1951a)—it is clear that the fault is of Mesozoic or younger age. We consider it probably Mesozoic and refer it to the Lewis orogeny of this paper. (See p. 123.) The structural implications of the blocks of Harmony Formation are thus such as to warrant notice, even though but little of the formation is here represented.

ORDOVICIAN SYSTEM

VALMY FORMATION

Name, distribution, and topographic expression

The type locality of the Valmy Formation is in the Antler Peak quadrangle; the name was derived from the railroad station of Valmy about 4 miles to the north (Roberts, 1951).

These rocks are exposed in probably more than two-thirds of the bedrock area of the Mount Lewis quadrangle. They are also present, though in less volume, in the Crescent Valley quadrangle, where they compose most of both the east slope of the Shoshone Range and the northwest slope of the Cortez Mountains.

The formation contains many massive beds of quartzite and chert, which form rugged topographic eminences (see fig. 37); the probably more bulky shale, sandstone, and greenstone units tend generally to form smoother slopes between the prominent ledges of more resistant rock (figs. 18 and 38).

Stratigraphy

The Valmy Formation includes a wide range of rock varieties: quartzite of remarkable purity and wide range of grain size; sandstone of several kinds; chert, ranging in color from dark gray through red and green, shale, siltstone; greenstone that varies from pillow lava to ash; and very minor limestone. Of these, the most abundant are sandstone and quartzite, but greenstone is also abundant.

The structural disruptions of the area are so great that it has been impossible to determine the depositional sequence of the several blocks into which the Valmy Formation has been sliced. Apparently unbroken sequences as much as 5,000 feet thick, and several of 2,000 to 4,000 feet, contain beds that cannot be matched in thickness and succession in other blocks. Owing to the scarcity of fossils, it is uncertain whether the differences in stratigraphy of the several blocks are due primarily to lateral facies changes within a relatively thin stratigraphic section or to their sequential, rather than partly contemporaneous, deposition. On the interpretation most conservative as to thickness—one that favors abrupt facies changes to account for the differing stratigraphic successions—the thickness of the formation must exceed 12,000 feet. On the much more likely interpretation of only moderate local facies changes, the thickness is estimated at 20,000 to 25,000 feet. As neither top nor bottom of—nor assured transitions between—the blocks of differing ages are identifiable, the original thickness may have been considerably greater. The facies changes recognizable within individual blocks are not conspicuous, and indeed many apparent variations along the strike are suspect because of the possibility that bedding faults and low-angle slicing, rather than depositional lensing, are responsible for them.

The map shows the Valmy Formation subdivided into lithologic members: quartzite, greenstone, chert, and "undifferentiated" (on pl. 1, indicated respectively as: Ovq, Ovg, Ovc, and Ov). As noted in the map explanation, these designations have no age significance; they are purely lithologic members whose distinction depends much more on mappability than on the thickness of a particular lithologic unit. For example, many thin, but definitely separable quartzites are shown in some areas whereas much thicker ones that constitute large parts of other sections are not distinguished because their boundaries are indefinite owing to poor exposures. The mapped subdivisions of the Valmy are thus useful only in showing structural trends; they do not inform the reader definitely as to the dominant local bedrock and cannot be placed in a stratigraphic succession valid throughout the area.

In general, these sections are described in rough age sequence, beginning with the oldest as determined by Mr. Ross' study of the shelly faunules and the study of the graptolite collections by Messrs. Ross and Berry.

Most of the fossil control is furnished by the graptolites. Collections have been too few and too irregularly scattered to permit confident piecing together of a section. Messrs. Ross and Berry felt that most of the collections could only be placed within fairly wide zonal limits of the British section (Elles and Wood, 1914): Arenig (Zones 3–5 incl.), Llandeilo or Llanvirn (Zones 6-8 incl.), Caradoc (Zones 9–13 incl.), and Ashgill (Zones 14-15).

Our fossil collections are in general too scanty and poorly preserved to allow the several sections assigned to each of these divisions to be arranged in order of age within them, and there may indeed be some gaps in the succession, owing either to incomplete collection, to faulting, or to disconformity. If there are such, they have gone unrecognized in this area of complex structure. No fossils definitely referable to the lowest Ordovician stage (Lower Tremadocian) have been recognized, though some of the trilobites seem more likely of Late Tremadocian age than of Arenigian, which is the lowest stage now recognized from the graptolites as definitely present in the Valmy (R. J. Ross, Jr., 1958).

In the discussion that follows, in which the stratigraphy of the formation is treated in as consistent a chronologic sequence as is warranted by our data, any intention of subdividing the Valmy into formal members is specifically and explicitly disclaimed.

Older parts (*Late Tremadocian-Arenigian, Zones 2–5 incl., of Elles and Wood, 1914*).—The oldest rocks referred to the Valmy Formation in this area crop out on the west side of Crum Canyon near and south of the mouth of Bateman Canyon. Although the section here appears homoclinal and contains no obvious faults, structure and paleontology demand at least two faults of large displacement. The rocks of probable Late Tremadocian age (fossil collns. F–1, F–4, pl. 4) form two blocks separated by a thin sliver of yellow-weathering sandstone belonging to the Elder Sandstone, from which graptolites of Silurian age (F–2, table 4) have been collected.

The more easterly, structurally lower block is the thinner. It rests on a fault slice consisting of chert and quartzite referred to the Valmy (pl. 1), but whose position in that formation is quite uncertain. The rocks that appear to comprise an unbroken section in the south part of sec. 17, T. 30 N., R. 46 E. include, at the base, platy, thin-bedded silty sandstone about 100 feet thick, overlain by interbedded quartzite and sandstone about 250 feet thick, by thin-bedded sandstone interlayered with chert (100 ft) and finally by a greenstone breccia about 400 feet thick from which small limestone lenses have yielded the faunule of F–2. The aggregate thickness of this section thus, is roughly 800 feet.

Although in sec. 17 these strata are immediately overlain by other Valmy beds of presumed late Tremadocian age (see pl. 1), a major fault must intervene, because half a mile farther south along the strike a sliver of Elder Sandstone from which collection F–2 was made separates the two blocks of Valmy strata.

The more westerly, tectonically higher sheet of Tremadocian strata rests on the lower sheet or on the sliver of Elder Sandstone in sec. 21, T. 30 N., R. 46 E. It consists of chert, and interbedded sandstone, siltstone, and fine chert conglomerate about 1,000 feet thick, overlain by greenstone breccia (300 ft) and this in turn by massive sandstone (250 ft) weathering yellow brown, massive quartzite (30 ft), interbedded sandstone, micaceous siltstone, and thin platy limestone (200 ft), massive quartzite (75 ft) and well-bedded sandstone (about 900 ft thick) containing a few beds of chert and quartzite. The aggregate thickness of this block is thus about 2,500 feet.

The greenstone breccia passes along strike, both to south and north, into pillow lavas, in the interstices of which are pockets of brown-weathering limestone. These are the source of collections F–1 and F–155, which have been considered of Late Tremadocian age (Ross, R. J., Jr., 1958; and p. 30 this report).

Although it is possible that the greenstone members of these two partial sections are the same, the differences between their other strata are such as to make it likely that the two sections are additive rather than equivalent.

Other sections that yielded fossils considered characteristic of the earlier Ordovician (Zones 3–5 of the standard British column) are as follows:

1. In the center S½ sec. 35, T. 30 N., R. 45 E., west of upper Lewis Canyon.
2. In the valley of the unnamed stream that drains the southeastern part of T. 28 N., R. 46 E., about half way from Gold Acres to the Utah mine camp.
3. Immediately east of Feris Creek, one-fourth mile above its junction with Indian Creek.
4. On the south side of Indian Creek near the Tomcat mine.
5. On the crest of the hill north of Corral Canyon, across from Bald Mountain.

At localities 1 to 3 faults are so closely spaced that only a few hundred feet of strata—no part of which diagnostically differs from similar strata in the thicker sections just enumerated—is dated by the fossils. Localities 3 and 4 are probably in the same tectonic unit, a section of sandstone and quartzite probably 1,000 feet thick. The section north of Corral Canyon appears to contain at least 2,000 feet of sandstone and quartzite, more than half of it quartzite. The quartzite appears to be thicker than that in any other of the sections of Valmy Formation characterized by fossils of Zone 5 (Late Arenigian) or greater age; however, it is not so much thicker as to preclude approximate equivalence to some others, especially in view of the fact that the other sections are in different thrust sheets.

In summary, then, the part of the Valmy Formation of Zones 2 to 5 (Late Tremadocian to Late Arenigian age) probably is at least 3,500 feet thick and perhaps very much thicker.

Intermediate parts (*Llanvirnian-Llandeilan, Zones 6–8 incl., of Elles and Wood, 1914*).—Only one collection, F–119, from near the SE cor. sec. 35, T. 29 N., R. 46 E., just north of the junction of Feris and Indian Creeks, has been referred by Messrs. Ross and Berry to the Llanvirnian and Llandeilan Stages of the Ordovician. Although we have failed to recognize any major structural break between the rocks here and those across the creek to the south, which furnished collection F–117 of Arenigian age, the attitudes of the rocks are so variable that we are uncertain as to how much of the section here is of one stage and how much of the other. A similar difficulty exists as to a boundary between these rocks and those of Caradocian age represented by collections F–97 and F–126, 2 miles to the north. The rocks of this general area, surely much more than 2,000 feet thick, are interbedded sandstone, chert, quartzite, and very subordinate shale.

Higher part (*Caradocian, Zones 9–13 incl., of Elles and Wood, 1914*).—More collections have been referred by Messrs. Ross and Berry to the Caradocian Stage than to all older and younger stages together. This fact does not, of course, imply that the bulk of the Valmy is of this age; it may only mean that rocks of this age are more abundantly fossiliferous, or that they split better along bedding than those of other parts of the column. The numerous thrust slices generally differ from adjacent ones in detail, though having a general resemblance, but their depositional sequence is no longer recognizable.

On both sides of Indian Creek, between Chicken Creek and the mouth of Feris Creek, we have made several fossil collections referred by Messrs. Ross and Berry to British Zones 9 to 13. These are from thin shale and platy sandstone intercalated in a section of dominant sandstone and quartzite. The rocks apparently belong to the same thrust sheet as the sandstones and quartzites south of Indian Creek and east of the mouth of Feris Creek—mentioned above as containing somewhat older fossils (Zones 3–5)—and those just mentioned as belonging to Zones 6 to 8. The apparent stratigraphic continuity implied by the map throughout this extensive age range is probably illusory, for none of the greenstone and chert and little of the shale elsewhere represented in strata of corresponding graptolite zones are to be found along this section of Indian Creek. Furthermore, although the structure is obscure, it seems unlikely that more than perhaps 2,000 feet of beds from which graptolites of the intermediate zones were locally obtained is represented here. This thickness is exceeded elsewhere by rocks of Zones 6 to 8.

Perhaps the thickest single slice of Zones 9 to 13 is that along the divide between Horse and Pipe Canyons. Here chert and cherty shale about 100 feet thick has yielded a small collection of graptolites (F–83) referred by Messrs. Ross and Berry to Zones 9 to 13 of the British section. This chert is overlain by greenstone (pillow lavas) about 500 feet thick, followed by more chert (500 ft) and interbedded sandstone, shale, and thin chert beds perhaps 600 feet thick, and interlensing greenstone and sandstone as much as 1,000 feet thick. The entire section seems to be continuous and conformable. The total thickness is uncertain because of structural complexities but must approach 3,000 feet. Obviously the fossils do not prove all these rocks to be of Zones 9 to 13—some of the strata may be younger. The section, however, seems considerably different from any other in the quadrangles and thus probably not a time equivalent of the other partial sections that have yielded Caradocian fossils.

A mass of greenstone on and south of Mill Creek may represent the same section as that north of Horse Canyon, just mentioned, but no fossils have been found in adjoining beds. Its thickness must be at least 2,000 feet, perhaps 2,500 feet. Although some pillow lavas are present, this mass is largely fragmental and presumably is composed of submarine tuffs. Although the rocks immediately north of Mill Creek and west of Shwin Canyon are separated from the greenstone by a thrust, they also contain pyroclastic debris interbedded with dark shale and brown-weathering sandstones. This shale yielded collection F–45, referred by Messrs. Ross and Berry to Zones 9 to 13 of the British section. The greenstone mass on and south of Mill Creek may be somewhat older than that on the Horse Canyon-Pipe Canyon divide (R. J. Ross, Jr., oral commun., 1960). If so, the aggregate thickness of greenstone in the Caradocian part of the Valmy may be near 4,000 feet—it is surely more than 2,000.

Other strata that have yielded fossils of roughly equivalent age are:

1. The chert about 1,000 feet thick near the mouth of Lewis Canyon.
2. The chert and associated dark shale about 500 feet thick paralleling Lewis Canyon on the west.
3. The shale and sandstone low on the mountain front between Horse and Crippen Canyons.
4. The sandstone, shale, chert, and quartzite on the divide between Feris Creek and Middle Fork of Mill Creek.
5. The sandstone and shale forming the divide between Middle and South Forks of Mill Creek.
6. The quartzite capping the divide between the South Fork of Mill Creek and Cooks Creek.
7. The sandstones and subordinate shale in the center of sec. 26 (unsurveyed), T. 28 N., R. 45 E., and extending southeastward across Elder Creek.
8. The shale and sandstone klippe in the SW¼ sec. 23 (unsurveyed), T. 28 N., R. 46 E.
9. The sandstone and shale that override the Devonian rocks just south and west of the abandoned site of Lander.
10. The thick chert, sandstone, and shale to the southwest of Triplet Gulch, on the east side of the Shoshone Range south of Goldquartz.
11. The sandstone and shale, several thousand feet thick, that make up the scarp of the Cortez Mountains in the southeast corner of the Crescent Valley quadrangle.

In summary, the Caradocian part of the Valmy consists of sandstone, greenstone, quartzite, shale, and chert, that certainly aggregate at least 2,700 feet in thickness; if, as seems very probable, the sandstone and quartzite of Indian Creek are not lateral equivalents of these rocks or of the cherts of Triplet Gulch and of the sandstone and shale of the Cortez Mountains, a thickness of 6,000 to 9,000 feet is reasonable.

The highest beds (Ashgillian, Zones 14–15 of Elles and Wood, 1914).—No collections have been definitely referred to the Ashgillian Stage of the Ordovician by Messrs. Ross and Berry, but Mr. Ross (oral commun., 1960) thinks it likely that one, F–140, may indeed be of this age. This collection, from the hilltop three-quarters of a mile south-southwest of the site of Lander on Indian Creek, is from shale interbedded with sandstone and associated with a thin lens of greenstone. Perhaps 1,000 feet of beds is represented in this block. Mr. Hass considered some conodonts from a nearby locality to be of Late Ordovician age (p. 32).

Many other collections may represent beds as young as or younger than these but none are diagnostic, as, indeed, these also are not.

In summary, despite the uncertainties of sequence and correlation within the Valmy Formation, the formation must consist of at least 12,000 feet, and quite possibly two or more times this thickness, of sandstone, quartzite, chert, greenstone, graywacke, and shale, as well as a few lenses of unfossiliferous limestone as much as 10 feet thick and a few hundred feet long. The members formally recognized by Roberts (oral commun., 1954) across the Reese River in the Antler Peak quadrangle have not been identified here, perhaps because of facies changes but more likely because the members readily separable in the small areas of Valmy in the Antler Peak quadrangle form only minor parts of the much thicker sections exposed in the Shoshone Range.

Petrography

The Valmy Formation includes a wide range of rock varieties: quartzite, sandstone of several kinds, chert, shale, siltstone, greenstone, and very minor limestone. The most abundant are sandstone and quartzite.

The quartz sandstones and quartzites are remarkably pure (specimens 1, 4, 245; and fig. 7). They consist of moderately to poorly sorted quartz grains that range from coarse-sand size to silt size, and whose distribution is somewhat skewed toward the silt size. The coarser grains are highly spherical and stand out conspicuously from the finer grains, which are angular to subrounded. Small authigenic outgrowths, in optical continuity with the central grain, cement many of the rocks so that they break across the grains, though the large spherical grains commonly do not break and stand out on the fractured surface. Sparse grains of chalcedony (some recrystallized into radially aggregated quartz needles) and the heavy minerals zircon, tourmaline (both green and brown), biotite, hornblende, and augite are present, along with presumably authigenic barite and pyrite.

Although the common cement is silica, some beds are cemented by dolomite (11 percent dolomite, 3 percent clay in specimen 19, table 2) and a few by a ferrian member of the variscite group—$(Al, Fe^{+3})(PO_4)\cdot 2H_2O$—according to an X-ray determination by A. J. Gude 3d. This ferrian variscite is probably related to the common turquoise, which is widespread in the region, usually in veins but also as cement.

A second variety of sandstone is perhaps even more abundant than the quartz-rich one. This is a cherty variety that ranges in grain size from fine gravel to silt. Angular to subrounded gravel-sized fragments of chert and siliceous shale comprise 30 to 60 percent of the rock (fig. 8; specimen 225, table 2). Highly rounded and spherical grains of quartz of sand size make up 15 to 40 percent; the remainder is angular silt-sized quartz with about 3 percent sericite.

These rocks consist of material from two sources—the well-rounded grains clearly from a different provenance than the chert.

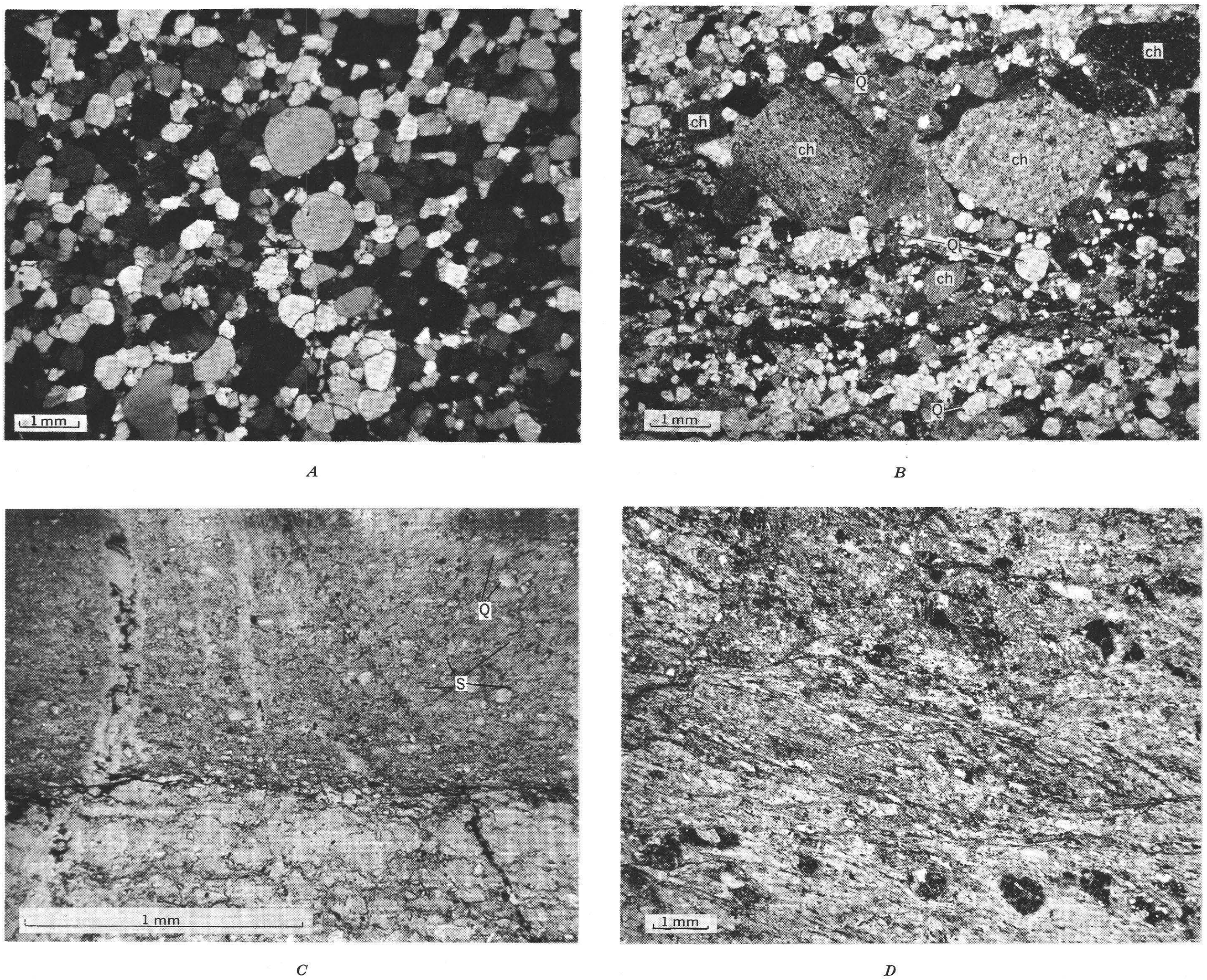

FIGURE 7.—Photomicrographs of rocks of the Valmy Formation. *A*, pure quartzite, showing highly spherical coarse grains in a matrix of finer and more angular grains cemented by mosaic-textured quartz. Also present are a very few detrital grains of chalcedony and sparse heavy minerals, including zircon, green tourmaline, brown tourmaline, biotite, epidote, hornblende, and augite. Crossed nicols. *B*, cherty sandstone—poorly sorted silty to very fine gravelly sandstone containing angular fragments of chert (ch), grains of well-rounded quartz (Q), and angular silt-sized particles of quartz, chert, and shale. Plain light. *C*, chert—a partly recrystallized very fine grained siltstone that contains sporadic larger well-rounded grains of quartz (Q). A few flakes of sericite (S) lie parallel to bedding. This is specimen 15 of table 2. Plain light. *D*, greenstone, showing lenticles of rock fragments, quartz, chalcedony, and calcite, in a schistose groundmass of chlorite, quartz, ilmenite, leucoxene, rutile, and apatite. Most of the rock fragments are metaquartzite, with granoblastic texture, some are quartz-biotite-calcite rock, and a few are microporphyritic volcanic rocks containing randomly oriented laths, formerly of plagioclase but fully replaced by quartz. See specimen 488 of table 2. The original rock was probably a soda-poor basic volcanic, containing accidental inclusions. Plain light.

Most cherts of the Valmy Formation (fig. 7*C*) are composed chiefly of very fine silt-sized quartz in a siliceous matrix. According to A. J. Gude 3d, the X-ray analyses indicates that only quartz is present in important amounts; if amorphous silica is present, it amounts to less than 5 percent. There is a little sericite, organic matter, and iron oxide. In both mineral content and chemistry, the chert is closely similar to the quartzite of the Valmy. (See table 2.) Other cherts of the formation contain considerably more sericite and organic material as well as tests of radiolaria. The radiolarian tests are spherical aggregates of quartz from 0.01 to 0.02 mm in diameter, some of which preserve punctate shell patterns and spines.

The greenstones of the Valmy range from pillow lava through pyroclastic breccia to fine ash; in a few places the same stratum may be seen to change along strike from one facies to the other within a few hundred yards. The pillow lavas are generally albitized but the cores of some pillows still contain andesine, though the feldspar in the exterior has altered to albite. The dark minerals are generally altered to chlorite. Some rocks that look like ashes have been so thoroughly altered as to retain no feldspar at all, though the former presence of plagioclase is indicated by lathlike pseudomorphs of quartz (fig. 7*D*).

Commonly the sparse limestones of the Valmy are closely associated with the greenstones. They consist of fragments of shells of trilobites, brachiopods, and gastropods, with some admixed volcanic fragments and the usual minerals of the sandstone and shale units of the formation.

Conditions of deposition

The eugeosynclinal association of greenstone, chert-rich sandstone, chert, siltstone, and subordinate limestone seems somewhat anomalous in association with the remarkably pure quartz sandstones and dolomitic sandstones of the Valmy. These quartz sandstones are quite as "mature," in the sense of being free from feldspar and other readily altered minerals, as any in what is normally thought of as a shelf facies. Minor coquina of shells of trilobites and brachiopods associated with the pillow lavas of Crum Canyon suggests that the depth of water was not more than 100 to 200 feet (Lochman, 1949). The rocks above and below are indistinguishable from those of the bulk of the formation, and show much current bedding. (See fig. 8.) Perhaps the scarcity of fossils other than graptolites throughout the formation does not signify any great depth, but merely that bottom conditions were too inhospitable for benthonic organisms except where hard bottom, such as was furnished by the pillow lavas, was available. The graptolites, being pelagic organisms, could survive in water of any depth; their preservation does not necessarily imply any abyssal depth for the deposits in which they are found.

Age

The fossil collections from the Valmy Formation are few and most are of rather scrappy material. The

TABLE 4.—*Graptolithina from the Mount*

[Identified by R. J. Ross, Jr., and W. B. N.

Species	USGS fossil colln. No. / Field colln. No.	D115-CO F-87	D114-CO F-117	D123-CO F-122	D95-CO F-138	D96-CO F-139	D113-CO F-119	D112-CO F-82	D116-CO F-86	D118-CO F-123	D121-CO F-121	D135-CO F-45	D103-CO F-126	D110-CO F-107	D119-CO F-92	D120-CO F-83	D124-CO F-81	D127-CO F-96	D102-CO F-125
	System	Ordovician																	
	Series	Arenig					Llandeilo or Llanvirn	Caradoc											
Dichograptus cf. *D. separatus* Elles						×													
sp		×																	
Tetragraptus fruticosus (Hall) (3-branched var.)				×															
quadribrachiatus (J. Hall)						×													
sp. (dependent type)						×													
Phyllograptus anna J. Hall						×													
cf. *P. anna* J. Hall					×														
ilicifolius J. Hall						×													
Didymograptus cf. *D. serratulus*																			
cf. *D. serratulus* (Hall)									×										
Didymograptus? sp		×	×	×						×	×								
?*Didymograptus* sp							×												
Cryptograptus tricornis (Carruthers)									×	×						×			
Glossograptus hincksii Hopkinson							×												
cf. *G. hincksii* (Hopkinson)																×			
cf. *G. hincksii* Hopkinson											×								
Glossograptus? aff. *G. hincksii* Hopkinson																			
hincksii var. *fimbriatus* Hopkinson									×										
cf. *G. hincksii* var. *fimbriatus* (Hopkinson)																			
aff. *G. hystrix* Ruedemann										×									
Nemagraptus sp									×										
exilis var. *linearis* Ruedemann										×									
Nemagraptus? sp												×							
Dicellograptus divaricatus var. *salopiensis* Elles and Wood									×										
aff. *D. divaricatus* var. *salopiensis* Elles and Wood										×									
gurleyi Lapworth?													×						
intortus Lapworth												×	×						
sextans (J. Hall)								×											×
sextans (J. Hall)?									×										
sextans var. *exilis*								×	×			×							
cf. *D. sextans* var. *exilis* Elles and Wood										×									
Dicellograptus sp								×				×							
Dicellograptus? sp																		×	
?*Dicellograptus* sp																			
Dicellograptus? sp. or *Dicranograptus*? sp. (Group IV of Elles and Wood)										×									
Dicranograptus contortus Ruedemann?															×				
Dicranograptus? sp																			
Dicranograptus? sp. or *Dicellograptus*? sp											×								
Climacograptus cf. *C. brevis* Elles and Wood														×					
eximius Ruedemann																			
exmius? Ruedemann																			
hastatus n. var																			
riddellensis Harris							×												
scharenbergi Lapworth																			
of type of *C. scharenbergi* Lapworth																			

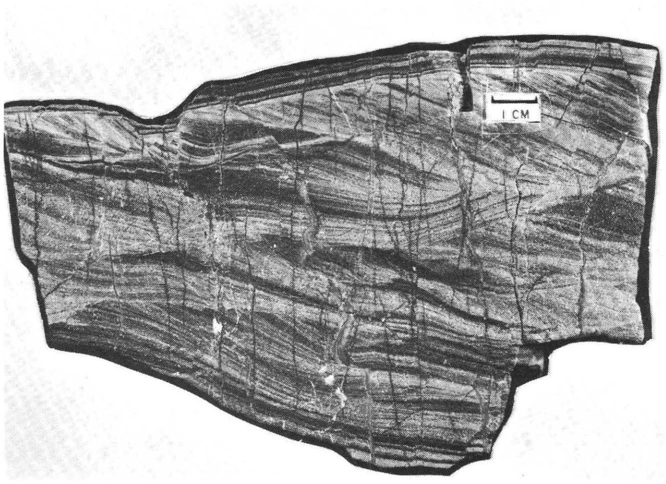

FIGURE 8.—Current-bedded clastic limestone of the Valmy Formation, sawed and etched to bring out the bedding structures. The specimen came from close above a greenstone bed on the west wall of Crum Canyon, opposite the mouth of Bateman Canyon.

most abundant fossils are graptolites. A few of the collections were studied by the late Josiah Bridge. Most of these and all later collections were studied by R. J. Ross, Jr., of the U.S. Geological Survey. He and W. B. N. Berry have again reviewed all the collections; their identifications and resulting age assignments are tabulated in table 4.

The Ordovician collections are included in a comprehensive study by Ross and Berry (1963) of graptolite collections made in the Basin Ranges through 1958 by Survey personnel. The species from Silurian collections are also summarized in table 4. It should be noted that F-67, F-79, F-93, and F-102 are from the Silurian Roberts Mountains Limestone or Elder Sandstone rather than from the Valmy Formation.

R. J. Ross, Jr. (written commun., 1960) stated:

The Ordovician graptolite collections range in age from Arenig through Early Caradoc, the same range covered by graptolites

Lewis and Crescent Valley quadrangles, Nevada

Berry. No fossils positively identified as Llanvirn]

Ordovician—Continued													Ordovician or Silurian			Silurian							Silurian undivided		
Caradoc—Continued																Llandovery		Tarranon	Ludlow						
D117-CO F-97	D130-CO F-75	D537-CO F-162	D104-CO F-127	D106-CO F-134	D111-CO F-113	D129-CO F-74	D131-CO F-104	D538-CO F-171	D128-CO F-80	D93-CO F-135	D98-CO F-143	D97-CO F-140	D107-CO F-131	D122-CO F-73	D125-CO F-110	D17-SD F-93	D19-SD F-102	D18-SD F-67	1274-CO F-2	D1277-CO F-5	1278-CO F-6	1279-CO F-7	D126-CO F-99	D109-CO F-116	D20-SD F-79
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TABLE 4—*Graptolithina from the Mount Lewis and*

Species	System	Ordovician																	
	Series	Arenig					Llandeilo or Llanvirn	Caradoc											
	USGS fossil colln. No.	D115-CO	D114-CO	D123-CO	D95-CO	D96-CO	D113-CO	D112-CO	D116-CO	D118-CO	D121-CO	D135-CO	D103-CO	D110-CO	D119-CO	D120-CO	D124-CO	D127-CO	D102-CO
	Field colln. No.	F-87	F-117	F-122	F-138	F-139	F-119	F-82	F-86	F-123	F-121	F-45	F-126	F-107	F-92	F-83	F-81	F-96	F-125
Climacograptid similar to *C. scharenbergi* Lapworth																			
Climacograptus cf. *C. scharenbergi* Lapworth									×										
aff. *C. scharenbergi* Lapworth																×			×
n. sp. (similar to *C. caudatus*)																			
2 n. spp																			
sp									×			×				×	×	×	×
sp. (possibly a minute species of type of *C. scharenbergi*											×								
Climacograptus? sp											×								
Diplograptus multidens var.?																			
sp							×												
Diplograptus? sp																			
Diplograptus? sp. or *Glyptograptus*? sp																			
Amplexograptus aff. *A. arctus* Elles & Wood							×												
Amplexograptus? sp							×												
Glyptograptus cf. *G. euglyphus* Lapworth							×												
cf. *G. euglyphus* (Lapworth)																			
euglyphus var. *pygmaeus* Ruedemann																			×
euglyphus var. *pygmaeus* Ruedemann?								×											
cf. *G. euglyphus* var. *pygmaeus* Ruedemann																			
teretiusculus (Hisinger)									×										
teretiusculus (Hisinger)?										×									
cf. *G. teretiusculus* Hisinger							×												
cf. *G. teretiusculus* (Hisinger)												×				×			
aff. *G. teretiusculus* (Hisinger)																			×
sp																			
Glyptograptus? sp										×						×		×	×
?*Glyptograptus* sp																	×		
Orthograptus calcaratus var.?																			
cf. *O. calcaratus* Lapworth																			×
Orthograptus? of the *O. calcaratus* type																			
Orthograptus calcaratus cf. var. *acutus* Lapworth																			
of the type of *O. quadrimucronatus*																			
whitfieldi (J. Hall)																			×
whitfieldi Hall?																			
Orthograptus? sp.																			
Trigonograptus ensiformis J. Hall																			×
cf. *T. ensiformis* (J. Hall)							×												
ensiformis var. *obtusus* Ruedemann																			×
Retiograptus geinitzianus (Hall)									×	×									
Monograptus cf. *M. pandus* Lapworth																			
aff. *M. convolutus* Hisinger																			
cf. *M. regularis* Törnquist																			
tumescens? var.?																			
sp																			
2 spp																			
Monograptus? sp.—2 spp																			
Caryocaris? sp		×																	

from the Valmy formation of the Antler Peak quadrangle. In both of these areas most of the collections are of Early Caradoc age. A very few have furnished Arenig and Llandeilo forms. This seems to be a common pattern in Nevada, although a few collections from other quadrangles suggest that Tremadoc and undoubted Ashgill species may eventually be found in the Mount Lewis and Crescent Valley quadrangles as well. Although F–140 is not quite well-enough preserved for certainty, I suspect that it may represent the Ashgill, though Dr. Berry does not agree. It is probable that the trilobites from Collection F–1 (USGS Colln. 1273–CO) and F–176 (USGS Colln. D548–CO) are Late Tremadoc or Early Arenig in age and therefore older than any of the listed graptolite collections.

On the basis of the graptolites and the three trilobite collections, the Valmy formation correlates with the Pogonip group, the Copenhagen formation of Merriam, and probably with the Eureka quartzite. No correlatives of the Hanson Creek formation are recognized with confidence.

Other fossils, though much fewer than the sparse graptolites, have furnished additional information. The following collection, from a limestone lens in greenstone, was made by James Gilluly and M. R. Mudge; the trilobites it contains were described by R. J. Ross, Jr. (1958).

F–1 (USGS colln. 1273–CO)—W. side Crum Canyon, alt 6, 400 ft, on spur 600 ft N. of center of S. section line, sec. 20, T. 30 N., R. 46 E.

Trinodus valmyensis Ross
Trinodus? sp.
Leiostegium mudgei Ross
sp.
Apatokephalus gillulyi Ross
Shumardia aff. *S. minutula* Harrington
Strigigenalis sp.
Four undet. gen. and spp.
Pliomeroides sp.

Josiah Bridge identified two gastropods: *Proplina* sp., and a probable new genus related to *Ophileta* and *Liospira*. Messrs. Ross and Bridge both considered this fauna to be probably of Late Tremadocian or Early

Crescent Valley quadrangles, Nevada—Continued

Ordovician—Continued													Ordovician or Silurian			Silurian							Silurian undivided		
Caradoc—Continued																Llandovery		Tar-ra-non	Ludlow						
D117-CO F-97	D130-CO F-75	D537-CO F-162	D104-CO F-127	D106-CO F-134	D111-CO F-113	D129-CO F-74	D131-CO F-104	D538-CO F-171	D128-CO F-80	D93-CO F-135	D98-CO F-143	D97-CO F-140	D107-CO F-131	D122-CO F-73	D125-CO F-110	D17-SD F-93	D19-SD F-102	D18-SD F-67	1274-CO F-2	D1277-CO F-5	1278-CO F-6	1279-CO F-7	D126-CO F-99	D109-CO F-116	D20-SD F-79
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Arenigian age, equivalent to Zone 2 or 3 of Elles and Wood (1914, p. 514–526). The presence of *Leiostegium*, *Apatokephalus*, and *Shumardia* indicates correlation with the Goodwin Limestone of the Eureka section (R. J. Ross, Jr., 1958, p. 559).

In F–4, a collection by M. R. Mudge from limestone associated with greenstone at an altitude of 6,020 feet, 1,100 feet north and 2,850 feet west of the SE cor., sec. 17, T. 30 N., R. 46 E., Josiah Bridge identified linguloid brachiopods and one small ribbed brachiopod, which he considered to be of probable Early Ordovician age.

In F–13, collected by James Gilluly from float on shale slope east of Crum Canyon at an altitude of 5,840 feet, 900 feet north and 950 feet east of the SW cor., sec. 16, T. 30 N., R. 46 E., G. A. Cooper identified a *Paterula* sp., which he considered of probable late Early Ordovician age, possibly Middle Ordovician.

In F–176 (USGS colln. D548–CO) collected by James Gilluly and Harold Masursky from a point 950 feet north, 3,150 feet west of SE cor. sec. 17, T. 30 N., R. 46 E., R. J. Ross, Jr., identified:

Brachiopods:
- *Apheoorthis*? sp.
- Undet. gen. and sp. (orthid)

Trilobites:
- *Leiostegium* sp.
- Undet. gen. and sp. No. 4 (Ross, 1958)
- Undet. gen. and sp. No. 2 (Ross, 1958)
- Hypostome, like Ross (1958, pl. 84, fig. 21)
- *Pliomeroides*? sp.
- Undet. gen. and sp.

Ostracode undet.

This fauna is partly the same as that from collection F–1 (Ross, R. J., Jr., 1958, p. 559). Therefore it should be Early Ordovician in age (R. J. Ross, Jr., written commun., 1960).

A few conodonts have also been found in the Valmy Formation during this survey. These were referred to W. H. Hass, who reported as follows:

F-163 from point 2,500 ft E., 400 ft S. of the NW cor. sec. 5, T. 28 N., R. 47 E., just south of Indian Creek.

Belodus? *mutatus* Branson and Mehl
Phragmodus insculptus Branson and Mehl
Balognathus sp.

The first two species were described from the Upper Ordovician Thebes sandstone of Missouri and the genus *Balognathus* was first reported in some Upper Ordovician material from England.

F-166 from point at alt 6,240 ft, about 1,200 ft up ridge from hill 6202, sec. 14 (unsurveyed), T. 28 N., R. 44 E., south of Harry Creek.

The collection consists of one conodont, which herein is identified as *Phragmodus*? sp. This identification is by no means certain, but if correct, it would suggest Middle or Upper Ordovician.

F-167 from point at alt 6,960 ft on spur of hill, at boundary of secs. 13 and 24 (unsurveyed), T. 28 N., R. 44 E.

This collection consists of one specimen each of:

Belodus ornatus Branson and Mehl
Oistodus sp.
Conodont fragment

Belodus ornatus is a very distinctive species which, to the best of my knowledge has been reported only from the Upper Ordovician.

Many collections were too fragmental for any very useful age assignments but *Hystrichosphaeridium* sp., smooth ostracodes, hexactinellid sponge spicules; and fragments of polygonal corallites (possibly *Favosites*), of both calcitic and phosphatic brachiopod shells, of crinoid columnals, bryozoa, and conodonts, have been observed in several collections by Jean Berdan.

Correlation

The probable correlatives of the Valmy Formation in more distant localities have been enumerated in the discussion by Mr. Ross (p. 30). It seems appropriate here to mention some strata of nearby areas, whose depositional relations bear more directly on the regional paleogeography.

The most immediate correlatives are of course the rocks of the type locality of the Valmy Formation in the Antler Peak quadrangle immediately to the northwest (Roberts, 1951; see fig. 1, this report). As defined, the Valmy in its type area includes strata of Early and Middle Ordovician ages only. The extension of the name to rocks of the northern Shoshone Range thus constitutes an expansion of the time span of the formation. There is, however, no question of the lithologic similarity of the formation in the Shoshone Range to that in the type locality, nor is there any clear difference between parts that are respectively of the same age as the type strata and of younger age. The differences that allow the formation in its type locality to be divided into members seem not to be identifiable in the more extensive exposures of this area. As with many variable formations, differences that locally appear conspicuous cannot be consistently recognized over larger areas of complex structure and stratigraphic diversity.

In the other direction, toward the southeast, the nearest correlative is the Vinini Formation (pl. 3), whose type locality, Vinini Creek, lies on the northeast flank of the Roberts Mountains, about 30 miles to the southeast of this area (Merriam and Anderson, 1942, p. 1694). The type Vinini, like the type Valmy, so far as known comprises only rocks of Early and Middle Ordovician age. In the Tuscarora Mountains, about 40 miles north of Vinini Creek, similar rocks pass upward into strata of Late Ordovician age, in a section that is at least 7,000 feet thick (Roberts and others, 1958, p. 2832). In the type locality, the lower (Early Ordovician) part of the Vinini consists of dominant siltstone and shale with considerable quartzite, limestone, limy sandstone and subordinate chert, tuff and lava; the upper (Middle Ordovician) part consists of chert and black organic shale. In general nearly every individual lithologic variety of either Vinini or Valmy can be matched in the other formation (though organic shale has not yet been found in the Valmy). Nevertheless the relative proportions of the several varieties are conspicuously different: far higher proportions of quartzite, volcanics and coarse sandstones are present in the Valmy; relatively much more shale and siltstone in the Vinini. Where the two formations are brought into contact by faults, as in the Cortez Mountains, consistent discriminations between them have been relatively easy to make (Gilluly and Masursky, 1965). They are sufficiently similar to leave little doubt of their close depositional association but different enough to suggest that the Vinini was deposited farther from the main source of both detrital and volcanic materials.

This point is highly significant as it implies a western, northwestern, or northern source, not only for the volcanics, chert, and graywacke materials of the Valmy and Vinini, but also for the pure quartzites. The time span of the Valmy includes that of the Eureka Quartzite and of the Kinnikinic Quartzite of the Bayhorse quadrangle, central Idaho (C. P. Ross, 1934, p. 950–952; 1937, p. 18; see fig. 1 and pl. 3, this report). The great thickness of pure quartzite in both Valmy and Kinnikinic and the thinning of these beds eastward and southward make very doubtful an eastern or southern source of the sands composing the Vinini.

Kirk (1933, p. 36, 40) suggested a southern or eastern source for the Eureka Quartzite, along with minor contributions from the west. Webb's reference to the Eureka (1958, p. 2368–2377) as including beds regressive and transgressive with respect to the land to the southeast implies a like source. Nolan, Merriam, and Williams (1956, p. 31–32), however, showed that miscorrelations had led Kirk to believe that the Eureka thickened eastward and southward. They suggested that perhaps the Eureka was deposited across a positive area that originally had separated the basins of deposition of the Vinini and Pogonip, and thereby implied a northwestern source for the sands of the Eureka. Perhaps, therefore, the Eureka represents a tongue of one of the quartzites of the Valmy or Kinnikinic. The suggestion that the Eureka may be a thin southward tongue of the much thicker Kinnikinic has been independently made by J. C. Hazzard and W. R. Moran of the Union Oil Co. (oral commun., 1952). Though such a source has not actually been demonstrated, this possibility is nevertheless surely worth considering in future stratigraphic studies of the northern Great Basin.

West of the Galena Range, two dominantly clastic formations of Ordovician age have been recognized in Nevada: the Sonoma Range Formation and the Comus Formation. (See pl. 3.) These undoubtedly correlate, at least in part, with the Valmy. The Sonoma Range Formation was described from the Winnemucca quadrangle (Ferguson, Muller, and Roberts, 1951a). No fossils were found in the formation during the reconnaissance upon which that report was based; however, the formation lithologically is so similar to the Valmy that it almost certainly is a lateral equivalent.

The Comus Formation, on the other hand, is known to be of the same age as part of the Valmy (Ferguson, Roberts, and Muller, 1952; Roberts and others, 1958) but differs notably in lithology. The Comus, as exposed both in its type locality in the Edna Mountains and in the Osgood Mountains, about 50 miles northwest of this area (fig. 1 and pl. 3), is composed of interbedded chert, shale, limestone, dolomite, and siliceous tuff. The proportion of carbonate to siliceous rocks is many times that in the Valmy. The suggestion that the Comus belongs to a facies transitional between the eastern and western, as used in this report, was made in a report by Roberts, Hotz, Gilluly, and Ferguson (1958, p. 2831). Indeed, the clastic rocks, except for the siliceous volcanics, do resemble those of the Vinini. Such an assignment might be taken to imply travel of the Valmy and Vinini for an additional 50 miles from the west, for presumably the transitional facies should have been deposited to the east of the Vinini, which contains but a small proportion of carbonates; such a conclusion seems premature, however, in the present state of our knowledge of paleogeography. Facies boundaries need not have been rectilinear. No siliceous volcanics other than those of the Comus have been recognized among the Lower Paleozoic rocks of the region.

Toward the southwest, the nearest Ordovician rocks that resemble the Valmy Formation crop out in the Mina and Coaldale (Ferguson, Muller, and Cathcart, 1954; 1953), and Silver Peak (Turner, 1902, p. 265–266; 1909, p. 243) quadrangles. (See fig. 1.) These rocks, originally described by Turner as the Palmetto Formation, consist of dark thin-bedded chert, gray slate, and sporadic limestone, with numerous interbeds of light-colored felsite. Their thickness, of which no accurate measurements are known to us, is estimated as at least 4,000 feet in the Mina quadrangle, where no base is exposed. The age ranges from Beekmantown (Early Ordovician) to Middle Ordovician. These siliceous clastic rocks are so near the dominantly carbonate rocks of equivalent age in the Inyo Range (Kirk, 1918, p 32–36) that it seems likely that a thrust zone corresponding to the Roberts thrust may pass between the two (H. G. Ferguson, oral commun., 1952). The reconnaissance work thus far done in this area has not, however, furnished direct evidence on this point.

In the southern Toiyabe and Toquima Ranges, 75 miles to the south of the Mount Lewis area, Ferguson (Ferguson, Muller, and Cathcart, 1953) included under the name of Palmetto Formation all the rocks equivalent to the Mayflower Schist, Zanzibar Limestone, and Toquima Formation as originally described in the Manhattan district (Ferguson, 1924, p. 20–25. See pl. 3, this report). Both the Mayflower and the Toquima resemble the Valmy, or perhaps more closely the Vinini Formation, but neither Valmy nor Vinini contains a limestone like the Zanzibar in either thickness or lithology. The structural complexities now known in the general region suggest the possibility that faulting may have caused intercalation of the Zanzibar Limestone. No fossils have been obtained from that formation at Manhattan. On the other hand, Ferguson's original interpretation, that the intercalation results from normal deposition, is by no means unlikely; somewhere, if our interpretation of major facies distribution is correct, there must have been interfingering of carbonate and siliceous facies of the Ordovician. Perhaps such an interfingering is preserved at Manhattan, either in autochthonous or allochthonous structural blocks. This is the interpretation of Kay (1960) and Lowell (1960) for the Toiyabe and Toquima Ranges.

A similar question arises in the country to the north of the Mount Lewis-Crescent Valley area. Not enough work has yet been published to enable conditions in the intervening area to be evaluated, but comparisons can be made with those of central Idaho (fig. 1). In the Mackay area, Umpleby (1917, p. 24–25) recognized a quartzite 1,600 feet thick overlain by about 950 feet of dolomite, both of Ordovician age. In Lemhi County, to the north, the quartzite had originally been assigned to the Cambrian (Umpleby, 1913, p. 32–33; 1917, p.

25), but a recent study by E. T. Ruppel and R. J. Ross, Jr., (R. J. Ross, Jr., oral commun., 1961) has revealed Ordovician fossils in it. In the Bayhorse quadrangle to the west, a comparable quartzite, the Kinnikinic, overlies the fossiliferous Ramshorn Slate of Early Ordovician age (C. P. Ross, 1934, p. 950). In the Wood River region, to the south near Hailey, the Ordovician is represented by an unnamed slate of Early Ordovician age and by the Phi Kappa Formation, more than 9,000 feet thick, of Early and Middle Ordovician ages. The Phi Kappa, as described, strongly resembles the Valmy, or perhaps more closely the Vinini, in its content of sandstone, shale, and chert (Umpleby, Westgate, and Ross, 1930, p. 18–23).

In the Bayhorse quadrangle, Idaho, C. P. Ross (1934, p. 942–956; 1937, p. 14–22) classified the Ordovician rocks into three formations: the Ramshorn Slate, 2,000 feet, at the base, overlain by the Kinnikinic Quartzite, about 3,500 feet, and the Saturday Mountain Formation, about 3,000 feet thick. The Kinnikinic was doubtfully considered Middle Ordovician and the Saturday Mountain definitely Late Ordovician in age (C. P. Ross, 1934, p. 950–956). In the Borah Peak quadrangle, just east of the Bayhorse, the Kinnikinic Quartzite, there about 3,000 feet thick, overlaps the Belt formations (Precambrian) along a pronounced angular unconformity and is in turn overlain by the Saturday Mountain Formation, there about 500 to 700 feet thick (C. P. Ross, 1947, p. 1102–1105). The Kinnikinic was referred to the Upper Ordovician in Ross' 1947 report, on the evidence of fossils thought to have been collected from it, but more detailed later studies (E. T. Ruppel, oral commun., 1961) make it appear that the fossils came from the basal transitional beds of the Saturday Mountain. In any case, it is highly probable that part, if not all, of the Kinnikinic is of Middle Ordovician age.

The Saturday Mountain Formation is almost certainly the equivalent of the Fish Haven Dolomite of Utah and thus of the Hanson Creek of this area (Edwin Kirk, quoted by C. P. Ross, 1934, p. 955; R. J. Ross, Jr., oral commun., 1955), though it contains considerably more clastic material, perhaps showing facies transition toward the siliceous facies here represented by the Valmy Formation. The Kinnikinic, though many times thicker, seems likely to represent the Eureka Quartzite of this area and thus to be equivalent to part of the Valmy Formation also. The relations between the facies of the central Idaho rocks should throw considerable light on the original facies transitions in the latitude of the Shoshone Range.

Provenance

The coarser sand beds of the Valmy record two sources, one to furnish the well-sorted, well-rounded quartz sandstone; and the other to supply the poorly sorted chert-rich sandstones with their admixture of angular siliceous shale particles.

The first of these sandstones is a very mature sediment, clearly derived from an older quartz sandstone from which all feldspar and other readily weathered minerals had been removed. The stratigraphic relations in central Idaho perhaps give a clue as to the source. In the Borah Peak quadrangle the Kinnikinic Quartzite overlaps two quartzite formations of the Belt Series (C. P. Ross, 1947, p. 1096–1104), either of which, being virtually pure quartzites with only hematitic cement, could have supplied the quartz for the Kinnikinic itself and for its approximate equivalents, the Eureka and the medial part of the Valmy. Indeed, even the heavy mineral suites of the Belt formations—zircon, tourmaline, and apatite—are like that of the pure quartzite of the Valmy. There is no reason why the source formations, if these quartzites of the Belt indeed did supply the highly quartzose sediments, could not have been exposed both earlier and later than the Middle Ordovician. The marked similarity of the quartzites of the Valmy—both those younger, older, and of the same age—to the Eureka, suggests a similar source. The absence of comparable quartzites in rocks of the shelf facies that are either older or younger than the Eureka makes the suggestion that all the pure quartzites were derived from the east (Kay, 1960; Lowell, 1960, and others) seem to us unlikely. The quartzites are both more abundant and thicker in the Valmy than in the Vinini. As the Valmy also contains more volcanics and cherts, it has more of a eugeosynclinal character than the Vinini and was therefore probably deposited farther west; this also suggests a more westerly source for the quartz sand.

The second component, the chert-rich sand, much of it angular and very poorly sorted, obviously had a very different source. Perhaps this material was derived from welts within the eugeosyncline itself in which the Valmy and Vinini were being deposited. One can readily imagine that if a newly deposited sequence of chert and shale were uplifted in a welt lying to the west, the shale might readily disaggregate and be transported eastward to form the shale facies of the Vinini while the interbedded chert might disaggregate to fine gravel and sand-sized materials such as compose the chert sandstones of the Valmy.

The many alterations of these contrasting deposits in the Valmy show that the currents bringing the pure quartz sands were repeatedly interrupted while local turbulent currents deposited the little-traveled cherty sands.

SILURIAN SYSTEM

ELDER SANDSTONE

The Elder Sandstone is here named from exposures in the drainage basin of Elder Creek, in the southeastern part of the Mount Lewis quadrangle. The hill extending northeastward from the sharp bend of Elder Creek in sec. 30, T. 28 N., R. 46 E., to the Utah mine is designated as the type locality.

Distribution and topographic expression

The Elder Sandstone crops out, at its type locality, over an area of about 3 square miles. It also crops out in a band roughly 2,000 feet wide in an isoclinal thrust sheet folded into a synform above the Kattenhorn fault, northeast, east, southeast, south, and southwest of Mount Lewis. There are also small exposures east of the lower fork of Elder Creek, near the Clipper mine; on the divide between Cooks Creek and Elder Creek; in the valley of the western tributary of Cooks Creek in sec. 21, T. 28 N., R. 45 E.; on the south side of Feris Creek between 1 and 2 miles above its mouth; in lower Crum Canyon; and in the low hills west and north of Gold Acres.

The dominant rocks of the Elder are moderately cemented sandstones, intermediate in resistance to erosion between the quartzite and chert of the Valmy and the sandstone and shale of that formation. It thus forms ridgetops in places, but is generally not topographically conspicuous.

Stratigraphy

Both lower and upper contacts of the Elder Sandstone are mechanical wherever exposed, so that nothing is known of the formation's original relations to either the Valmy or Slaven Formations.

The formation is dominantly composed of fine-grained sandstone, much of it silty, with subordinate interbeds of siltstone, sandy siliceous tuffite shale, and thin, platy, light-tan, light-gray, and yellow-brown chert. There is some cherty shale and a very little yellow-brown quartzite. Much of the sandstone and siltstone is notably feldspathic (see fig. 9, upper photo.; and specimen 608, table 2), and some is true arkose. Some specimens contain chalky and rusty grains that probably are altered feldspar. Euhedral pyrite is very common, and the notably yellow-brown hue of most of the formation, which sets it apart from other sandstones in the area, is doubtless a result of oxidation of the pyrite.

Some sandstone is crossbedded and some ripple marked. It is chiefly in beds a few inches to a few feet thick that are so uniform in resistance to erosion that they are not notably expressed in the topography. Most of these beds show finer laminae $\frac{1}{8}$ to $\frac{1}{4}$ inch thick, somewhat resembling in scale the laminations of the Roberts Mountains Limestone, but the rocks do not readily split on the partings of these laminae.

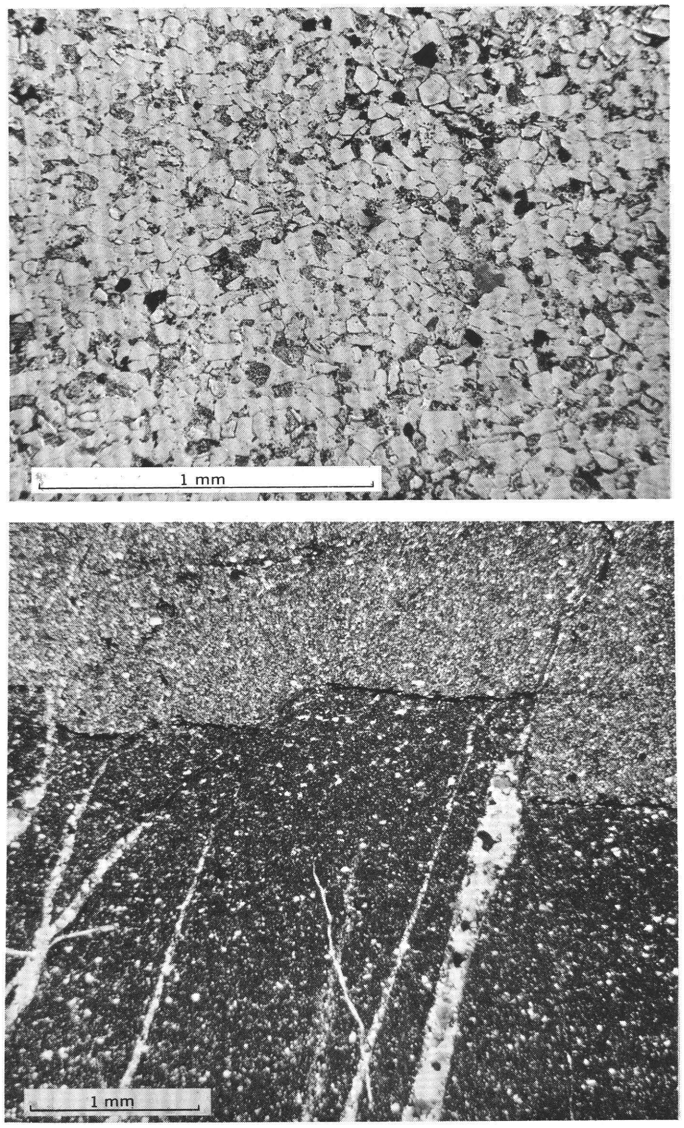


FIGURE 9.—Photomicrographs of rocks of the Elder Sandstone. *Feldspathic siltstone* (*upper photo.*).—Note angularity of grains. Quartz about 80 percent, potassium feldspar about 15 percent, muscovite about 4 percent, accessory augite, apatite, hornblende, green tourmaline, brown tourmaline, and zircon. Plain light. *Chert* (*lower photo.*).—Very fine to medium silt-sized quartz grains (0.005–0.02 mm) make up 70 percent, sericite about 20 percent, coarse silt-sized grains of potassium feldspar about 10 percent. A little organic matter accents the bedding. Crossed nicols.

Both sandstone and siltstone are composed of 70 to 80 percent quartz, 15 to 25 percent potassium feldspar, about 5 percent muscovite, and a little albite. Ghosts of shards in the shaly siltstone suggest that the potassium feldspar and mica may be pyroclastic additions to the quartz silt, which otherwise resembles some silt in the Valmy. There are also a few beds of limy black laminated siltstone composed of about 70 percent calcite, 20 percent quartz, 5 percent potassium feldspar, 2 percent muscovite, and 3 percent authigenic pyrite.

True quartzite—rock that breaks across the grains—is sparse in the Elder Sandstone; this is in marked contrast with the Valmy. The chert of the Elder Sandstone differs from that of the Valmy and Slaven Formations in being a very minor constituent, in layers rarely exceeding 3 feet in thickness. Most of it also differs in color, being nearly everywhere tan to light brown, rather than red, green, black, or dark gray, as in those formations. A few thin beds of dark-gray chert, however, do occur. Furthermore, the chert of the Elder Sandstone contains much more sericite (as much as 20 percent), and as much as 10 percent of potassium feldspar, therein differing very markedly from the very pure silica of the cherts of the Valmy Formation. (See fig. 9, lower photo.) Some of the chert seems to grade across the bedding into sandstone. Some sandstone contains white porcelaneous grains that appear to be devitrified volcanic glass, and some grades laterally into siliceous ash beds.

The structure of the several bodies of the Elder Sandstone is so complex as to preclude an accurate estimate of the formation's thickness. In the type locality the structure is so much disturbed that no closely controlled guess can be made. It seems reasonable, though, in view of the observed dips, topographic relief, areal distribution, and mechanical contacts at both top and bottom, to estimate the thickness as at least 2,000 feet and more probably about 4,000 feet.

Age and correlation

Fossils are sparse; only two generically identifiable collections were obtained from the Elder Sandstone during this survey. These were referred to R. J. Ross, Jr., who, with W. B. N. Berry, reported as follows:

F-93=D 17 SD, from sandstone at alt 6,800 ft., at point 450 ft. S., 3,300 ft. W. of SW cor. sec. 19, T. 28 N., R. 45 E. Collected by M. R. Mudge.

Monograptus cf. *M. regularis* Törnquist; ranges through zones 19–22 of the British section of Elles and Wood.

Age: Late Llandovery.

F-102=D 19 SD, from cherty shale 1½ mi. SW. of Utah mine at alt 6,200 ft., 1,250 ft. N., 1,300 ft. E. of SE cor. sec. 19, T. 28 N., R. 45 E. Collected by James Gilluly.

Monograptus aff. *M. convolutus* Hisinger. The single specimen, although remarkably well preserved under the circumstances, has not been identified specifically with certainty.

Age: Probably Late Llandovery.

Insofar as the Llandoverian age assignment applies to the Elder Sandstone, this formation is the approximate age equivalent of the Roberts Mountains Limestone, but of course it may also include both older and younger strata. Negative paleontologic evidence from a formation so sparsely fossiliferous is of minor value.

Unidentifiable fragments of algae, bryozoa, ostracodes, and brachiopods have also been found in the formation, but these have been of slight value as age clues.

Very few and thin carbonate beds have been found in the Elder Sandstone, and only very thin sandstones in even the westernmost representatives of the Roberts Mountains Limestone in the Mount Lewis quadrangle—the parts richest in clastic components. There is, however, much more silt in the Roberts Mountains Limestone of the Mill Creek window than in that of the Goat window, and still more is present in the Horse Mountain window. This suggests a transition between carbonate and siliceous facies, though it is not emphatic. A minimum of several miles could well have intervened between the respective sites of deposition of the Roberts Mountains and any part of the Elder. Despite this, there is surely no necessity for postulating separate depositional basins even though the two formations were in fact contemporaneous, as remains unproved.

The only other dominantly clastic Silurian rocks so far described from the general region are in the Trail Creek Formation of central Idaho (C. P. Ross, 1937, p. 22); however, the Trail Creek seems to be less sandy than the Elder. In this respect the Trail Creek resembles an unnamed shale and argillite unit exposed in the Cortez Mountains near Four Mile Canyon, a few miles south of the Crescent Valley quadrangle (Gilluly and Masursky, 1965). Black shales, from which *Monograptus* was obtained, have also been recorded from the Simpson Park Range, Tuscarora Mountains, and Pinon Mountains (Roberts and others, 1958, p. 2835). All these units seem nearly enough alike to have been deposited in the same basin. It may therefore be significant that the Trail Creek Formation has been described as in depositional contact with the Phi Kappa Formation in the Wood River region, Idaho (Umpleby, Westgate, and Ross, 1930, p. 23). The Phi Kappa, as just discussed, strongly resembles the Valmy of this area.

HIATUS BETWEEN SILURIAN AND DEVONIAN SYSTEMS

At no place has a depositional contact been recognized between the Elder Sandstone and Slaven Chert. As fossils are so sparse in both formations, however, especially in the Elder, there is no evidence of any significant hiatus—the Elder may include rocks younger than Early Silurian and the Slaven may include rocks older than Middle Devonian. Clearly, much more collecting must be done before we will know how great an hiatus, if any, exists in the western facies section between the Silurian and Devonian Systems.

DEVONIAN SYSTEM

MIDDLE DEVONIAN AND UPPER(?) DEVONIAN SERIES

SLAVEN CHERT

The Slaven Chert is here named from its exposures on the hills along and west of Slaven Canyon in the eastern part of T. 30 N., R. 46 E., where it is widely exposed.

Distribution and topographic expression

The Slaven Chert among the bedrock formations of the area is second only to the Valmy Formation in areal extent. It is widely exposed in Slaven Canyon and on the north face of the Shoshone Range as far west as the mouth of Crum Canyon. Other large exposures lie along the west face of the range between Trout Creek and Crippen Canyon, on the hills south of Harry Creek, on the east side of the range from Corral Canyon south to the edge of the map area, and in the drainage areas of Elder, Feris, and Indian Creeks.

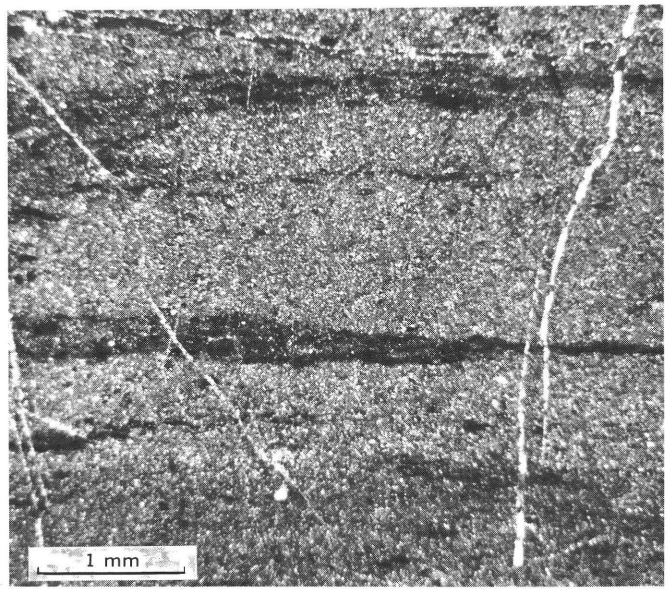
It should be noted that the discrimination of Slaven Chert from chert of the Valmy is not everywhere certain, so that some of these bodies are perhaps not correctly assigned to the Slaven and perhaps others mapped on plate 1 as Valmy should have been classed as Slaven. We were unable to establish criteria for distinguishing small masses. The larger bodies, however, have generally distinctive characteristics and probably are adequately differentiated on the map.

Over much of its extent the Slaven Chert is fairly conspicuous topographically, so that it evidently is fairly resistant to erosion. Except where it has been baked by igneous intrusives, as near O'Haras Peak, its exposures generally lack detail because the generally thin-bedded chert composing so much of the formation breaks into thin plates that are strewn over the surface. (See fig. 26.)

Stratigraphy

The Slaven Chert is in fault contact with older rocks so that nothing is known of its depositional base. Furthermore, structural complexities, poor exposures, and scarcity of fossils generally obscure the overall sequence of the strata. It is not known whether the numerous fault slivers are chiefly laterally equivalent or sequential, and if sequential, in what order.

The dominant rock of the Slaven is black chert, generally in nodular beds 1 to 4 inches thick. It is composed of about 85 percent quartz of very fine to fine silt size, about 10 percent iron oxides and organic matter, and about 5 percent sericite. (See fig. 10, upper photo.) Dark carbonaceous shale generally forms partings between the chert layers, but in places makes up beds 4 to 10 feet thick. Such interbeds are absent from, or at least uncommon in, the chert of the Valmy Formation.



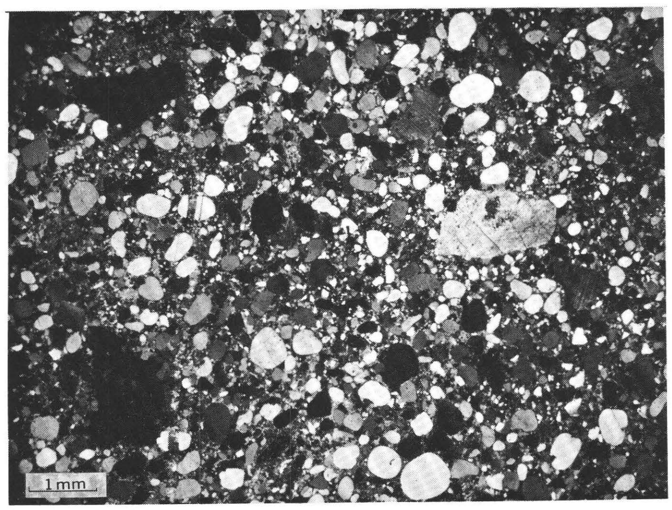


FIGURE 10.—Photomicrographs of rocks of the Slaven Chert. *Chert (upper photo.)*.—Composed of quartz fragments of very fine to fine silt size, sparse grains of fine-sand size, authigenic pyrite (now altered to iron oxides), and sericite and organic shreds. The microlaminae are defined by organic matter and iron staining. Much of the quartz is recrystallized into chalcedonic sheaves and bundles possessing a mosaic fabric. Crossed nicols. *Quartz sandstone (lower photo.)*.—The dark angular grains are shale, the light poorly rounded grain to the right of the center of the photograph is limestone. Note large well-rounded quartz grains and poor sorting. Cement is calcite. Crossed nicols.

Many sections contain a few beds of limy brown-weathering sandstone, from a few inches to 4 feet thick, that have no counterpart in the chert sections of the Valmy. Most of the fossils of the Slaven have come from them. These very poorly sorted limy sandstones (fig. 10, lower photo.) contain angular to subrounded fragments of chert, shale, greenstone, and limestone as much as 2 mm in diameter, making up about 25 percent of the rock; quartz grains ranging from silt to coarse-sand size, making up about 55 percent; and a finely crystalline carbonate matrix. Other beds that lack carbonate and are cemented with sericite could be called graywacke. The larger quartz grains are very well rounded like those of the quartzites of the Valmy; the smaller grains are angular.

Feldspathic siltstone is a subordinate component of the Slaven. This rock is very fine sandy siltstone composed of moderately well sorted angular grains of low sphericity. Quartz makes up about 75 percent of the rock, potassium feldspar about 20 percent, and muscovite, iron oxide minerals, and organic matter make up the rest. The siltstone resembles some beds in the Elder Sandstone. (See fig. 9, upper photo.; and specimens 608 and 264, table 2.)

A few layers of brown-weathering limestone occur near the abandoned Hilltop barite mine, on the north slope of the mountain front west of Slaven Canyon. Most of these are less than 2 feet thick but one (fig. 45) is about 20 feet thick. Similar beds crop out in Crippen Canyon, and in the valleys of Harry and Mill Creeks, but all are subordinate to the associated chert and shale.

Near the Greystone mine (fig. 47), near the divide between Cooks and Elder Creeks, a section of fossiliferous limestone 40 to 60 feet thick crops out. It is so distorted that its thickness and stratigraphic relations are uncertain, but it is intimately associated with chert representative of the Slaven and is here considered part of that formation. The rock is bluish gray on fresh fracture, weathers dark brown, and shows partings of brown shale spaced a few inches to 2 feet apart. Although little except chert, barite, and limestone is exposed in the bulldozed trenches of the Greystone mine, considerable shale and fine-grained sandstone are also probably present here as they abound in the float on the hillside above the mine. Similar rocks are also poorly exposed a few miles to the west near Mound Springs, in the adjacent Mount Moses quadrangle, associated with similarly fossiliferous baritized beds. In reconnaissance mapping (Ferguson, Muller, and Roberts, 1951b) these rocks were included in the Pumpernickel Formation, of doubtful Mississippian age. The fossils show both bodies to be more probably Devonian (p. 40–41) and they are considered to be parts of the Slaven.

Inasmuch as we have been unable to determine any stratigraphic succession among the several rock bodies referred to the Slaven Chert, we are unable to report anything concerning the original upper limit of the formation. All contacts with other pre-Tertiary formations in the area are faults.

Thickness

A reliable estimate of the thickness of the Slaven Chert is difficult because of its poor exposures, contorted structure, and faulted contacts. Many individual fault blocks are several hundred feet thick and exposures between Crum Canyon and the mouth of Slaven Canyon suggest a thickness of more than 2,000 feet. Sections north of the Gold Acres window and between Indian Creek and Mud Spring Gulch also imply thicknesses of 2,000 to 3,000 feet. An estimate of 2,000 feet is probably conservative—perhaps roughly 4,000 feet would not be excessive, but there simply are not enough data to make other than a guess.

Age and correlation

The fauna of the Slaven Chert is sparse both in genera represented and in individuals. The collections helpful in making age assignments and the localities furnishing them are listed in table 5.

TABLE 5.—*Fossils from the Slaven Chert in the Mount Lewis and Crescent Valley quadrangles, Nevada*

[Determinations by: Jean Berdan, 1; W. H. Hass, 2; Jean Berdan and I. J. Sohn, 3; L. G. Henbest, 4; Helen Duncan, 5]

Species	Field colln. No.																					
	F-25	F-26	F-27	F-28	F-29	F-48	F-57	F-58	F-59	F-68	F-71	F-72	F-77	F-78	F-103	F-106	F-109	F-112	F-130	F-160	F-161	F-165
Cervicoinodes sp	---	2	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Icriodus cf. *I. latericrescens* Branson and Mehl	2	2	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
sp	---	2	2	---	2	---	---	---	---	---	---	---	---	2	---	---	---	---	---	---	---	---
Polygnathus linguiformis Hinde	2	---	---	2	---	---	2	---	---	---	2	---	---	---	---	---	---	---	---	---	---	---
sp	2	---	---	2	---	---	---	---	---	---	2	---	---	---	---	---	---	---	---	---	---	2
Polygnathus?	---	---	---	---	---	2	2	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Spathognathodus sp	2	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	2	---	---	---	---
Hindeodella	---	---	---	---	---	2	---	2	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Barlike conodont fragment	---	---	---	---	---	---	2	---	---	---	2	---	---	---	---	---	---	---	---	---	---	---
Ozarkodina sp	---	---	---	---	---	---	---	---	---	---	2	---	---	---	---	---	---	---	---	---	---	---
Tentaculites sp	1	2	2	2	2	1	1	1	---	1	2	1	1	1	---	1	1	---	1	---	1	---
Hibbardia? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Treposella sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Phlyctiscapha? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Sulcicuneus sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Aechmina sp	---	1	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---

TABLE 5.—*Fossils from the Slaven Chert in the Mount Lewis and Crescent Valley quadrangles, Nevada*—Continued

Species	Field colln. No.																				
	F-25	F-26	F-27	F-28	F-29	F-48	F-57	F-58	F-59	F-68	F-71	F-72	F-77	F-78	F-103	F-106	F-109	F-112	F-130	F-160	F-161	F-165
Bollia sp	1	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Schweyerina? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Hellina sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Hollinella sp	1	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Adelphobolbina sp	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Abditoloculina sp. aff. *A. insolita* Kesling	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Abditoloculina sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Subligaculum sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Hanaites sp. aff. *H. platus* Kesling and McMillan	1	1	---	---	1	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Hanaites? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Nezanyslia? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Chironiptrum sp	1	1	---	---	1	---	---	---	---	1	---	---	---	---	---	1	---	---	---	---	---	---
Arcyzona sp	---	1	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Kirkbyella (*Berdanella*) sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Halliella sp	1	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Eukloedenella? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Hypotetragona sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Kloedenella sp. aff. *K. opisthorhysa* Kesling and Kilgore	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Poloniella sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Trinota? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Barychilina? sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Libumella sp	1	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Paraparchites sp	1	1	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Parahealdia sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Tricornina sp	1	1	---	---	---	---	---	---	---	---	---	1	---	---	---	1	---	---	---	---	---	---
Octonaria sp	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Tubulibairdia sp	---	1	---	---	---	---	---	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---
Bairdiocypris sp	---	1	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---
Acanthoscapha sp	---	---	---	---	---	---	---	---	---	---	---	---	---	---	---	1	---	---	---	---	---	---
Pelmatozoan columnals	---	---	---	---	---	---	---	---	1	---	---	---	---	---	4	---	---	---	---	5	1	---
Hexactinellid sponge spicules	---	---	---	---	---	---	---	---	1	---	---	---	---	---	4	---	---	---	---	---	---	---
Fenestellid bryozoan	---	---	---	---	---	---	---	---	1	1	---	---	---	---	4	1	---	---	---	5	1	---
Osagia?	---	---	---	---	---	---	---	---	---	---	---	---	---	---	4	---	---	---	---	---	---	---
Molluskan plates with prismatic (or alveolar?) structure	---	---	---	---	---	---	---	---	---	---	---	---	---	---	4	---	---	---	---	---	---	---
Brachiopod spines	---	---	---	---	---	---	---	---	---	---	---	---	---	---	4	---	---	---	---	---	1	---
Spumellaria	---	---	---	---	---	---	---	---	---	---	---	---	---	---	4	---	---	---	---	---	---	---
Tasmanites?	---	---	---	---	---	---	---	---	---	---	---	---	---	---	4	---	---	---	---	---	---	---

SOURCE AND LOCATION OF FOSSIL COLLECTIONS

F-25. From thin-bedded shaly and limy sandstone at alt 6,080 ft, on the spur W. of the Hilltop barite mine, in the NE¼ sec. 10, T. 30 N., R. 46 E. Collected by James Gilluly and H. E. Malde.

F-26. From thin-bedded limy sandstone at alt 6,000 ft, 1,000 ft NW. of F-25 and about 200 ft stratigraphically higher. Collected by James Gilluly and H. E. Malde.

F-27. From sandstone at alt 5,960 ft, 4,800 ft N., 1,850 ft W. of the SE cor. sec. 23, T. 30 N., R. 46 E. Collected by H. E. Malde.

F-28. From limy sandstone at alt 6,580 ft at a point 3,000 ft N., 4,700 ft W. of the SE cor. sec. 23, T. 30 N., R. 46 E. Collected by H. E. Malde.

F-29. From limy sandstone at alt 6,240 ft at a point 2,000 ft N., 2,600 ft W. of the SE cor. sec. 23, T. 30 N., R. 46 E. Collected by H. E. Malde.

F-48. From limestone at alt 7,000 ft in Crippen Canyon, 550 ft N., 220 ft E. of the SW cor. sec. 10, T. 29 N., R. 45 E. Collected by R. B. Neuman.

F-57. From limy sandstone at alt 6,380 ft, S. of Harry Creek, 2,900 ft S., 400 ft W. of SW cor. sec. 12, in sec. 13 (unsurveyed), T. 28 N., R. 44 E. Collected by R. B. Neuman.

F-58. From Slaven Chert at alt 6,700 ft, 1,600 ft N., 4,600 ft W. of SE cor. sec. 13 (unsurveyed), T. 28 N., R. 44 E. Collected by Olcott Gates and R. B. Neuman.

F-59. From Slaven Chert at alt 6,600 ft, 200 ft N., 2,600 ft W. of SE cor. sec. 13 (unsurveyed), T. 28 N., R. 44 E. Collected by R. B. Neuman and Olcott Gates.

F-68. From calcareous shale and sandstone on S. side of Horse Creek at alt 6,000 ft, center of NW¼ sec. 5, T. 29 N., R. 45 E. Collected by M. R. Mudge.

F-71. From sandy limestone at alt 6,480 ft, 1,800 ft N., 1,500 ft. W. of SE cor. sec. 5, T. 28 N., R. 45 E. Collected by James Gilluly.

F-72. From sandy limestone on spur SW. of junction of North Fork with Mill Creek at alt 6,560 ft, 1,600 ft N., 1,400 ft W. of the SE cor. sec. 4, T. 28 N., R. 45 E. Collected by James Gilluly.

F-77. From thin gray limestone on slope S. of Crippen Canyon at alt 6,200 ft, 2,500 ft S., 850 ft E. of NW cor. sec. 8, T. 29 N., R. 45 E. Collected by James Gilluly.

F-78. From point above Crippen Canyon at alt 6,600 ft, 1,850 ft N., 3,200 ft W. of the SE cor. sec. 8, T. 29 N., R. 45 E. Collected by James Gilluly.

F-103. From shale interbed in chert at alt 5,960 ft, 250 ft S., 300 ft W. of NE cor. sec. 29, T. 28 N., R. 46 E. Collected by James Gilluly.

F-106. From thin gray limy sandstone at alt 7,440 ft, 2,500 ft N., 2,200 ft W. of the SE cor. sec. 6, T. 28 N., R. 46 E. Collected by James Gilluly.

F-109. From thin gray sandy limestone just S. of hillcrest at alt 6,975 ft, 1,300 ft N., 1,850 ft E. of the SW cor. sec. 28, T. 29 N., R. 46 E. Collected by James Gilluly.

F-112. From yellow sandstone E. of saddle of alt 7,750 ft, 1,250 ft S., 2,500 ft E. of the NW cor. sec. 7, T. 28 N., R. 46 E. Collected by James Gilluly.

F-130. From limestone at alt 6,560 ft, 100 ft W. of portal of Lovie mine on E. slope of Bullion Mountain, sec. 30, T. 29 N., R. 47 E. Collected by Olcott Gates.

F-160. From finely conglomeratic sand at alt 6,560 ft, 3,500 ft N. 45° W. of Utah Mine camp, T. 28 N., R. 46 E. Collected by Olcott Gates.

F-161. From sandstone on hilltop at alt 6,360 ft, 2,100 ft N., 2,700 ft W. of the SE cor. sec. 22, T. 29 N., R. 46 E. Collected by James Gilluly.

F-165. From sandy limestone at alt 6,620 ft, 4,000 ft S., 1,000 ft E. of the SW cor. sec. 12, in sec. 13 (unsurveyed) of T. 28 N., R. 44 E. Collected by James Gilluly.

W. H. Hass reported as follows:

Icriodus, Polygnathus linguiformis Hinde, and *Cervicoinides* have reported ranges of Middle and Upper Devonian. *Icriodus latericrescens* suggests a Middle Devonian age.

Polygnathus linguiformis is present in the Genundewa limestone lentil of the Geneseo shale of New York. The Geneseo is classified as early Late Devonian by the U.S. Geological Survey but G. A. Cooper would exclude the Genundewa from the Geneseo and would classify the former as Late Devonian and the Geneseo as Middle Devonian. *P. linguiformis* is present in the basal sandstone of the Chattanooga shale of the Eastern Highland Rim of central Tennessee; it is also present in the Trousdale shale of Pohl in north-central Tennessee, in the Blocher formation of Campbell in Indiana, and in the basal foot of the black shale sequence in the Duffield area of western Virginia. The Trousdale shale of Pohl is considered by me to be a part of the Chattanooga shale, which in the central Tennessee area is classified as Upper Devonian; and the Blocher formation of Campbell is considered to belong to the Upper Devonian part of the New Albany shale. In these two stratigraphic units, as well as in the basal beds of the black shale sequence of the Duffield area of western Virginia, *Polygnathus linguiformis* has been found associated with the brachiopod *Schizobolus*. G. A. Cooper has classified the Trousdale of Pohl and the Blocher of Campbell as late Middle Devonian but I have classified these same units as early Late Devonian chiefly because their correlative in New York—the Geneseo shale—is classified by the U.S. Geological Survey as early Late Devonian. I have also seen specimens of *P. linguiformis* in collections reported to have come from the Middle Devonian limestone of Ohio.

Jean M. Berdan commented as follows concerning the ostracodes from collections F–25, F–26, F–29, and F–106:

With the possible exception of F–25, which contains two genera not represented in the other lots, all these collections appear to represent the same horizon. The presence of many genera of hollinid ostracodes suggests a Middle Devonian age, as at that time there was a great diversification in this group. Several of the genera listed here, as far as is known at present, are confined to the Middle Devonian; for example, the hollinids *Adelphobolbina, Abditoloculina, Subligaculum* and *Hanaites*, and others such as *Treposella* and *Sulcicuneus*. *Hanaites* was originally described from the Givetian of Czechoslovakia, and has since been reported from the Givetian of Russia, the Hamilton Group of New York, and the Middle Devonian part of the Traverse Group of Michigan. Accordingly, on the basis of the ostracodes, the horizon represented by these collections would be considered Middle Devonian in age, and probably Givetian.

Although *Tentaculites* is a long-ranging genus, in the Great Basin it seems to be confined to Devonian rocks, according to C. W. Merriam (oral commun., 1951). Collection F–59 contains many small, unidentifiable brachiopods, corals, bryozoa, pelmatozoan columnals, hexactinellid sponge spicules, and some other ostracodes, but the only generically identifiable fossils are the ostracodes *Tubulibairdia* sp. and "*Bythocypris*" sp., which might be of Silurian age but are more likely Devonian.

Helen Duncan, though unable to identify the material even generically, considered the fragmentary debris of corals, bryozoa, and pelmatozoa in collection F–160 as probably younger than Pogonip and older than Carboniferous, thus lending at least permissive support to a Devonian assignment.

Perhaps the most interesting as well as puzzling fossils collected from the Slaven Chert are baritized brachiopods. Many of these have been found at the Greystone mine (fig. 47) and, a dozen miles farther southwest, the same genus is found more sparsely at the Mound Springs barite mine, in the Mount Moses quadrangle.

These fossils, first collected from the Mound Springs locality in 1950, were identified by several paleontologists as belonging to the genus *Halorella*, a genus hitherto known in North America only from Triassic rocks. In 1955 similar fossils were found at the Greystone barite mine in rocks otherwise indistinguishable from those of the Slaven Chert. The true age assignment thereby became crucial to the understanding of both the areal and historical geology of this area. The internal structure of the specimens from Mound Springs is not well preserved; search by Keith B. Ketner, N. J. Silberling, and James Gilluly in 1957 yielded several somewhat more completely preserved specimens from the Greystone locality.

N. J. Silberling, of the U.S. Geological Survey, reported (Dec. 3, 1957) on the question as follows:

This report is concerned with large plicate rhynchonellid brachiopods from the Shoshone Range, Nev., in and near the Mount Lewis quadrangle. For convenience the collections studied can be separated by locality into four lots, as follows:

A. Mountain Springs (= Mound Springs) barite mine, T. 28 N., R. 44 E., Mount Moses quadrangle; brachiopods preserved as molds in barite; collected by Ketner, 1957 (loc. 410) and by Ketner and Silberling, 1957.

B. Greystone barite mine, NE¼ sec. 26, T. 28 N., R. 45 E., Mount Lewis quadrangle; brachiopods preserved as molds in barite; collected by Ketner, 1957 (loc. 411 and 418).

C. Greystone barite mine, Carico claim; brachiopods preserved in dark crystalline limestone on strike with barite in easternmost quarry; collected by Ketner and Silberling, Aug. 8, 1957.

D. Valley View barite mine, northeast part of Mount Lewis quadrangle; Stanford University collection 36430, labeled as follows: "W side of range, N of Hilltop; approximately 17 mi. SE of Battle Mountain, Nev.; Barium Products Co.; collector: H. E. Wheeler, November 1938." Identified by H. E. Wheeler as *Tetracamera*? or *Camaphoria*? [sic].

The external morphology of the specimens in all four lots is generally the same except for some variation in the strength of the radial ribs. Likewise the internal structures, visible in several specimens from lots A and D and examined by means of serial sections in two specimens from lot C, are in close agreement. Hence the brachiopods from all three barite mine areas apparently represent the same species.

In a report to Gilluly, dated Feb. 26, 1954, Mackenzie Gordon, Jr., identified an earlier collection of brachiopods from the

Mound Springs barite quarry as *Halorella* sp. and assigned this collection to the Triassic. The object of the present report is to plead caution in using these brachiopods for a definite age assignment.

The arguments in favor of the Triassic age assignment are:

1. The external shape and the ornamentation of the Shoshone Range brachiopods is in agreement with specimens of *Halorella* from rocks of known Late Triassic age.
2. Both the Shoshone Range specimens and the Upper Triassic specimens of *Halorella* possess the same kind of simple internal structures, namely a median septum in the brachial valve and dental plates in the pedicle valve.
3. *Halorella*, with which the Shoshone Range specimens show a general morphologic agreement, is known from upper Upper Triassic beds at several localities in the western United States, whereas *Halorella*-like brachiopods have not been described from the western Paleozoic.

Other factors lending doubt to a positive Triassic assignment are as follows:

1. Although both the Shoshone Range specimens and *Halorella* from known Upper Triassic rocks have the same kind of internal structures, there appears to be a consistent difference in the proportions and position of these structures. In the Shoshone Range specimens the dental plates are small and set close together in the beak of the pedicle valve. In contrast, specimens of *Halorella* from the Upper Triassic of the Paradise Range, Nev., and the Blue Mountains, Oreg., have more widely spaced and larger dental plates, which extend anteriorly into the pedicle valve about three times farther than the dental plates in specimens of comparable size from the Shoshone Range. Moreover, the median septa of the Shoshone Range specimens appear to be more prominent than the septa present in *Halorella* from the Upper Triassic.
2. Termier (1936; and in subsequent papers) has applied the name *Halorella* to rhynchonellid brachiopods from the Upper Devonian of Morocco. This author (1938, Soc. géol. France, Compte rendu, no. 7) also includes *Trematospira? baschkirica* Tschernyschew from the Upper Devonian of the Ural Mountains in the genus *Halorella*. Although Termier (1950, Morocco, Serv. Géol., Notes et Mem., no. 74) defends his assignment of the Devonian specimens from Morocco to *Halorella* on the basis of their internal structure, he has not illustrated the details of this structure, and hence comparison with the Shoshone Range specimens is not possible.
3. The preponderance of bedded chert in the sections from which the Shoshone Range specimens were obtained makes these rocks distinctly different from any known Mesozoic rocks in western Nevada. On the other hand bedded cherts are well developed in the Paleozoic sections of areas adjacent to the Mount Lewis quadrangle. I am admittedly strongly influenced by this particular point as I feel that we have a fair knowledge of the Triassic in northwest Nevada and the occurrence of a thick bedded-chert section of Triassic age in one small area would indeed be an anomaly.
4. The Valley View mine (Hilltop Barium mine, on the Mount Lewis topographic map) area, from which the specimens of lot D were obtained, is included in Gilluly's Slaven chert. Gilluly's fossil collection F–26, in the Slaven chert, about a mile southwest of the Valley View mine, produced Devonian ostracodes and conodonts. If the *Halorella*-like brachiopods from the Valley View mine and the Devonian microfossils are in fact from the same section, a Devonian age would be established for these brachiopods and they could be used in turn to date the rocks at the Greystone and Mountain Springs mines. The locality data for lot D are quite vague, but this is the collection mentioned by Gianella (1941) from the Valley View mine. According to Gianella the brachiopods in this collection were identified by Wheeler and Muller as *Camaphoria* [sic]. Presumably Gianella meant *Camerophoria*, but the internal structures of the specimens are not in accord with this genus.

In summary, I do not feel that an unequivocal age assignment can be based on these brachiopods from the Shoshone Range, but the two best guesses are either Late Triassic or Devonian (or at least Middle Paleozoic). According to Mackenzie Gordon, Jr., there are no brachiopods like these known in the Carboniferous and Permian, but it should be mentioned that superficially similar rhynchonellids do occur in the Middle Jurassic of Europe.

Owing to the importance of this diagnosis, both regionally and locally, P. E. Cloud, Jr., of the U.S. Geological Survey, made further studies and reported (Sept. 23, 1958) as follows:

Radiolaria obtained from the residues of rock trimmed from around the brachiopods in these collections proved indeterminate to Dr. Wm. Riedel of the Scripps Institution of Oceanography, a specialist in this group—he considers it unsafe even to guess whether pre-Mesozoic or post-Paleozoic.

So I have restudied and critically compared the brachiopods with the result that I find them to be more like the Upper Devonian (Famennian) specimens from central Morocco which Henri Termier called *Halorella crassicosta* and *H. intermedia* than they are like any Triassic species of *Halorella*. Excluding the smooth and non-septate Upper Triassic species which have been called *Halorella*, there is striking homeomorphy between the Devonian and Triassic shells studied in ribbing, internal structure, tendency toward asymmetry, and presence of opposed anterior sulci, producing a rectilinear margin. The difference is in beak and hinge-line characteristics, so poorly preserved in the Nevada specimens. The Moroccan specimens, however, and one specimen from Mound Springs barite mine lack the sharply defined beak ridges and palintropes which are so distinctive of the Triassic forms. In contrast also, the beak of the Triassic specimens is suberect, while the Famennian specimens from Morocco and that of the better specimen from Mound Springs have a slightly incurved beak.

These differences are great enough to be considered of subgeneric or even generic value among the brachiopods, especially considering the apparent time gap. I know of no named genus having such characteristics, but as an interim designation we might call the Nevada specimens "*Halorella*" aff. "*H.*" *crassicosta* Termier.

It would be unsafe to correlate with a North African horizon on the basis of a single fossil, so the age remains in doubt; but it seems reasonable to consider it Devonian(?), and perhaps Late Devonian.

Thus the balance of evidence indicates that the Slaven Chert is of Middle Devonian and perhaps in part of Late Devonian age.

Correlation

This age assignment makes the Slaven Chert the time equivalent of the upper part of the Nevada Limestone and perhaps of the Devils Gate Limestone of the eastern facies.

In the western facies no demonstrable equivalents have been recognized from other ranges. The fact, however, that rocks closely resembling those at the Greystone mine, and containing identical fossils, have been mapped as Pumpernickel Formation at Mound Springs (Ferguson, Muller, and Roberts, 1951b) raises the question as to the proper assignment of the Pumpernickel Formation elsewhere.

As originally described (Muller, Ferguson, and Roberts, 1951) the Pumpernickel was considered of probable Pennsylvanian age, because of its supposed conformity with the overlying Havallah Formation, which had yielded fusulines of probable Wolfcamp and Leonard ages. The formation was mapped in reconnaissance widely over the adjacent Winnemucca (Ferguson, Muller, and Roberts, 1951a), Golconda (Ferguson, Roberts, and Muller, 1952), and Mount Moses (Ferguson, Muller, and Roberts, 1951b) quadrangles. More detailed work by Roberts in the Antler Peak quadrangle (Roberts and others, 1958, p. 2848) has shown both that the Havallah is at least in part as old as Middle Pennsylvanian, and that it is unconformable on the Pumpernickel. There would, therefore, be nothing incongruous if much of the Pumpernickel Formation turned out to be of Devonian age like that part mapped as Pumpernickel at Mound Springs.

Much of the material included in the Pumpernickel as mapped in the Mount Tobin, Winnemucca, and Golconda quadrangles is chert, with some sandstone and shale; these rocks could be direct equivalents of the Slaven Chert as here described. But much of the Pumpernickel includes volcanic rocks, slate, and phyllite, none of which are present in any considerable amounts in the Slaven. It remains, then, for future detailed work to determine whether the Devonian age of the so-called Pumpernickel at Mound Springs is representative of the formation as a whole. In any event, it is clear that a eugeosynclinal area existed to the west in Devonian time, whence the Roberts thrust moved the materials of the Slaven Chert into the map area.

PRE-PENNSYLVANIAN UNCONFORMITY (THE ANTLER OROGENY)

No rocks of Mississippian or earliest Pennsylvanian age have been recognized in the northern Shoshone Range. Inasmuch as the formations of Pennsylvanian, Permian, and Early Triassic(?) ages are here virtually as highly deformed as the older ones, this stratigraphic hiatus does not necessarily result from orogenic disturbance and unconformity. There are, however, three different tectonic blocks in the Mount Lewis quadrangle in which the Battle Conglomerate, of Middle Pennsylvanian age, rests in depositional contact directly upon the Valmy Formation. These strong unconformities clearly record the Antler orogeny (Roberts, 1951).

This orogeny is more definitely attested—or at least better dated—in the Antler Peak quadrangle, across the Reese River Valley, where the Battle Conglomerate unconformably transgresses the Dewitt thrust, a major thrust that involves Ordovician and Cambrian beds (Roberts, 1951). The Battle Conglomerate is there far less deformed than the rocks on which it rests, although it is itself isoclinally folded at nearby localities. The dating of the Antler orogeny as Middle to Late Mississippian is less clearcut than it was thought to be at the time the orogeny was named. Later fossil discoveries have led to revision of the ages of the Scott Canyon and Harmony Formations, both formerly considered as probably of Mississippian age. Because their faulted contacts with the Valmy are planed off and overlain by the Battle Conglomerate this relation was considered as dating the orogeny as Late Mississippian. The Scott Canyon and Harmony Formations are now classified as of Cambrian rather than Mississippian(?) age (Roberts and others, 1958, p. 2827–2829; see pl. 3, this report). Now, therefore, in the type locality near Antler Peak, the orogeny can only be bracketed between Ordovician (Valmy) and Middle Pennsylvanian times, instead of between the narrow limits of Late Mississippian to Middle Pennsylvanian, as was once thought.

There is, nevertheless, rather persuasive regional evidence that the orogeny was post-Devonian and even post-Early Mississippian. This evidence has been recently reviewed for the region as a whole (Roberts and others, 1958). Local suggestions of the correctness of this, the original, dating lie in the overall similarities in sedimentary facies of the Valmy, Elder, and Slaven Formations. These similarities imply that much the same kind of depositional conditions existed either continuously or at intervals between Early Ordovician and Late Devonian time. Transitional beds between these formations have not been recognized so that it is of course conceivable that they do not represent uninterrupted deposition. It has been emphasized that the Elder Sandstone is notably rich in potassium feldspar and contains siliceous rather than andesitic volcanics, thereby contrasting with the Valmy and Slaven Formations. But the overall mutual similarities of these rocks are so great that it seems gratuitous to assume that they represent three very different geographical environments. Furthermore, rocks of Early Mississippian age, in the Pine Valley area about 8 miles south of Carlin (see figs. 1 and 2), include conglomerates that interfinger with shale and thus seem to record the early stages of the Antler orogeny (Roberts and others, 1958, p. 2839–2840).

To the west, in the Osgood Mountains and northern Hot Springs Range, the Goughs Canyon Formation composed of andesite flows and pyroclastics, lenses of clastic limestone, chert, and shale has yielded fossils of Early and early Late Mississippian age (Hotz and Willden, 1961; Helen Duncan, written commun., 1962). It is in fault contact with Cambrian and Upper Pennsylvanian rocks so that its normal stratigraphic relations are unknown. Perhaps it correlates, at least in part, with the Inskip Formation of the Winnemucca quadrangle (pl. 3). The Inskip, which is allochthonous, as is the Goughs Canyon Formation in the Osgood Mountains, is at least 9,000 feet thick and consists of graywacke, conglomerate, chert, and greenstone—a typical eugeosynclinal association. It is of probable Mississippian age, as determined by Helen Duncan (Roberts and others, 1958, p. 2847).

In summary, then, the evidence is clear that eugeosynclinal conditions continued in some parts of western Nevada at least into Late Mississippian time, though conglomerates testify to nearby uplifts as early as Early Mississippian. The Antler orogeny probably ended eugeosynclinal conditions in the eastern part of the depositional basin but not throughout its extent. The question of dating the orogenies is further discussed in the section of this report dealing with structure.

STRATA YOUNGER THAN THE ANTLER OROGENY

PENNSYLVANIAN SYSTEM

MIDDLE PENNSYLVANIAN SERIES

BATTLE CONGLOMERATE

Name

The Battle Formation was named in the Antler Peak quadrangle (Roberts, 1951), where it consists dominantly of conglomerate but does contain considerable sandstone, shale, and even limestone, from which fossils of Middle Pennsylvanian (Des Moines) age were collected. In the quadrangles here described the formation is so predominantly conglomerate that its local designation should emphasize the fact; accordingly it is here called the Battle Conglomerate.

Distribution and topographic expression

Outcrops of the Battle Conglomerate are few and most are small; they are scattered in the northern part of the Mount Lewis quadrangle:

1. On the divide north of Horse Canyon, about a mile above the mouth.
2. In Rocky Canyon near its head.
3. On the slope southwest of the head of the west fork of Lewis Canyon.
4. On the south side of Horse Canyon, in sec. 10, T. 29 N., R. 45 E.

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5. On the high northeast spur of Mount Lewis.
6. On the west side of the hillcrest half a mile north of the Dean Mine.
7. On the crest of Havingdon Peak.

All these localities lie above the Whisky Canyon thrust fault. The significance of this distribution is discussed in the "Structural geology" section.

The Battle Conglomerate is a ledge-forming unit in this area, but is generally too thin to make imposing topographic features.

Stratigraphy

The basal contact of the Battle Conglomerate is exposed in three of the localities just cited (1, 4, and 5). In each of them, the conglomerate overlies the Valmy Formation. The contact is an obvious unconformity, with a relief of 1 to 2 feet in 10 feet of bedding length, but is rather smoothly worn. In locality 1, the underlying beds are chert; in locality 4, quartzite, forming angular rubble 4 to 10 feet thick; and in locality 5, sandstone. Fragments and boulders of the underlying beds are incorporated in the lower part of the Battle Conglomerate, so there is no question of its depositional contact on the Valmy Formation, even though the tectonic units of which the conglomerate forms a part, and which consist of Valmy, Battle, and local higher beds, are all transported.

Most of the formation is coarse conglomerate, though there are some sandstone beds and, in Horse Canyon, some mottled red and gray marls and mudstones 4 to 10 feet thick. The boulders and pebbles range in size from more than 1 foot to less than 1 inch in length; most are between 2 and 6 inches in diameter and well rounded. They consist of quartz, quartzite, chert, and subordinate argillite in a finer matrix of quartz and subordinate feldspar and rock fragments.

The formation ranges in thickness from about 50 to about 400 feet, probably because of irregularities in the floor upon which it was deposited. At the top the conglomerate gives way rather abruptly to the Antler Peak Limestone, without evidence of disconformity.

Age and correlation

No fossils have been found in the Battle Conglomerate in the Shoshone Range. The lithologic similarity with the Battle Formation of the nearby type locality, however, strongly supports the correlation, especially as the Antler Peak Limestone conformably overlies the conglomerate in both localities.

In the Antler Peak quadrangle, fossils from limestone beds intercalated in the Battle Formation (Roberts, 1951) indicate a Des Moines (Middle Pennsylvanian) age.

This age would make the Battle Conglomerate virtually contemporaneous with part of the Ely Limestone of eastern Nevada (pl. 3) and of the Moleen and Tomera formations of Dott in the Carlin area (Dott, 1955, p. 2286), and perhaps with part of the Brock Canyon Formation of the Cortez Mountains, next to be described.

Twenty-five miles farther northwest, in the Edna Mountains (Golconda quadrangle), the Highway Limestone, as much as 200 feet thick, of Middle Pennsylvanian age, conformably intervenes between the Battle Conglomerate and the Antler Peak Limestone (Ferguson, Roberts, and Muller, 1952; see pl. 3 this report). It is considered by Ferguson, Roberts, and Muller (1952) to be an offshore facies of the Battle Conglomerate and contemporaneous with at least the upper part of the Battle Formation at Antler Peak, the type locality. No comparable beds have been recognized in the northern Shoshone Range.

The Antler Peak Limestone is of Late Pennsylvanian and Early Permian (Wolfcamp) age, according to the late James Steele Williams of the U.S. Geological Survey. There is thus an apparent hiatus involving the latter part of Middle Pennsylvanian time and early Late Pennsylvanian time between the Battle Conglomerate and the Antler Peak Limestone in the type locality, even though no evidence of erosion has been recorded there (Roberts and others, 1958, p. 2843; Dott, 1955, p. 2287). Also no interruption of sedimentation is recognized in the area of this report. This may be taken either as evidence that the Battle Conglomerate as mapped here is somewhat younger than in the type locality or that this is another of the obscure disconformities of considerable age value and slight geometrical expression as has been inferred at Antler Peak. A third possibility, that perhaps should not be cavalierly dismissed, is that the ranges of some fossil species in the Great Basin may differ from their ranges in the better known midcontinent so that the apparent conformity of Battle and Antler Peak is genuine and should be accepted at its face value—the whole Middle and Late Pennsylvanian may be present. On the other hand the strong angular unconformity between the Tomera and Strathearn Formations near Carlin (Dott, 1955, p. 2289) is evidence of an emergence during late Des Moines time only a few score miles to the east of the Shoshone Range. Even though the Battle and Antler Peak Formations are here rootless after traveling an unknown distance from the west, an unconformity at this horizon anywhere in the region would not be surprising.

PENNSYLVANIAN AND PERMIAN SYSTEMS

BROCK CANYON FORMATION

Name and distribution

The Brock Canyon Formation, which consists of dolomite, conglomerate, sandstone and limestone, is here named from its exposures in Brock Canyon on the northwest face of the Cortez Mountains, in and east of the southeast corner of the Crescent Valley quadrangle. The formation extends southward and northeastward along the mountains for unknown distances of at least several miles; it has also been reported from the Simpson Park Range, about 20 miles to the south (R. J. Roberts, oral commun., 1954).

Stratigraphy

The few thousand feet of basal contact exposed in the Crescent Valley quadrangle is everywhere a fault. R. J. Roberts (oral commun., 1954) reported, however, that the formation is autochthonous in the Simpson Park Range, where it has almost the same makeup as here. Accordingly it seems likely that the fault which separates the formation from the underlying Valmy Formation is not of great displacement. Perhaps it represents the dislodgment of the formation from the foundation on which it was deposited by the drag of an overriding thrust plate, such as we know to have operated farther west in Mesozoic time. (See p. 123–125.)

Whatever its structural significance, the fault has produced a much-brecciated mass of dolomite along the contact, and locally has sheared out the basal dolomite member completely. Internally, this lower dolomite member of the formation, however, generally is disturbed only moderately. It consists of dense fine-grained gray dolomite in beds 2 to 12 feet thick and ranges in thickness from a wedge-out to about 240 feet. It is uncertain how much of this variability may be due to structural duplication—certainly the local cutting out of the member is mechanical.

The dolomite is separated from the next higher unit, of conglomerate and sandstone, by another near-bedding fault. The conglomerate and sandstone is finer grained than the Battle Conglomerate and contains few pebbles more than 3 or 4 inches in diameter. The pebbles are of black chert, gray chert, quartzite, quartz, and sandstone, generally well rounded, and contained in a matrix of sand, silt, and finely comminuted rock fragments. Many of these pebbles appear to have been derived from the Valmy Formation. On fresh fractures the rock is gray but it weathers red brown. Local dark and carbonaceous shale wedges are interbedded, and some of these have yielded poorly preserved plant fragments. The member is much faulted and contorted so that its thickness is difficult to evaluate accurately, but it is at least 400 feet and may be as much as twice this.

The uppermost member of the Brock Canyon Formation within the map area is shaly gray limestone with considerable sandstone interbeds, which range from only a few inches to 1 or 2 feet in thickness. This limestone seems conformable upon the underlying coarse clastic member and to be about 600 feet thick. No higher beds are exposed in the area and the total thickness in the Crescent Valley quadrangle is perhaps 1,200 to 1,500 feet. As much as 6,000 feet of the formation, chiefly sandstone, quartzite, siltstone, and chert-pebble conglomerate, has been estimated in the adjoining Frenchie Creek quadrangle (L. J. P. Muffler, written commun., 1961).

Age and correlation

Two small collections of fossils were obtained from the Brock Canyon Formation in this area. One, of very poorly preserved vascular plants, was collected by Olcott Gates and K. B. Ketner from a shale interbed in the conglomerate and sandstone of the middle member, at a point 250 feet northwest of the crest of the 6,820-foot hill, near the southeast corner of the Crescent Valley quadrangle. It was examined by S. H. Mamay of the U.S. Geological Survey, who reported that the plant remains are unidentifiable but may possibly represent the very primitive group of *Psilophyteles*. If so, the age would be Middle Silurian to Middle Devonian. Considerably better material collected in the adjoining Frenchie Creek quadrangle by L. J. P. Muffler was identified by Mr. Mamay (report of Oct. 28, 1959, to L. J. P. Muffler) as either Late Pennsylvanian or Permian on the basis of abundant specimens of *Taeniopteris*.

The second collection was made by R. J. Ross, Jr., from limestone of the upper member of the formation at a point just east of the Crescent Valley quadrangle boundary and about 2,000 feet southwest of the mouth of Brock Canyon in T. 28 N., R. 48 E. This was referred to Mackenzie Gordon, Jr., of the U.S. Geological Survey, who reported that the collection contained *Astartella* sp. and *Naticopsis* sp. These species are similar to those in a somewhat more extensive collection made by Robert Lehner of the U.S. Geological Survey in the Simpson Park Range, about 20 miles to the south. In light of this association in a larger faunule, Mr. Gordon considered the age suggested as Late Pennsylvanian, perhaps as young as Early Permian.

The age suggested by the combined evidence cited would be Late Pennsylvanian or Early Permian. This assignment is also reasonable in view of the regional geology, which suggests that the earliest post-Devonian limestones of central Nevada are Pennsylvanian and Permian (Roberts and others, 1958, p. 2839–2850). It is noteworthy that the formation rests on the Valmy Formation and probably is nearly in its place of deposition. The unconformity beneath it should therefore record the Antler orogeny; it is unlikely that the formation is older than Early Mississippian.

ANTLER PEAK LIMESTONE

Name and distribution

The Antler Peak Limestone was named by Roberts (1951) from its exposures in the Antler Peak quadrangle. Only small masses of the formation are present in the map area, all above the Whisky Canyon thrust and closely associated with the Battle Conglomerate. Except for a small body on the northeast spur of Mount Lewis, and another, poorly exposed along the range front for a quarter of a mile east of the mouth of Lewis Canyon, its exposures are chiefly on the west wall of Lewis Canyon, the ridge crest to the west of it, and on the spur between Whisky and Rocky Canyons. The largest body lies in the head of the south fork of Horse Canyon, and a smaller body is on the divide north of Horse Canyon, about a mile above its mouth.

Stratigraphy

Wherever the two formations are in contact the Antler Peak Limestone seems to lie conformably on the Battle Conglomerate. As was mentioned in describing the conglomerate, the contact appears almost gradational, despite the considerable time gap that has been inferred to intervene between them.

The Antler Peak Limestone consists almost wholly of light-gray limestone, locally containing some irregular chert nodules, in beds that range from a few inches to 5 feet in thickness. Most of it is rather thick bedded, averaging perhaps 3 or 4 feet. Some is coarsely granular and some almost aphanitic. A few beds are shaly and there are one or two beds of limy yellow sandstone on the spur west of Whisky Canyon. The formation is about 700 feet thick; it is generally so highly deformed and each block is individually so small that no attempt was made to recognize the many members distinguished by R. J. Roberts (oral commun., 1952) in the type locality.

Age and correlation

The Antler Peak is a highly organic limestone, and fossil fragments abound in it, yet they are so difficult to free from the matrix that few are identifiable. Hence the age is not as accurately known as could be wished. Fossils from it collected during this survey are listed in table 6.

The paleontologists have supplied the following comments. As to collection F–16, Mackenzie Gordon, Jr., reported:

> In this list no attempt has been made to supply specific names, as not enough features of the shells are preserved for precise determinations and some of the species are probably undescribed. Similar specimens of *Chonetes* (the deeply sinuate form), *Buxtonia*, and *Cancrinella* occur with *Waagenoconcha* in Roberts "H" unit of the Antler Peak formation in the Antler Peak quadrangle. On this basis, the collection is considered to be Antler Peak, although the composition of the fauna in itself would not restrict the age assignment more narrowly than between Early Pennsylvanian and Early Permian.

TABLE 6.—*Fossils from the Antler Peak Limestone, Mount Lewis and Crescent Valley quadrangles, Nevada*

[Determinations by: J. S. Williams, 1; Helen Duncan, 2; Mackenzie Gordon, Jr., 3. Doubtful determinations queried]

Species	Field colln. No.														
	F-16	F-35	F-36	F-18	F-19	F-20	F-21	F-31	F-32	F-39	F-41	F-43	F-44	F-90	F-38
Caninia sp.								2	2						
Carruthersella sp.	2														
Neokoninckophyllum sp.	?2														
Crinoids, indet.				1	1			1	1						1
Fenestella sp.						?2		2							
Fistulotrypa sp.											2				
Hexagonella sp.	?2			?2											
Polypora sp.	2			2		2		2			2		?2		
Lophophyllidium sp.				2											
Rhabdomeson sp.				2		2									
Rhombopora sp.	2		2	2	2					2					
Rhomboporoid, indet.						2	2	2							
Rhomboporella sp.	2			2						2	2				
Rhombotrypella sp.			2	2							2				
Stenoporoid, indet.		2		2						2					
Streblotrypa sp.				?2											
Cystodictya sp.				2											
Ambocoelia sp.	1	1	1				?1	1						?3	
Buxtonia sp.	3		?1							3					
Cancrinella sp.	3									3					
Chonetes sp. (radially lirate)	3														
sp. A. (prominent sinus)	3														
Chonetid, indet.			1											3	
Composita sp.				?1					?1	?1					?1
Crurithyris sp.	3											3	3		
Dictyoclostus sp.				1							3				
Dielasma sp.											?3				
Hustedia sp.	3				1						3				
Juresania sp.	3				1				?1						?1
Linoproductus sp.											3				
Marginifera sp. A (coarse costate)	3										?3				
sp. B (nearly smooth)	3														
Neospirifer sp.	3								?1						?1
Productus sp. (large costate)	3														
sp. indet.					1			1	1						1
Spiriferina sp. indet.	3			1											
Astartella											3				
Pleurotomariid gastropod, indet.	3								?1	1					1
Straparolid gastropod, indet.			1												

LOCATION OF FOSSIL COLLECTIONS

F-16. From point at about 5,920 alt, on ridge SW. of Betty O'Neal mine, in sec. 22, T. 30 N., R. 45 E. Collected by Helen Duncan, Jean Berdan, R. B. Neuman, and James Gilluly.

F-18. From point at alt 8,680 ft, 2,900 ft S., 1,600 ft W. of NE cor. sec. 1, T. 29 N., R. 45 E. Collected by H. E. Malde and James Gilluly.

F-19. From point at the head of the SW. fork of Lewis Canyon, alt 8,280 ft, 2,100 ft N., 2,400 ft W. of SE cor. sec. 2, T. 29 N., R. 45 E. Collected by James Gilluly.

F-20. From point at alt 8,260 ft, 3,150 ft N., 3,500 ft W. of SE cor. sec. 2, T. 29 N., R. 45 E. Collected by James Gilluly.

F-21. From point at alt 8,200 ft, 450 ft S., 1,500 ft E. of NW cor. sec. 2, T. 29 N., R. 45 E. Collected by Olcott Gates and James Gilluly.

F-31. From point on divide between Lewis and Horse Canyons at alt 8,600 ft, 1,000 ft N, 3,000 ft W. of SE cor. sec. 2, T. 29 N., R. 45 E. Collected by James Gilluly.

F-35. From point on divide W. of Whisky Canyon, at alt 6,620 ft, 550 ft W., 3,200 ft N. of the SE cor. sec. 27, T. 30 N., R. 45 E. Collected by Helen Duncan, Jean Berdan, R. B. Neuman and James Gilluly.

F-36. From point just W. of divide W. of Whisky Canyon, at alt 7,000 ft, 650 ft W., 2,000 ft N. of SE cor. sec. 27, T. 30 N., R. 45 E. Collected by Helen Duncan, Jean Berdan, R. B. Neuman, and James Gilluly.

F-38. From point at alt 6,370 ft, 400 ft W. of SE cor. sec. 32, T. 30 N., R. 45 E. Collected by James Gilluly.

F-39. From ridge at alt 6,340 ft, 250 ft N., 500 ft W. of SE cor. sec. 32, T. 30 N., R. 45 E. Collected by James Gilluly.

F-41. From point on spur W. of Mount Lewis at alt 8,400 ft, 100 ft N., 500 ft E. of SW cor. sec. 2, T. 29 N., R. 45 E. Collected by Olcott Gates, R. B. Neuman, E. M. Shoemaker, and James Gilluly.

F-43. From point at alt 7,920 ft 500 ft S., 550 ft E. of NW cor. sec. 11, T. 29 N., R. 45 E. Collected by R. B. Neuman.

F-90. From point on the divide between Whisky and Rocky Canyons at alt 7,200 ft, SE¼ sec. 27, T. 30 N., R. 45 E. Collected by M. R. Mudge.

With respect to the corals of this collection, Miss Duncan commented that the Russian species of *Neokoninckophyllum* occur mainly in the Middle and Upper Carboniferous, but the genus occurs also in beds currently (1962) assigned to the Lower Permian. *Carruthersella* and its close relatives are well represented in the Upper (Middle and Upper of the Russian terminology) Carboniferous of Europe and Asia, although they first appear in the Lower Carboniferous. Actually the coral association might be of Mississippian rather than of Pennsylvanian age inasmuch as both genera belong to groups that range through the Carboniferous and into the Permian. Fortunately, however, Miss Duncan was able to identify fragmentary bryozoans whose general aspect very strongly suggested the Antler Peak Limestone. Even though the stratigraphic ranges are too little known for confidence, she inclines to an assignment to the Late Pennsylvanian or Early Permian.

Age assignments of the other collections identified by James Steele Williams (JSW), Helen M. Duncan (HMD), and Mackenzie Gordon, Jr. (MGJr) were:

F-18..Late Pennsylvanian (HMD); late Paleozoic (JSW).
F-19..Pennsylvanian or Permian (HMD); Permian (JSW).
F-20..Pennsylvanian or Permian (HMD).
F-21..Late Paleozoic (JSW).
F-31..Early Pennsylvanian to Permian (JSW).
F-32..Early or Middle Pennsylvanian (HMD).
F-36..Pennsylvanian or Early Permian (JSW and HMD).
F-38..Early Pennsylvanian to Permian (JSW).
F-39..Late Pennsylvanian or Early Permian (JSW).
F-41..Late Pennsylvanian or Permian (HMD); Early Pennsylvanian to Early Permian (MGJr).
F-43..Devonian to Permian (MGJr).
F-44..Devonian to Permian (MGJr).
F-90..Mississippian to Permian (MGJr).

Although, as noted in the individual lists, the range of many of these groups of fossils is very much wider, the general aspect of the assemblages is such that Miss Duncan and Messrs. Gordon and Williams all incline to correlate these rocks with the Antler Peak Limestone in the Antler Peak quadrangle. The most probable assignment seems to be Late Pennsylvanian and Early Permian. Although one or two collections suggest an Early or Middle Pennsylvanian age, they do not, apparently, contain fossils that closely resemble forms that occur in the Highway Limestone in the Edna Mountains (Golconda quadrangle). But several of these collections are associated with others of Early Permian aspect. Accordingly the suggestion of an Early or Middle Pennsylvanian age is not considered compelling and the rocks are correlated rather confidently with the Antler Peak Limestone of the type locality as Late Pennsylvanian and Early Permian.

This assignment implies correlation with the Brock Canyon Formation of the Cortez Mountains, the Strathearn Formation of Dott in the Carlin-Elko area, and in part with the Garden Valley Formation of the eastern Roberts Mountains and the Carbon Ridge Formation of the Eureka district. All these formations contain considerable amounts of coarse siliceous clastic rocks, which imply notable relief near their areas of deposition.

Although the Antler Peak Limestone is allochthonous in the northern Shoshone Range, it is autochthonous at Antler Peak a few miles to the northwest. The pattern of the depositional areas implied by the distribution of all these clastic limestones of Late Pennsylvanian and Early Permian age and by their common association with coarse conglomerates seems one of an archipelagic sea, dotted with islands of high relief. The Havallah Formation, which is, in part at least, contemporaneous with the Antler Peak and Brock Canyon, is a dominantly clastic formation but of fine grain compared to the Brock Canyon and Garden Valley Formations. Though widespread in the Mount Tobin, Sonoma Range, and Antler Peak areas, the Havallah is allochthonous wherever exposed. Its depositional site may have been considerably farther to the west or north so that the paleogeography of the time is wholly speculative.

HAVALLAH FORMATION

Name and distribution

The Havallah Formation was named in the Tobin Range, about 40 miles to the west, by Muller, Ferguson, and Roberts (1951). Only two small bodies are referred to this formation in the area of this survey. These are both on the spur west of Mount Lewis, one on either side of the divide between the heads of Horse Canyon and Lewis Canyon. Both are fault blocks closely associated with other fault blocks composed of Battle Conglomerate, Antler Peak Limestone, and China Mountain(?) Formation. Obviously their present situations bear no necessary relation to their sites of deposition nor original stratigraphic relations.

Lithology

The only rock variety recognized in the fault blocks mapped as Havallah on plate 1 is calcareous mudstone that approaches an argillite in lithification. Faint color banding of darker and lighter gray at spacing of a few millimeters may represent bedding but it does not control the fracture, which is conchoidal. The argillite weathers dark rusty brown and contains glistening carbonaceous—almost graphitic—corkscrew-shaped markings, with scalloped and feathery edges. These presumably algal remains are identical with those in the Havallah in Hoffman Canyon in the Tobin Range. According to H. G. Ferguson (oral commun., 1950) these are the features referred to in the description of the type Havallah Formation as "feathery algae." They seem to differ from any other organic features in the experience of the many geologists to whom we have shown our specimens. Accordingly the presence of such peculiar identical markings in identical lithology seems an adequate ground for correlating the local rock bodies with the type Havallah. The "algal" corkscrew-like markings are fairly abundant and not difficult to find. Worm tracks are also common.

The bedding is so obscure that little can be said about the thickness exposed in this area. If the vertical dips of the southern mass are not isoclinal, a thickness of about 1,000 feet is present.

Age

No fossils diagnostic of age have been found in this area; age assignment is therefore based on data from the Antler Peak quadrangle, whence the only specifically identifiable fossils thus far known from the formation have been obtained. Fusulinids of Wolfcamp (Early Permian) and possibly Leonard (Early Permian) age have been collected from the upper part of the section in the Antler Peak area (Roberts, 1951); the middle member has not yielded fossils; the lower member contains fusulinids assigned by Raymond Douglass of the U.S. Geological Survey to an Atoka age (Middle Pennsylvanian). So far as known, the formation contains no breaks in sedimentation.

LATE PERMIAN HIATUS

Rocks of Late Permian age are unknown in the northern Shoshone Range. Relations in China Mountain (Golconda quandrangle), 25 miles to the west (fig. 1), indicate that the Koipato Formation, of middle Permian to Early Triassic age (Norman J. Silberling, oral commun., 1957), overlaps a thrust fault that brings Pumpernickel Formation over Havallah—a post-Wolfcamp structure having thus been beveled before the middle of Permian time (Ferguson, Roberts, and Muller, 1952). Farther north, in the Edna Mountains, the Edna Mountain Formation of Phosphoria age overlaps a folded sequence of Antler Peak strata and rests with sharp unconformity on the Cambrian (Ferguson, Roberts, and Muller, 1952). These two unconformities are on opposite sides of the Golconda thrust, which brings different facies of roughly equivalent ages into juxtaposition. They therefore were doubtless formed in positions several tens of miles apart, and the structures they transect need not have been formed at precisely the same time.

The Koipato Formation wedges out and has not been found east of China Mountain, which was apparently near the original eastern limit of deposition. The Edna Mountain Formation is represented in small patches in the Antler Peak quandrangle but not farther east. It is everywhere rootless in that quadrangle but autochthonous farther west in the type locality in the Edna Mountains. Although the Edna Mountain Formation may have originally covered the area of the Mount Lewis-Crescent Valley quadrangles, it probably derived some of its sediments from here, as it is not known to be autochthonous anywhere east of the type locality. The coarseness of the conglomerates in the China Mountain Formation, of Early and Middle Triassic(?) age, suggests that high relief prevailed while the formation was being deposited but inasmuch as these rocks are all allochthonous also, inferences based on them apply to a locale some unknown distance westward, rather than to the immediate vicinity of the northern Shoshone Range. It seems more than likely that the map area was undergoing erosion during both Late Permian and Early Triassic time and that even the unknown source area (whence the thrust blocks of Triassic rocks were brought) was likewise being eroded during the Late Permian.

TRIASSIC(?) SYSTEM

LOWER AND MIDDLE(?) TRIASSIC(?) SERIES

CHINA MOUNTAIN(?) FORMATION

Name and distribution

The type locality of the China Mountain Formation is in China Mountain, in the southwest part of the Golconda quadrangle, about 25 miles west of the map area (Ferguson, Roberts, and Muller, 1952). The formation there rests in erosional unconformity on the Havallah and Koipato Formations in the upper plate of the Golconda thrust. The formation has not been identified to the east of China Mountain except with some question in the Mount Lewis quadrangle. Here its exposures form a discontinuous band a few hundred feet wide, extending from a point on the west wall of Lewis Canyon at the south boundary of T. 30 N. to the spur west of Whisky Canyon just south of the northern mountain front.

Stratigraphy

In the type locality, the formation consists of conglomerate, sandstone, shale, and impure dolomite aggregating about 500 feet in thickness. The basal conglomerate is dominantly of chert and quartzite pebbles with boulders as much as 2 feet across. Most of the formation seems derived from the Havallah; some fine-textured conglomerate derived from the lavas of the Koipato is restricted to the upper part (Ferguson, Roberts, and Muller, 1952).

In the Shoshone Range the formation here referred to as China Mountain(?) is much faulted and several fault blocks contain no older rocks. A few of them, however, expose depositional contacts of the formation on the Antler Peak Limestone. The unconformity is channeled but not notably angular.

The formation is here best exposed on the divide between Whisky and Rocky Canyons, where the lower part of the China Mountain(?) Formation is a very well sorted limy sandstone about 20 feet thick, which weathers to a pinkish buff and grades upward into poorly bedded maroon mudstone about 30 feet thick. This is overlain by well-bedded green siltstone, 20 feet; poorly exposed maroon siltstone, 50 feet; and a 40-foot thickness of buff-weathering conglomerate containing rounded quartz pebbles an inch across and angular fragments of limestone as much as 4 inches across.

This buff conglomerate gives way upward to about 200 feet of maroon conglomerate, which contains boulders of fossiliferous limestone as much as 4 feet long, in a matrix of maroon mudstone. This member is almost surely a fanglomerate. The top member exposed is a very limy yellow sandstone about 50 feet thick. The total thickness is estimated at about 400 to 500 feet.

About a mile farther south, in sec. 35, T. 30 N., R. 45 E., there seems to be less maroon mudstone in the formation, but this may be the result of bleaching by the adjacent intrusive body.

The materials in the coarse fraction of the fanglomerate and conglomerate include quartz, quartzite, limestone, greenstone, white sandstone, chert, and red jasper. No felsites such as comprise much of the Koipato Formation farther west were found. The materials could all have been supplied from the Valmy, Slaven, and especially from the Battle Conglomerate and Antler Peak Limestone. Many of the limestone boulders have been definitely identified as coming from the Antler Peak on the basis of their fossil content.

Fossils from individual boulders were segregated and submitted for paleontological examination. From one boulder Mackenzie Gordon, Jr., identified:

Dielasma? sp.
Reticulariina? sp.
Hustedia sp. *A*.
Crurithyria? sp.

From another:

Buxtonia sp.
Productid indet.
Rhynchopora sp.
Spiriferinid indet.

And from a third:

Reticulariina sp.
Hustedia sp. *A*.
Composita? sp.
Bryozoan.

Helen Duncan identified from one boulder a syringoporoid coral resembling *Pseudoromingeria*, perhaps *Drymopora*?, and from another the bryozoan *Rhomboporella*. From another boulder she recognized fragments of fistuliporoid? and rhomboporoid bryozoans.

James Steele Williams identified, from one boulder, fragments of spiriferoid and other brachiopods, and from another

Crinoid columnals.
Juresania.
Indeterminate productid and other brachiopods.
Two indeterminate gastropods.

Though no single collection is diagnostic, in the opinion of these paleontologists, all these lots suggest Late Mississippian to Pennsylvanian or Permian age and all are consistent with derivation from the Antler Peak Limestone.

Age and correlation

The only fossils found in the matrix of this conglomerate and considered indigenous to it were examined by E. L. Yochelson of the U.S. Geological Survey. He reported that they are indeterminate capuliform gastropods, of little value in age determination, though marine and perhaps more likely post-Paleozoic than earlier in age.

The oldest rocks in contact with and also younger than the China Mountain(?) Formation are the intrusives of presumable middle Tertiary age. There is therefore no way of proving that the formation is of Triassic age, as is implied by correlation with the type China Mountain. This age assignment is of course tentative; it is based on the general resemblance in lithology and stratigraphic position to the China Mountain Formation in the Golconda quadrangle.

It should be pointed out that, at the type locality, the China Mountain Formation is part of the upper plate of the Golconda thrust (Ferguson, Roberts and Muller, 1952). On the other hand, everywhere in more westerly localities, the Valmy is confined to the lower plate of that thrust. In the Shoshone Range, however, the Whisky Canyon thrust sheet, which contains the formation here called China Mountain(?), is overridden by the Valmy in the higher Pipe Canyon thrust sheet. H. G. Ferguson of the U.S. Geological Survey has on this ground tentatively objected to the correlation here suggested (oral commun., 1954).

The objection does not seem insuperable as the Pipe Canyon faults may be younger than the Golconda. The Whisky Canyon fault carries Havallah, Battle, Antler Peak, and China Mountain(?) in its upper plate, yet these represent both plates of the Golconda thrust in Antler Peak and more westerly localities. There is nothing incredible about the possibility that masses of lower plate rocks may be sheared off and mixed with those of the upper plate of a major thrust. It is, therefore entirely possible that the Whisky Canyon and Golconda faults are the same.

Too little is known of structures to the west to make profitable a discussion of the many other possibilities. Suffice it to say that we consider the rocks here called China Mountain(?) to resemble that formation more than they do any other formation younger than Antler Peak known within a radius of 100 miles. These include such formations as the Newark Canyon of Eureka (Nolan and others, 1956, p. 69) and the King Lear of Western Humboldt County (Willden, 1958, p. 2382–2391), both of Early Cretaceous age, and the Edna Mountain of Permian age. The presence of marine gastropods, even if unidentifiable with certainty, points to an early Mesozoic rather than to a later age. The uncertainty of the correlation seems adequately indicated by the query attached to the name.

LOWER TRIASSIC(?)-MID-TERTIARY HIATUS

No stratified rocks of age intermediate between Early Triassic(?) and mid-Tertiary have been recognized in the northern Shoshone and nearby ranges. All pre-Tertiary rocks in the area have been intensely deformed into fault blocks clearly involved in large deformation. One orogenic episode, the Antler orogeny, appears from regional evidence to have occurred in Early or mid-Mississippian time. But, since the Lower Triassic(?) rocks are also involved in major thrusts, there must have been one or more later orogenies in Mesozoic or early Tertiary time.

Evidence for Mesozoic and early Cenozoic orogeny is indeed abundant in the region. Nolan (Nolan, Merriam, and Williams, 1956, p. 69–70) has pointed out the strong unconformity near Eureka that separates the Lower Cretaceous Newark Canyon Formation from the Carbon Ridge (Permian) and older formations, and that the Newark Canyon is itself strongly deformed.

Thrust faults involving rocks as young as Late Triassic are known in the Mount Tobin quadrangle (Muller, Ferguson, and Roberts, 1951). In the Mount Tobin quadrangle also there is excellent evidence of post-Havallah and pre-Koipato (post-Leonard(?)-pre-Late Permian) and of post-Koipato and pre-Tobin (post-earliest Triassic but still in Early Triassic) deformations. There is, then, nothing unusual in our inference of two orogenies in the northern Shoshone Range—one Mississippian, the other post-Early Triassic(?), perhaps Jurassic. To facilitate reference we refer to this second orogeny, of probable Mesozoic age, as the Lewis orogeny from relations near Mount Lewis (p. 123).

The oldest sedimentary Cenozoic rocks in the area are of Miocene age. These rest on an erosion surface of considerable relief, carved across thick piles of thrust sheets in such a way that literally miles of structural relief have been transgressed by the unconformity. So far as the local record can be read, there is every likelihood that the area was one of almost uninterrupted erosion from some time during the mid-Mesozoic to the mid-Tertiary.

TERTIARY SYSTEM

GENERAL FEATURES

The rocks of Tertiary age in the northern Shoshone Range record intrusion, both of deep-seated and volcanic masses, widespread explosive volcanism, and later flood eruptions of lava. Many sedimentary rocks are intercalated between the abundant layers of tuff, welded tuff, agglomerate, and lava flows that buried an irregular topography cut in the highly deformed Paleozoic and lower Mesozoic(?) rocks. Subsequent erosion destroyed much of the volcanic cover, leaving only remnants scattered around the margins of the range and in a few fault blocks within it.

Most of the igneous rocks, both intrusive and extrusive, are intermediate between quartz diorite and granite in composition: granodiorite, dacite, quartz monzonite, and quartz latite. Rhyolite, andesite, basalt, and quartz diorite are subordinate.

The igneous rocks appear to fall into three age groups, though many local masses cannot be assigned definitely to any particular epoch because they are isolated geographically from the areas where the age distinctions are evident.

The oldest of the igneous episodes is represented by the course granodiorite of Granite Mountain, and perhaps also by the similar stocks on the ridge west of Goat Peak. The stock at Granite Mountain, at least, is definitely older than the bulk of the igneous rocks of Mount Lewis. The evidence lies in the fact that the basal tuffs erupted from the Mount Lewis vents in Indian Creek valley contain many very large waterworn boulders of coarse granodiorite identical with that on Granite Mountain.

Furthermore, quartz latite tuff of the mid-Tertiary cycle rests unconformably on a small plug of granodiorite east of Tub Spring Gulch on the east side of Granite Mountain. Pebbles of granodiorite also form inclusions in the breccia of Mount Lewis. Enough time must have elapsed between the two igneous episodes for erosion to have deroofed the Granite Mountain stock (surely erosion of many hundred feet was involved) before the onset of the volcanic episode of the Mount Lewis vents. Because the quartz diorite plugs west of Goat Peak so much resemble the Granite Mountain mass, they probably also represent the same (earlier) episode.

There seems to be no reliable way of establishing the age sequence of most of the rest of the intrusive and extrusive igneous rocks of the area: they are here described simply in accordance with their areal distribution. The basaltic andesite flows of the Mal Pais, however, appear to rest on an erosion surface cut across all other rocks and are therefore considered to belong to a third, still younger, igneous episode.

The tenuous age assignments of the several sequences are discussed after the descriptions of their geologic and petrologic relations.

IGNEOUS ROCKS OF EARLY TERTIARY AGE

GRANITE MOUNTAIN STOCK

DISTRIBUTION AND GEOLOGIC RELATIONS

The Granite Mountain stock occupies an area of about 6 square miles between Hilltop in the Mount

Lewis quadrange and the head of Mud Spring Gulch in the Crescent Valley quadrange. The stock is named for Granite Mountain, which is the highest ridge within it. Alluvium at the head of Corral Canyon—"The Park" on the map (pl. 1)—conceals any connection between the intrusive near Hilltop and the main mass of Granite Mountain, but several outcrops of granodiorite along the southwest side of the Park suggest that the intrusive rocks in the two areas comprise one body, continuous except for the break by the Corral Canyon fault. The small stock of granodiorite near Tenabo in the Crescent Valley quadrangle may be connected at depth with the Granite Mountain stock, as the belt of Devonian chert and siliceous shale that extends southeast from the Granite Mountain area toward Tenabo has been considerably altered and hornfelsed.

At Hilltop and Tenabo, the granodiorite is poorly jointed and weathers to form rounded subdued topography. On Granite Mountain, however, the stock has a blocky rugged topography controlled largely by the well-developed vertical joints and sheeting. The steep face west of O'Haras Peak is a fault-line scarp whose steplike profile is due to the jointing. The east face of the stock from Granite Mountain to Mud Spring Gulch is a slope that follows the sheeting.

Uplift and tilting of the stock along the Corral Canyon fault and subsequent erosion have modified the original shape of the intrusion. Prior to the faulting, it was probably continuous from the Hilltop area to Mud Spring Gulch. The stock cut across both small-scale (on a scale of tens of feet) and large-scale (on a scale of thousands of feet) structures in the wallrocks without notably deflecting them. There is therefore no evidence of forceful intrusion, at least at the present levels of exposure. Nor is any structural or stratigraphic control for the original shape or location of the stock apparent.

The contact with Paleozoic rocks along the southeast side of the stock is regular, with few embayments, apophyses, or dikes. A small satellitic plug which cuts the local structures of the Slaven Chert crops out on the south slope of Bullion Mountain. Good exposures near the Gray Eagle mine and near Rock Spring show that the contact is either vertical or dips steeply southwestward and truncates Paleozoic chert and shale. In contrast, the contacts along the east and northeast sides of the stock are very irregular, with many dikes and small satellitic plugs. Exposures are too poor and relief too low to prove the attitude of the contact here, but the general pattern suggests that the roof of the stock may dip generally northeastward. No breccia nor swarms of inclusions were seen along any contacts between granodiorite and Paleozoic rocks.

The intrusion of the small crescent-shaped plug on the south slope of Bullion Mountain brecciated the wallrocks both along the contact between granodiorite and chert and in an offshoot that cuts hornfelsed shale. Figure 11 is a field sketch of this breccia dike. The breccia is made up of angular fragments and blocks, some as much as 10 feet long, of the immediately adjacent sedimentary rocks. A few of the large blocks close to the margins of the dike have the same orientation as the adjacent wallrocks; others have been rotated and tilted. The breccia is cemented by cream-colored quartz and chalcedony. Small apophyses of granodiorite invade the breccia. In a few places the breccia appears to be folded and contorted; but most of it shows no sign of deformation. As slickensides, fault gouge, and mullions are lacking, this breccia does not appear to be fault breccia. Perhaps it formed during the intrusion by rock bursting along a crack opened by the granodiorite magma.

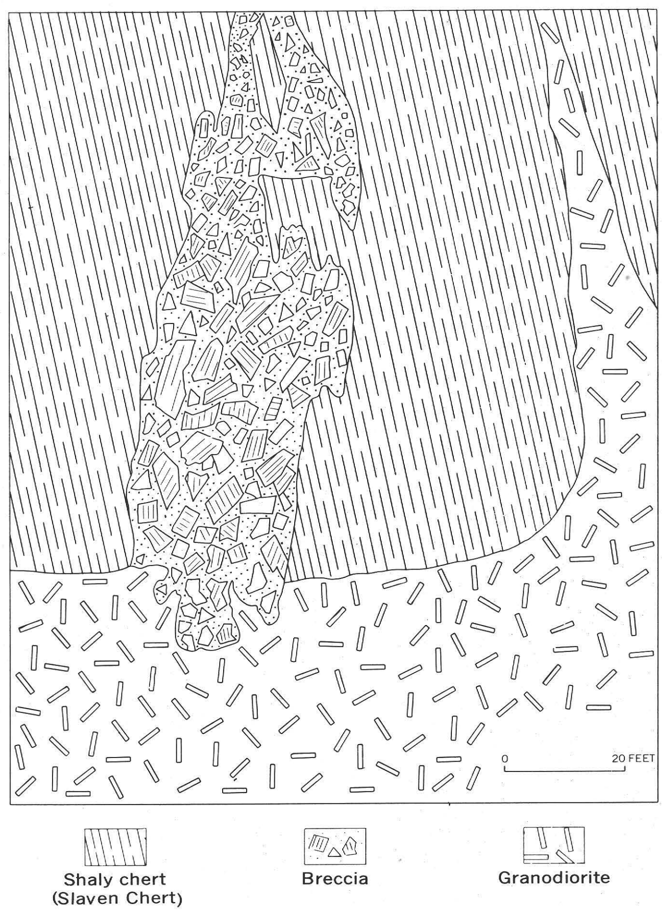


FIGURE 11.—Sketch of breccia dike on south slope of Bullion Mountain.

Potash metasomatism took place along the contact of granodiorite with chert and shale of the Slaven Chert in the gully above the Gray Eagle mine. About 50 feet of slide rock separates the granodiorite from the nearest exposure of metasomatized wallrocks, which consist of coarse-grained, sugary-textured rocks containing biotite, quartz, and feldspar, interlayered with granular, primarily quartzose beds. Thin laminae, from ¼ to 1 inch thick, consist of an aphanitic, gray porcelaneous material not identifiable with a hand lens. The bedding thickens on the noses and thins on the flanks of the isoclinal folds in the metasomatized rocks.

The microscope reveals that the coarse-grained feldspathic beds consist of quartz (50 percent), potassium feldspar whose optic angle is 55° (40 percent), and biotite and muscovite (10 percent). The quartz and K-feldspar form a granular mosaic, the grains of which are in the 0.05–1 mm size range. The biotite forms euhedral crystals as much as 2 mm long. The muscovite occurs as small shreds between quartz and feldspar grains and also as replacement of biotite. The granular siliceous beds have the same texture and minerals as the feldspathic ones but feldspar, biotite, and muscovite are very subordinate compared with quartz. The thin porcelaneous layers are a granular intergrowth of apatite, K-feldspar, and quartz.

From this outcrop, nearest the granodiorite, the country rock can be traced southward almost continuously. The amounts of K-feldspar and biotite gradually decrease; 500 feet from the first exposure the rocks are typical hornfelsed chert and siliceous shale, consisting primarily of granular quartz with a little muscovite and biotite. Minute bubblelike pods of K-feldspar, however, can be seen scattered through the quartz grains, and apparently reflect incipient potash metasomatism. The decline in K-feldspar content away from the intrusive, the preservation of bedding in the feldspathic rocks, and the continuity from these to the hornfelses, indicate that K-feldspar was added to chert and siliceous shale near the granodiorite contact. The thin porcelaneous beds rich in apatite may be metamorphosed layers of sedimentary phosphate, though phosphate was not recognized where the strata are unmetamorphosed.

This is the only area of metasomatism found near the granodiorite, but hornfels is widespread. Chert has recrystallized to a granular mosaic of quartz with a few flakes of biotite or muscovite. Siliceous shale has been altered to granular quartz, biotite, and muscovite. Some of the shales apparently contained considerable organic carbon, for hexagonal flakes of graphite are present in some of the hornfels. Limy shales or sandstones that have been hornfelsed contain patches of actinolite, tremolite, or diopside; and calcite has recrystallized to form large-grain mosaics. Throughout the hornfels zones around the stock are many veinlets and seams of granular quartz.

JOINTING

Over the whole area of exposure, the granodiorite of Granite Mountain is broken into approximately rectangular blocks by two sets of almost vertical joints (fig. 12)

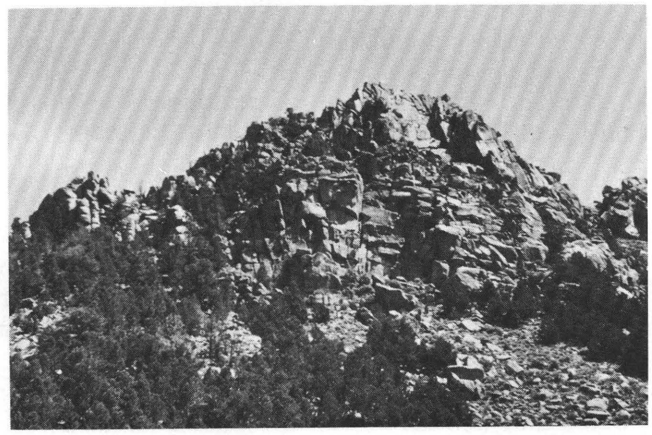


FIGURE 12.—Granite Mountain from the southwest, showing conspicuous jointing.

and a well-developed sheeting. The attitudes of about 300 joints, measured within 5° on a grid pattern, are plotted in figure 13. The sheeting strikes about N. 10°

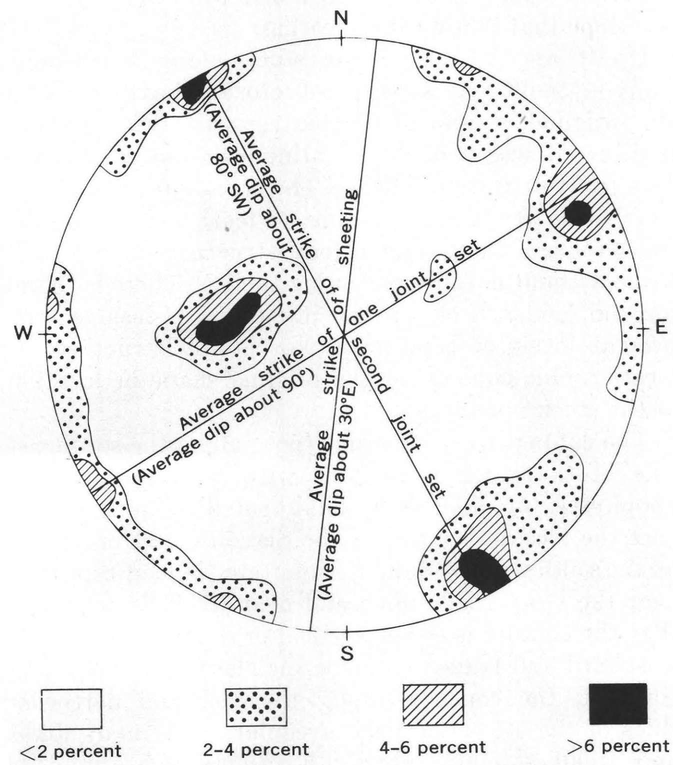


FIGURE 13.—Equal-area projection of poles to planes of jointing and sheeting in the Granite Mountain stock. Lower hemisphere, 300 poles.

E., approximately parallel to the east side of the stock, and dips about 30° E. One set of vertical joints strikes about N. 60° E., normal to the trend of the stock from

Hilltop toward Tenabo; the other set strikes N. 25° W., conforming to the trend of the stock, and dips about 80° W.

The two sets of joints and the sheeting mutually cut each other without offset, and thus all appear to be of the same age. The vertical joints, but not the sheeting, continue several hundred feet into hornfelsed wallrocks and fade as the hornfelsing decreases. Lamprophyre dikes in the granodiorite follow the vertical joints, but pegmatites bear no relation to the jointing. These facts suggest that the joints are related to cooling of the granodiorite and formed before or contemporaneous with injection of the lamprophyres. Neither foliation nor lineation is apparent so the jointing or sheeting cannot be shown to be related to the intrusion of the granodiorite magma.

PETROGRAPHY

Granodiorite

The stock consists primarily of unfoliated biotite-hornblende granodiorite of monzonitic texture, though facies ranging from quartz diorite to aplite in composition and from hypidiomorphic granular to porphyritic in texture are also present. Cognate inclusions are common, but inclusions of wallrock are scarce. The granodiorite differs from the subordinate facies in texture and proportions but not in the species of minerals. Quartz, plagioclase, K-feldspar, hornblende, and biotite are the common minerals, with magnetite, sphene, apatite, calcite, and allanite as accessories.

Granodiorite forms at least 70 percent of the exposed bulk of the stock. In outcrop it is white to light gray; black biotite and hornblende give a pepper-and-salt appearance. Megascopic minerals include quartz, plagioclase, hornblende, and biotite and the texture is equigranular. Near the margins of the stock, plagioclase forms phenocrysts in an equigranular groundmass of quartz, plagioclase, biotite, and hornblende.

The microscope shows anhedral quartz, subhedral zoned plagioclase, and subhedral to anhedral biotite and hornblende in a hypidiomorphic granular texture. The interstices are filled by anhedral K-feldspar and quartz. Most thin sections show slight deuteric alteration—a dusting of saussurite in some of the plagioclase crystals and a little chlorite bordering some crystals of biotite and hornblende. Accessory minerals, magnetite, sphene, apatite, calcite, and allanite, make up less than 1 percent of the rock.

Quartz-plagioclase contacts are sutured and irregular, and some quartz grains enclose small euhedral crystals of hornblende, biotite, or plagioclase. Many quartz grains are strained and contain dusty or bubbly inclusions, many of which appear to lie along healed cracks as the wisps and strings of inclusions cross contacts between grains.

The plagioclase is intergrown with quartz, hornblende, and biotite. Some crystals enclose small euhedral crystals of hornblende or biotite but rarely quartz. It invariably shows complex oscillatory zoning; contacts between the zones range from gradational to abrupt, with an unconformity between zones. The plagioclase has the average composition of andesine; although the zoning is oscillatory with many reversals, the composition ranges from about An_{55} in the centers to An_{25-35} on the outside. Commonly, a thin, ragged and discontinuous zone of oligoclase faces areas filled with interstitial K-feldspar. A wormy intergrowth of K-feldspar and oligoclase follows some of the oligoclase-K-feldspar contacts. Both the rim of oligoclase and the wormy intergrowth are absent from other mineral contacts.

Hornblende ranges from euhedral crystals to anhedral ragged splinters and appears typical with pleochroism: X=pale green to light bluish green, Y=dark green, Z=very dark green to dark bluish green, $2V=60°$ to 70°, negative, and $Z \wedge c$=about 20°. Biotite forms subhedral to anhedral, somewhat ragged books, it is pleochroic in green and brown to reddish brown, with a small optic angle.

Biotite and hornblende are present together in the granodiorite, in widely varying relative proportions. This variation was noted in the field and an attempt made to map it, but no recognizable pattern was found. Study of thin sections of specimens collected on a grid pattern also failed to indicate any systematic variation. Biotite is commonly within or alongside hornblende and vice versa. Both appear equally fresh and may be poikilitically enclosed in plagioclase or interstitial K-feldspar. Apparently both grew simultaneously throughout most of the period of crystallization.

K-feldspar averages about 15 percent by volume in the granodiorite, but is almost absent in the quartz diorite facies and makes up as much as 25 percent of the quartz monzonite facies. Within the granodiorite itself, it ranges from about 11 to about 20 percent, the greatest variation of any of the main constituents.

In the granodiorite, the K-feldspar occurs only as anhedral crystals filling interstices along with anhedral quartz. It commonly poikilitically includes euhedral to subhedral crystals of plagioclase, hornblende, and biotite, but rarely of quartz. The interstitial K-feldspar also embays and veins some of the adjacent plagioclase crystals. The embayments and veinlets of K-feldspar are optically continuous with the main mass of interstitial K-feldspar. The K-feldspar thus appears to have not only been the last main constituent to crystallize, but also to have reacted with and replaced plagioclase.

The K-feldspar has 2V ranging from 35° to 60°, generally 47° to 55°, and the optic plane is in the orthoclase position. Indices are 1.522, 1.526, and 1.528 and extinction on (010) is 6° to 7°. These properties suggest orthoclase cryptoperthite (Tuttle, 1952, p. 557 559, 563) with a composition of about Or_{75-65}, Ab_{25-35} (Larsen and others, 1938, p. 419; Edmondson Spencer, 1937, p. 457).

Where the K-feldspar forms crystals 1 or 2 mm long, it commonly shows very fine albite-hair perthite, oriented parallel to (010). There is no evidence of soda metasomatism so the perthite doubtless results from exsolution.

An average of eight modal analyses of granodiorite and granodiorite porphyry specimens gives the following percentages by volume: Quartz, 24; plagioclase, 47; K-feldspar, 14.5; biotite, 10; hornblende, 4.5. Included in the biotite and hornblende averages is a little chlorite which formed from slight deuteric alteration of these minerals. Accessory minerals comprise less than 1 percent of the total. Calcite, which occurs as anhedral crystals filling part of an interstitial area in a few fresh and unaltered samples, may be magmatic rather than hydrothermal.

An analysis of a granodiorite porphyry specimen from the Granite Mountain stock is given in column 7, table 7.

Quartz monzonite

Porphyritic quartz monzonite, as a facies of the Granite Mountain stock, is present in several small knobs surrounded by alluvium in Mud Spring Gluch and also on the northeast side of Bullion Mountain on the spur just northwest of Rock Spring. These small bodies are not distinguished from the dominant granodiorite on the map (pl. 1).

The transition from the granodiorite to porphyritic quartz monzonite takes place in a zone at least 300 feet wide in which masses of each enclosed the other. Irregular stringers of each rock type intrude the other, and locally there is much swirling and interlensing of the two. Veinlets, stringers, and irregular patches of aplite are particularly abundant. The structure of the contact zone resembles that of a marble cake.

The porphyritic quartz monzonite consists of the same minerals as the granodiorite, but contains abundant euhedral phenocrysts of K-feldspar, as much as 10 cm long and averaging about 4 cm. Zones of inclusions parallel to the faces of the phenocrysts are clearly visible with a hand lens. Quartz also forms euhedral phenocrysts as much as 2 mm in diameter. The groundmass resembles typical granodiorite, but locally shows some granulation.

Several of the large phenocrysts were studied in thin section. They consist of orthoclase cryptoperthite whose indices and optic angles are generally the same as those of the interstitial K-feldspar in the granodiorite. But instead of fine hair perthite, the phenocrysts contain irregular rectangular patches of albite alined rigorously parallel to the (010) cleavage of the phenocryst, the albite twinning being parallel to the cleavage. The optic angle of the K-feldspar in areas where the albite is relatively abundant is 53° and the extinction angle on (010) is 8°. Where albite is absent, the optic angle is 58° and the extinction angle 6°. These variations suggest a decrease in the albite content of the K-feldspar in the perthitic areas. This conclusion, the regularity of orientation of the albite, and the absence of deuteric or hydrothermal alteration indicate that this perthite, like the hair perthite of the granodiorite, is probably the result of exsolution rather than replacement.

The centers of the K-feldspar phenocrysts are crowded with randomly oriented crystals of quartz, plagioclase, hornblende, and biotite. The inclusions decrease in number outward, but they become progressively better oriented, so that biotite flakes and hornblende laths approximately parallel the pinacoidal faces of the phenocrysts. Plagioclase inclusions are even more rigorously oriented: their albite or pericline twin planes parallel the (010) or (100) faces of the phenocrysts within a range of only 5°.

As the crystals of plagioclase, biotite, and hornblende in the surrounding granodiorite are not so rigorously oriented, the only explanation for the orientation within the phenocrysts seems to be that the freely growing euhedral phenocrysts impinged on euhedral crystals of plagioclase, hornblende, and biotite, also growing in the magma. As a pinacoidal surface of K-feldspar grew outward, it gradually rotated smaller euhedral plagioclase phenocrysts until their crystal faces were parallel. The outward trend toward alinement of inclusions may reflect increasing euhedrism of the larger crystal faces of the K-feldspar relative to those of the enclosed crystals.

The K-feldspar phenocrysts not only enclosed plagioclase crystals but also extensively replaced them; many plagioclase inclusions are anhedral, with irregular, embayed, and scalloped outlines, cut by veins and patches of K-feldspar. Zoning is abruptly truncated, and rims of oligoclase are missing, although these are present on the plagioclase in the groundmass.

Quartz inclusions are also irregular and embayed and may have been replaced to some extent. Hornblende and biotite seem to have been replaced only slightly if at all.

The matrix of the quartz monzonite porphyry is granodiorite typical of most of the stock. In a few specimens, bent biotite flakes and strained quartz grains and K-feldspar phenocrysts suggest that the rock has been somewhat crushed.

Quartz diorite porphyry

Quartz diorite porphyry forms offshoots of the Granite Mountain stock in Mud Spring Gulch, the small crescent-shaped satellitic plug on the south slope of Bullion Mountain, and a probable extension of the stock in the Hilltop area. It also makes up the small plug near Tenabo, on the trend of the Granite Mountain stock.

The quartz diorite porphyry contains subhedral phenocrysts of oscillatively zoned plagioclase, some of rounded and embayed quartz, and of subhedral horn blende and biotite. These phenocrysts are in a fine- to medium-grained groundmass of quartz, plagioclase, and a very little K-feldspar. The quartz diorite porphyry facies thus resembles the main mass of granodiorite of Granite Mountain except for its strongly porphyritic texture and its low content of K-feldspar. In the Hilltop area and in Mud Spring Gulch, the plagioclase has been much saussuritized and the hornblende and biotite replaced by chlorite. At Tenabo small clots of green hornblende and of phlogopitic biotite are present in the quartz diorite porphyry.

An analysis of a specimen from a small intrusive near Havingdon Peak is given in column 6, table 7.

Inclusions

Three varieties of cognate inclusions were found within the Granite Mountain stock. The size, angularity, and content of quartz, biotite, and K-feldspar increase in order of decreasing age of the three types.

Clots of amphibolite are the oldest inclusions. These are scattered throughout the stock, both in the granodiorite and its facies and in the younger inclusions. They are spheroidal, average about 10 cm in length, and stand out conspicuously against the white granodiorite. A representative amphibolite inclusion has the following modal volume percentages: hornblende, 64; plagioclase, 18.5; quartz, 11; biotite, 6; and K-feldspar, 0.5. These are the same mineral species as in the surrounding rocks, even to the zoning of the andesine. There are no reaction rims between inclusions and host rocks. The textures are xenomorphic to hypidiomorphic granular; grains range from 0.1 to 1.5 mm in diameter. Although the plagioclase zoning might suggest that these inclusions are magmatic rather than recrystallized xenoliths, if reaction between magma and xenolith were complete, perhaps the zoning would reflect a like thermal history for both matrix and inclusion, as has been inferrred in the intrusions of the Colorado Plateau (Hunt, Averitt, and Miller, 1953, p. 160–164; Waters and Hunt, 1958, p. 349–351).

The next younger inclusions are of quartz diorite porphyry. Most range from 6 inches to 4 feet in diameter but a few are as much as 40 feet across. They tend to be rectangular with rounded corners. Conspicuous phenocrysts of white plagioclase and glassy quartz in a dark matrix of small hornblende and biotite crystals give the outcrops a highly porphyritic appearance. The following mode is representative: plagioclase, 42; hornblende, 40.5; quartz, 8; biotite, 7; K-feldspar, 2.5. Euhedral to subhedral phenocrysts of zoned andesine and anhedral to subhedral phenocrysts of quartz, both in the 1- to 3-mm size range, are in a matrix of smaller crystals of hornblende, biotite, plagioclase, and quartz, forming a hypidiomorphic granular texture. A little K-feldspar fills interstices between the other minerals. These quartz diorite inclusions themselves contain inclusions of amphibolite.

The youngest inclusions are large blocks of granodiorite porphyry, which contain inclusions of both amphibolite and quartz diorite. One is a block at least 100 feet long and 50 feet wide, at the north end of the stock near where the road crosses the divide between The Park and Mud Spring Gulch. Contact with the surrounding granodiorite is sharp and corners are angular. Dikes of granodiorite cut the block. Many inclusions of granodiorite porphyry also occur along the south side of the stock.

In hand specimens the granodiorite porphyry of these inclusions shows conspicuous subhedral to euhedral phenocrysts of white plagioclase 2 to 5 mm long in a matrix of quartz, plagioclase, hornblende, and biotite that has a dark pepper-and-salt appearance. In thin section, the following mode was measured in volume percent: plagioclase, 47; K-feldspar, 17; quartz, 17; biotite, 13; and hornblende, 6. All mineral species are the same as those in the host granodiorite. The K-feldspar appears to have been last to crystallize; in the matrix irregular anhedral crystals, some as much as 1 mm long, poikilitically enclose smaller euhedral to subhedral crystals of hornblende, biotite, plagioclase, and quartz. Some of the included crystals of zoned andesine appear to be embayed by the K-feldspar.

The inclusions of granodiorite porphyry differ from the host granodiorite, especially the porphyritic border facies, primarily in a higher percentage of biotite and hornblende. The large blocks of granodiorite porphyry along the border of the stock may not be inclusions, but remnants in place of a slightly older border facies that has been invaded by the granodiorite.

Inclusions of Paleozoic rocks are smaller and much less abundant than the cognate inclusions. Angular fragments of chert and quartzite, mostly ranging from 3 to 12 inches across, are very sparsely distributed throughout the stock, with no apparent concentration along the borders.

Chert inclusions have recrystallized to a granular mosaic of slightly dusty quartz having an average grain size of about 0.05 mm and containing subhedral crystals of hornblende from 0.01 to 0.05 mm in length. The quartzite inclusions consist of quartz grains that retain their original rounded shapes, quartz grains that have recrystallized with sutured contacts, and interstitial subhedral crystals of green hornblende and dusty K-feldspar. The hornblende and K-feldspar have the same optical properties as in the enclosing granodiorite. Though the granodiorite contains more biotite than hornblende, the inclusions contain no biotite. The inclusions of chert and quartzite have apparently undergone some metasomatism. The absence of biotite in them perhaps reflects failure of sufficient water to penetrate the inclusions.

Aplite

Most of the aplite associated with the Granite Mountain stock is in the broad contact zone between granodiorite and prophyritic quartz monzonite. It forms lenses, pods, and ramifying veinlets and narrow dikes rarely more than a foot wide. In many places the host rock grades to aplite through a zone of several inches marked by decreasing biotite and hornblende and a gradual change from coarse hypidiomorphic granular texture to fine-grained xenomorphic granular. Some aplites have streaks of slightly coarser grained quartz and feldspar, and many have vugs about an inch long lined with euhedral crystals of quartz.

The microscope shows that the aplite consists of an irregular mosaic of anhedral plagioclase, K-feldspar, and quartz whose average grain size is about 1 mm. Some plagioclase is oscillatively zoned andesine like that of the granodiorite but most is gradationally zoned saussuritized albite. K-feldspar and quartz are micrographically intergrown and in places occupy irregular masses and deep embayments in the albite. Some K-feldspar encloses small unconnected, fuzzy patches of albite all of which show parallel extinction and twinning, thereby suggesting replacement of albite by K-feldspar. The only ferromagnesian mineral is biotite; it constitutes less than 1 percent of the aplite.

The aplite dikes are much coarser grained next to their borders than in their centers. K-feldspar crystals as much as 1 mm long are riddled with angular masses and wormy stringers of quartz which give a very pronounced cuneiform micrographic texture. Plagioclase is absent. Some micrographic K-feldspar crystals are in optical continuity with K-feldspar of the quartz monzonite alongside.

As the granodiorite is so homogeneous, it is impossible to tell whether the aplite is a fracture filling or a replacement body, but the textures and minerals suggest that they formed by replacement, primarily by addition of K-feldspar and quartz and saussuritization of the andesine to give cloudy residual albite.

Lamprophyre

Lamprophyre dikes cut the Granite Mountain stock, particularly just east of the Corral Canyon fault. They clearly follow the two vertical joint sets. They range in thickness from about 2 feet to about 30 feet and sharply cut the granodiorite. These dikes weather more rapidly than the granodiorite and form linear depressions. Many have irregular brownish-red oxidized areas. They are dark gray to black and very fine grained or aphanitic. Irregular clots of calcite occur in a few, and one contains rounded grains of quartz surrounded by thin calcite rims.

The remnant texture of these dikes suggests that they once consisted of pyroxene (and perhaps hornblende) phenocrysts in a matrix of plagioclase laths. The phenocrysts are now a mass of chlorite, calcite, and chalcedony. The groundmass consists of andesine, chlorite, and calcite. Clearly the dikes have undergone extensive hydrothermal or pneumatolytic alteration, chloritization, and calcitization; the altering volatiles must have been rich in water and CO_2.

EMPLACEMENT

The Granite Mountain stock cuts across the structure of the Paleozoic rocks and no obvious control for its shape, size, orientation, or location is apparent.

The Bateman thrust must have joined the Trout Creek transcurrent fault in the area now occupied by the Granite Mountain stock. (See p. 113.) These two faults doubtless bound a major structural block on the south and on the bottom; their junction might thus present a zone unusually susceptible to intrusion. Perhaps the dikes that mark the Trout Creek fault for much of its length and the localization of the two small intrusive masses of quarts monzonite on the ridge west of Goat Peak at the west end of the Trout Creek fault are further indications of such a control. The fact is, however, that the Bateman fault in part of the area of the Granite Mountain stock lay at very shallow depths prior to the emplacement of the stock (sections *F–F′* and *11–11′*, pl. 2) and elsewhere at somewhat deeper levels (sections *E–E′* and *9–9′*, pl. 2), yet does not seem in any way to control the outlines of the intrusive mass (sections *E–E′*, *F–F′*, *G–G′*, *9–9′* and *10–10′*, pl. 2).

It is of course conceded that the outlines of the stock may really differ considerably from those shown on these sections; the position of the faults, though, should be reasonably correct. The independence of the faults and the main outline of the intrusion at the surface testifies to their independence at depth also.

The problem of emplacement and room is most acute for the comparatively large Granite Mountain stock. Even granting that this mass may not widen downward to the extent we suggest in the section of plate 2, at the surface it is 4½ miles long and about a mile in maximum width, yet it does not deflect such thrust sheets as those of the Bateman and Hilltop thrusts near the northwest end of the stock. The Bateman thrust is cut off at an oblique angle with the strike but at a very steep contact north of the head of Hilltop Canyon; neither it nor the rocks of footwall or hanging wall show any deflection from their regional trends. Similarly, the many dikes between the main forks of Hilltop and Bateman Canyons testify to a larger mass at depth, yet the Hilltop thrust above and the fault slices associated with it show no deviation from their regional trend.

The Slaven Canyon fault west of O'Haras Peak is dipping into the intrusive and certainly shows no tendency to parallel the contact. The belt of Slaven Chert surrounding much of Granite Mountain is not wider in the area of the intrusive than it is to the north in Slaven Canyon. On Granite Mountain itself the steep intrusive contact can be followed through a relief of more than 1,000 feet. Neither to north nor south do the attitudes of the beds of the Slaven Chert show any deflection by the intrusive from their general trend.

There is therefore no evidence of forceful displacement of the wallrocks during the emplacement of the stock, nor the slightest suggestion of a laccolithic form. Perhaps, though, this lack of evidence of forceful intrusion and shouldering aside of wallrocks does not negate such a mechanism so completely here as it would in areas of simpler structure.

The cherts, quartzites, and siliceous shales of the Shoshone Range are much fractured and jointed. Presumably during an intrusion the rocks alongside would be readily broken and would not transmit an uplifting force far from the contacts. The disturbing effect of the intrusive might thereby by considerably more localized than it would be in areas of less broken, more homogeneous, and more persistent stratigraphic units. A floored intrusive, if such there were, would not here tend to develop a laccolithic form so much as a bysmalithic one. However, we have no evidence of a floor in any case.

It is doubtful, however, whether a stoping mechanism could have operated here. No large blocks of Paleozoic rocks and few small ones were found in the Granite Mountain stock. The breccia alongside the small plug on Bullion Mountain has not foundered into the underlying quartz diorite. Inclusions of amphibolite and diorite are common in the Granite Mountain stock and presumably were suspended as the magma crystallized. Chert and quartzite probably would not sink in a magma viscous enough to hold amphibolite inclusions in suspension.

The alternative is that the displaced wallrocks were removed ahead of the rising magma, and eventually reached the surface as inclusions in volcanic breccias, block agglomerates, tuffs, and lavas. The rising quartz monzonite magma in the Horse Canyon pipe (p. 66–70) was preceded by a cap of breccia. The early magmas in all three breccia pipes of the Miocene and Pliocene volcanic cycle and in the cores of the dacite and rhyolite plugs on the southeast flank of Mount Lewis carry abundant fragments of the wallrocks, whereas later magmas of this cycle were free of inclusions. This suggests that the upper parts of the magma columns were richer in inclusions than the lower. Furthermore, the welded tuffs, tuff breccias, and black agglomerates of the upper Tertiary volcanic rocks average about 3 percent total volume of chips and fragments of Paleozoic rocks. Perhaps this was also true during the earlier igneous episode to which this stock belongs.

Although this hypothesis may account for the space occupied by the intrusives, there still remains the question of the mechanism of intrusion. Within the Shoshone Range are small late Tertiary dikes of breccia, complex breccia pipes, small plugs with breccia along their contracts or containing it in offshoots, and the comparatively large early Tertiary Granite Mountain stock whose contacts with Paleozoic sedimentary rocks are sharp, crosscutting, and unbrecciated. The magmas in all these are clearly much alike. The swarm of dikes and the hornfelsing near the breccia pipes suggest that these may be underlain by a stock as large as the Granite Mountain stock. Whether or not the Granite Mountain stock was capped by a breccia pipe or volcano depends on interpretation of rather circumstantial petrographic evidence discussed in the next section. Although no transition from a stock to a breccia pipe can be seen, the several varieties of intrusive bodies and breccias may represent stages at successive depths in the crust of the emplacement of a younger stock like that of Granite Mountain. The mechanism would be brecciation by eruptive activity, "volcanic pumping," and rock bursting above the rising magma, as postulated for the breccia pipes. (See p. 73–75.)

PETROGENESIS

The granodiorite of Granite Mountain is essentially a hornblende-biotite quartz diorite with interstitial granitic material. Zoned plagioclase, quartz, hornblende, and biotite crystallized contemporaneously until the magma was a mush of crystals, at which time the residual liquid crystallized to quartz, K-feldspar, and a little oligoclase. In the prophyritic quartz monzonite facies, K-feldspar phenocrysts crystallized during a slightly earlier stage than did those in the granodiorite. Crystallization began in the porphyritic dike rocks and in the extrusive rocks as it did in the granodiorite but was arrested before the granitic fraction crystallized. Phenocrysts of zoned andesine, quartz, hornblende, and biotite are surrounded by a glassy groundmass which contains normative orthoclase and more normative than modal quartz.

There is no textural evidence that biotite formed by reaction of hornblende with the magma; nor any correlation between amounts of biotite and hornblende as would be expected if biotite replaced hornblende. Both minerals apparently crystallized simultaneously. Hydrothermal experiments on the stability of pargasite (Boyd, 1956, p. 199) and phlogopite (Yoder and Eugster, 1954, p. 163) suggest that biotite and hornblende may both be stable at the same temperature and water pressure. The two breakdown curves, above a water-vapor pressure of 1,000 bars and a temperature of 1050°C, are very close together. These curves represent maximum temperatures for end members of two solid-solution series. Addition of other components, such as quartz, would lower the temperatures; and iron in solid solution in the micas and amphiboles would not only lower temperatures further but would also produce changes owing to pressure of oxygen (Eugster, 1956.) Although the breakdown curves thus must be used with caution, they do suggest that simultaneous crystallization of biotite and hornblende might occur if water-vapor pressure were sufficient. In the Granite Mountain stock, the amount of biotite shows a roughly linear increase with increasing amounts of K-feldspar. This suggests that the amount of potassa in the magma is the primary control for the proportion of biotite relative to hornblend

The interstitial K-feldspar, quartz, and a little ologoclase have essentially the composition of a granite whose composition is near the thermal trough in the $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2 system at 1,000 bars of water-vapor pressure (Tuttle and Bowen, 1952, p. 39; see fig. 14, this report).

The increases in proportions of quartz, K-feldspar, and biotite of the cognate inclusions with decreasing age, indicate that the granitic components in the magma gradually increased with successive intrusions. If the granitic interstitial filling resulted solely from fractionation, the replacement of plagioclase by K-feldspar becomes difficult to explain. Plagioclase ranging from oligoclase to calcic andesine, was breaking down at the same time that K-feldspar was crystallizing from the residual magma. Gillson described (1929) very similar replacement in a quartz monzonite from Pioche, Nev. There the replacement was much more extensive than in the Shoshone Range, but the general textural relations and the composition, optic angle, and perthitic character of the replacing K-feldspar closely resemble those in the Granite Mountain stock. Gillson concluded that much of the replacement was deuteric after complete consolidation of the magma but also stated that the introduction of K-feldspar by gaseous transfer must have begun in a late magmatic stage.

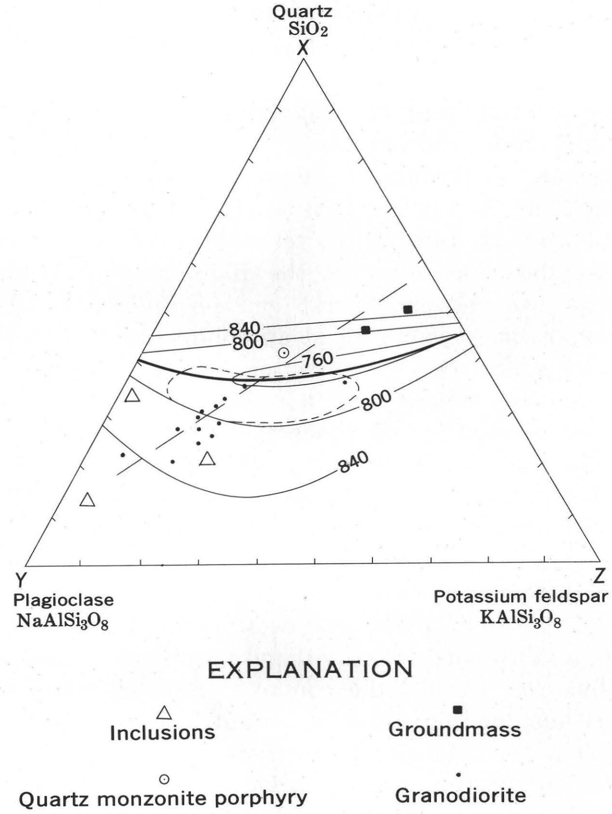


FIGURE 14.—Plot of modes of rocks from the Granite Mountain stock, superimposed upon crystallization diagram of mixtures of albite, K-feldspar, and quartz under 1,000 atm pressure of water vapor. (From Tuttle and Bowen, 1952, p. 39.) On the modal plot, *X* is quartz, *Y* is plagioclase, and *Z* is K-feldspar. On the crystallization diagram, *X* is SiO_2, *Y* is albite, and *Z* is K-feldspar. The isotherms are on the liquidus and the heavy line separates the fields of quartz and feldspar. Line of short dashes encloses compositions of many granites measured by Chayes (1956, p. 213). Line of long dashes shows general trend of composition change in the Granite Mountain stock.

In the Shoshone Range, the replacement of plagioclase by K-feldspar was late magmatic, as is shown by several lines of evidence: (1) The granodiorite of Granite Mountain is fresh. Deuteric alteration was confined to slight saussuritization of the plagioclase and chloritization of the mafic minerals. (2) The replacing feldspar is generally optically continuous with that in the interstices. (3) The exsolution perthite indicates crystallization above the subsolidus exsolution curve in the $KAlSi_3O_8$-$NaAlSi_3O_8$-H_2O system (Bowen and Tuttle, 1950, p. 497). (4) Alinement of the crystals poikilitically included in the K-feldspar phenocrysts implies growth of the phenocrysts in a magma.

Bowen and Tuttle (1950, p. 497–500) showed that an increase in water-vapor pressure lowers the temperatures of crystallization of the orthoclase-albite solid solution series and stated that with sufficient pressure the solidus should intersect the subsolidus solvus. This intersection would produce either two solid solution series with a reaction relation or a eutectic between the two; which is produced depends on the way in which solidus and solvus meet. The replacement of plagioclase by K-feldspar in the Granite Mountain stock can hardly be explained this way The K-feldspar of the exsolution perthite must have crystallized as a solid solution. Furthermore, Tuttle and Bowen (1952, p. 38) reported that a water-vapor pressure of about 4,000 atm is required to lower the solidus enough to intersect the solvus, even with the addition of silica as another component to help lower the temperature. This pressure corresponds to a depth of about 9 miles; it is very doubtful whether the Granite Mountain stock crystallized at any such depth. A much more likely estimate would be 2 or 3 miles.

The anorthite-leucite-silica system (Schairer and Bowen, 1947, p. 75, 80) contains a eutectic between K-feldspar and anorthite if leucite is neglected. The albite in the granodiorite of Granite Mountain constitutes a third component in the system, and Schairer and Bowen (1947, p. 87) warn that this may drastically change the paragenesis. Certainly most granodiorites and quartz monzonites contain both K-feldspar and plagioclase feldspars of oligoclase to andesine composition without any petrographic evidence of disequilibrium. In the Shoshone Range, the quartz latite welded tuff from the ridge south of Mill Creek contains both discrete euhedral K-feldspar and zoned andesine phenocrysts. Some boundary must exist between the alkali feldspar and the plagioclase solid solution series, perhaps in the general position indicated by Bowen (1928, p. 231). Bowen (1928, p. 227, 233) postulated that a reaction relation between K-feldspar and plagioclase might produce K-rich magmas, but subsequent work (Schairer and Bowen, 1947) showed no experimental evidence for such a reaction.

Nonetheless, reaction clearly took place in the granodiorite of Granite Mountain. K-feldspar would crystallize and plagioclase simultaneously break down if the course of crystallization of the magma were displaced onto the feldspar solidus from a boundary curve such as Bowen proposed. This would require addition to the magma of the constituents of K-feldspar, an increase in temperature, a change in pressure of volatiles, or a combination of all three. The diagrams of Bowen and Tuttle (1950, p. 497) show that an increase in water-vapor pressure moves the minimum in the alkali solid solution series slightly toward albite. An increase in water-vapor pressure might thus shift the boundary between the alkali and plagioclase feldspars toward plagioclase. If this took place rapidly, a temporary disequilibrium might cause melting of the plagioclase and rapid crystallization of K-feldspar.

If the Granite Mountain stock underlay an explosive volcano, gases might stream upward through the crystallizing magma during an eruption, bringing heat, potassium, and abundant volatiles to add to the residual magma, thus causing disequilibrium of the plagioclase and rapid crystallization of K-feldspar as suggested in the preceding paragraph. A similar change in the Cornelia Quartz Monzonite at Ajo, Ariz., has been interpreted in the same way (Gilluly, 1946, p. 29, 73–81). Perhaps the calcium of the irregular patches of interstitial calcite was derived from the replaced plagioclase.

This hypothesis would account for the quartz monzonite porphyry facies of the granodiorite of Granite Mountain. The large K-feldspar phenocrysts suggest both abundant potassium and slow cooling. Upward-streaming gases might have caused enough turbulence to slightly crush and granulate the granodioritic matrix and produce the swirled irregular "marble-cake" contact zone between porphyry and granodiorite. The aplite pods and dikes in this zone may indicate a still later continuation of this gaseous introduction of K-feldspar after essentially complete crystallization. The locale of the quartz monzonite porphyry facies within the stock may thus represent the main eruptive channel through the stock to a volcano above.

The rocks of the Granite Mountain stock provide little evidence concerning the origin of the magma itself. Perhaps the scattered amphibolite inclusions are a clue that selective fusion of amphibolite at depth gave rise to the magma. Similar inclusions in the laccoliths of the LaSal Mountains, Utah, led Waters and Hunt (in Hunt, 1958) to propose selective fusion of amphibolite as a magma-generating mechanism. The main magmatic sequence in the LaSal Mountains is from diorite porphyry to feldspathoidal syenite, accompanied by a final stage of aegerine granite and soda-rhyolite porphyry derived from alteration of syenitic residual magma by volcanic gases. In the Granite Mountain stock and satellitic bodies, amphibolite inclusions are less abundant than in the LaSal Mountains; the rocks contain considerable quartz; and although the mechanism proposed for making the aegerine granite and rhyolite in the LaSal Mountains is similar to that proposed here for the reaction between K-feldspar and plagioclase, there is no evidence that the residual magma was syenitic.

STOCK ON RIDGE WEST OF GOAT PEAK

A small stock of porphyritic granodiorite split by a septum of Roberts Mountains Limestone cuts the carbonate rocks of the Goat window on the ridge crest west of Goat Peak. The irregular borders are steep to vertical and locally the carbonate rocks have been recrystallized to diopside, garnet, and tremolite.

The fact that the two masses, which certainly must coalesce at shallow depths, crop out near the center of the upfold in the Roberts thrust (see section *H–H′*, pl. 2) might suggest that they are part of a deroofed laccolith. This location is, however, independent of the upfold in the thrust, for the upfold took place simultaneously with the thrust faulting (see p. 105–109) in Paleozoic time, and had attained essentially its present form long before the intrusion took place. The location of the intrusion on the axis of the upfold is thus no more significant of a genetic connection between folding and intrusion than is the location of the Horse Canyon (section *5–5′*, pl. 2) and Mount Lewis (sections *F–F′* and *7–7′*, pl. 2) intrusive bodies in steep synforms of the thrust sheets. As has already been noted, the Granite Mountain stock cuts through both antiform (sections *12–12′* and *11–11′*, pl. 2) and synform indifferently (sections *10–10′* and *9–9′*, pl. 2). No relation is apparent between the folding and the much later igneous intrusions.

The granodiorite of this stock resembles that of Granite Mountain in mineralogy and texture except for the absence of hornblende. Euhedral to subhedral oscillatively zoned andesine, anhedral quartz, and subhedral biotite are in a matrix of granular quartz and K-feldspar. Some of the interstitial K-feldspar encloses plagioclase and biotite crystals.

AGE

The petrographic similarity of the stock west of Goat Peak to the Granite Mountain stock suggests that they are of the same age. The Granite Mountain stock is surely of early Tertiary age. Not only have zircons from this stock yielded a lead alpha age of 50± 10 million years (T. W. Stern, written commun., 1962), which is compatible with an Eocene or Oligocene age, but boulders derived from it are contained in gravels beneath the pyroclastics of middle or late Tertiary age in Indian Creek valley.

This assignment cannot be confidently made with respect to all similar intrusive rocks, however, because in this area the same species of oscillatively zoned andesine, green hornblende, and biotite are found in the granodiorite as in the younger extrusives. The similarity even extends to the presence of a little allanite in both the granodiorite and the younger volcanics. Lindgren, Graton, and Gordon (1910, p. 46) commented on a similar parallelism in New Mexico: "there is a striking similarity in composition between the generally uniform Tertiary monzonitic intrusive rocks and the latites and latite-andesites of the effusive epoch. The suggestion is justified that they were derived from essentially the same source * * *."

ROCKS OF MIDDLE OR LATE TERTIARY AGE

EXTRUSIVE ROCKS

GENERAL FEATURES

The extrusive rocks exposed in the Mount Lewis and Crescent Valley quadrangles are remnants of a sequence of middle or upper Tertiary volcanic rocks which probably covered most, if not all, of the Paleozoic and lower Mesozoic rocks now making up the northern part of the Shoshone Range to a very considerable depth. On the west side of the range, downfaulted volcanic rocks crop out on the basin side of several of the bordering Basin Range faults. No reasonable structural assumptions could justify an inferred thickness of less than 1,500 feet for the volcanics northeast of the mouth of Lewis Canyon, yet volcanics are absent for several miles from the upthrown block alongside. In the south and southwest parts of the Mount Lewis quadrangle, some of the extrusive rocks are faulted against Paleozoic rocks and some unconformably overlap them.

On the east side of the range, a few patches of extrusive rock rest unconformably on Paleozoic sedimentary rocks and in turn are covered by Quaternary alluvium. Within the range, several hundred feet of coarse volcanic breccia and tuff is preserved in a fault block at the headwaters of Indian Creek.

The lithology of the extrusive rocks differs from place to place within the two quadrangles. In order to keep the number of map units within reasonable limits, many separate and diverse bodies of rock have been included under the rubric "Volcanic rocks, Tv" on the geologic map (pl. 1). This designation includes tuff, bedded tuff, mudflows, and lacustrine and stream deposits rich in pyroclastic material. Not all these rock types are of course present in every area marked with this symbol. The symbol Ts on the geologic map (pl. 1) indicates mappable conglomerate and sandstone relatively free of pyroclastic material.

INDIAN CREEK VALLEY, MOUNT LEWIS QUADRANGLE

Tertiary sedimentary (Ts) and supracrustal volcanic rocks (Tv) cover an area of about 4 square miles near the head of Indian Creek and on the divide between Indian Creek and Crum Canyon. They occur in a small fault block which has been downdropped along two normal faults on the west side of the valley. Surrounding ridges and peaks of Paleozoic rocks are as much as 1,400 feet above the highest outcrops of Tertiary rocks. As the fault displacement is probably not more than a few hundred feet, the sediments must have been deposited in a deep valley.

SEDIMENTARY ROCKS

The rocks mapped as Tertiary sedimentary rocks (Ts) in Indian Creek valley consist of coarse, pebble to cobble conglomerate and a few lenses of dirty coal. These rocks rest unconformably on Paleozoic rocks and are overlain by Tertiary volcanics. The coal lenses crop out along the normal faults to the west. Though the coal is of very poor quality, it was mined for fuel in the silver-boom days and some dumps still remain.

The conglomerate consists of well-rounded pebbles and cobbles in a sandy matrix. Bedding is very crude and locally absent, and local lenses of coarse sandstone are common. Most pebbles are of Paleozoic rocks but a few are of granodiorite, probably from the Granite Mountain stock. Fragments of Tertiary extrusive and fine-grained intrusive rocks are absent. The coal is lignitic and contains much fragmental plant material too much shredded for identification.

This lower conglomerate member is lenticular and discontinuous. It is overlain by a dominantly volcanic sequence, described below, and this in turn is overlain by an upper member of conglomerate, sandstone, and siltstone.

Coarse, pebble to cobble conglomerate fills channels cut into the block agglomerate and tuff. The pebbles and cobbles are subangular to well rounded and are of Tertiary volcanic and Paleozoic sedimentary rocks. Silicified logs are present in some sandy lenses. Torrential crossbedding, which dips more steeply eastward than the layering, and the lithology of the volcanic rocks, which matches closely many in the vents near Mount Lewis, both suggest a westward source.

Sandstone and grit form lenses and streaks within the block agglomerate, tuff, and conglomerate. Grains of Tertiary rocks are subangular and those of Paleozoic rocks generally round. Feldspar, quartz, and biotite are common. Crossbedding, cut-and-fill structures, and lensing of beds abound. Leaf and twig fragments are numerous.

The siltstone is generally black to dark gray, well-bedded, and full of plant fragments, mostly grasses and rushes. The microscope shows abundant very finely divided ash mixed with organic material, shreds of biotite and chlorite, and small grains of feldspar and quartz.

VOLCANIC ROCKS

The volcanics (Tv) of Indian Creek valley consist dominantly of coarse block agglomerate—probably largely volcanic mudflow or avalanche deposits. Other rock types interbedded with the block agglomerate include lapilli tuff, tuffaceous sandstone and siltstone, black organic mudstone, and coarse conglomerate of mixed Paleozoic and Tertiary fragments. These rocks overlie the basal conglomerate and coal beds and are overlain unconformably by Quaternary gravel. Maximum thickness of the volcanic rocks is about 500 feet. As the upper surface is erosional, the thickness was originally greater and, in view of the wide distribution of such volcanic rocks and the considerable relief of the surface on which they lie, it may have been very much greater.

Block agglomerate also crops out on the small knob (alt 8,283 ft) on the northeast spur of Mount Lewis. Here the agglomerate rests on a small pod of coal and overlaps directly, without intervening conglomerate, onto Paleozoic rocks. Several bodies of block agglomerate, too small to map, also rest on Paleozoic rocks near the Dean mine. These, however, may be landslide deposits of post-Tertiary age.

The stratigraphy is highly variable because of lensing and wedging out of beds, channeling of coarse stream conglomerates, torrential crossbedding, and local erosional discontinuities.

Block agglomerate

The block agglomerate weathers to rounded knobs; the unit slumps readily on steep slopes. It is dark purple to gray where fresh and weathers dark tan. It is generally massive, but locally shows faint very crude bedding. Small pods, generally not more than 4 to 5 feet thick and 10 to 50 feet long, of much contorted black tuffaceous mudstone and siltstone full of plant remains, mostly grasses and rushes, are scattered throughout. In a few places, the mudstone is bleached and baked along contacts with the agglomerate.

The block agglomerate consists of fragments of hornblende dacite ranging from a few inches to 6 feet across. Large blocks are well rounded and nearly spherical; small blocks are subangular to subround. Small chips of dacite, broken crystals of hornblende, biotite, quartz, and feldspar, and grains and pebbles of chert, quartzite, greenstone, and granodiorite make up the matrix. Some of the fragments of Paleozoic rocks are well-rounded pebbles and probably were incorporated into the agglomerate from stream gravels. Pieces of wood, some silicified and some charred, and small carbonaceous plant fragments are scattered throughout the block agglomerate.

In thin section the dacite blocks are seen to be much altered; with phenocrysts of plagioclase, hornblende, biotite, and quartz set in a divitrified matrix containing minute microlites of feldspar. The original euhedral plagioclase phenocrysts are oscillatively zoned, but have been altered to dusty, splotchy masses of minute grains of calcite, sericite, epidote, chlorite, and opaque minerals. The composition of the remaining plagioclase ranges from calcic albite to sodic oligoclase, depending on the degree of saussuritization. The original hornblende phenocrysts have been altered to an intimate mixture of chlorite and a yellow chloritic or clay mineral (nontronite?) having a higher virefringence than most chlorite. A fine dust of leucoxene and hematite rims many of the pseudomorphs after hornblende. Biotite shows similar alteration except that some books contain small anhedral crystals of K-feldspar.

The groundmass of the dacite fragments has been completely devitrified to a fine-grained patchwork of quartz, feldspar, chlorite, and opaque dust. A few fragments show perlitic fractures now filled with chlorite. Small microlites of hornblende and biotite are completely opaque.

The microscope shows that the matrix of the block agglomerate consists primarily of small fragments of dacite and broken phenocryst minerals. The matrix has been altered like the dacite blocks except that it has more abundant calcite. Mixed with the fragments of dacite are a few chips of chert and quartzite.

The block agglomerate on knob 8283 consists wholly of hornblende dacite in large blocks ranging in length from 2 to 8 feet and separated by a mortar of dacite fragments. Most blocks are crudely rectangular with rounded corners. The entire outcrop is an unbedded jumble of large and small blocks which are pale green to purple, in contrast with the brownish red of the matrix.

In thin section, this agglomerate resembles that in the valley except that some of the phenocrysts of andesine and hornblende remain unaltered. The conspicuous red color of the matrix attests abundant interstitial hematite dust.

An analysis of a representative specimen is recorded in column 5, table 7.

Tuff

The second most abundant rock type within Indian Creek valley is vitric crystal tuff and tuff breccia ranging from dacite to rhyolite. The tuff forms lenses from 10 to 100 feet thick in the block agglomerate. In places it is underlain and overlain by tuffaceous fine-grained bedded siltstone and mudstone.

In outcrop, the tuff is pale olive green to white and has a platy parting that from a distance resembles bedding. Small crystals of feldspar, quartz, and biotite, a few fragments of collapsed pumice, and scattered inclusions of Paleozoic rocks can be seen in hand specimen. Biotite flakes lie approximately parallel to the platy parting.

The microscope shows crystals, many broken, of quartz, plagioclase, sanidine, and biotite in a matrix of glass shards. Quartz, where unbroken, has the rounded embayed outline typical of quartz phenocrysts in the intermediate rocks of the area. Sanidine is clear and untwinned and plagioclase is oscillatively zoned andesine. Biotite is in small flakes and shreds, many of which are bent.

The matrix consists partly of glass shards, many recrystallized to minute wedges of twinned tridymite. The shards are not closely packed and fused together as in welded tuffs but are cemented by devitrified glass whose index of refraction is higher than that of the shards. This devitrified cement contains shreds of chlorite and opaque materials and in a few places shows very dim outlines of glass fragments. Probably it was originally very finely divided volcanic dust which, through solution, redeposition, and recrystallization, has cemented the shards. An analysis is given in column 4, table 7.

Origin

The hornblende dacite block agglomerate probably came from the dacite plug on the southeast spur of Mount Lewis. The minerals and texture of the dacite in the plug resemble closely those of the block agglomerate, and the plug has a central core of dacite breccia. The dacite probably first erupted as a breccia which avalanched downhill, rounding the blocks and incorporating vegetation, pebbles and cobbles from underlying gravels, and shard-filled mud from small ponds. The charring of wood fragments, the oxidation of hornblende and biotite, the alteration of the plagioclase, and the baked and bleached contacts with some of the mudstone lenses all suggest that the avalanche was hot and wet—a steaming, churning mass of dacite blocks, gravel, mud, and plant remains.

Tanakadate (1927, p. 158–164) gave graphic descriptions of volcanic mudflows during the eruption of Tokati-Dake in Japan, and Scrivenor (1929) described the lahars of the East Indies as volcanic mudflows. We believe their origin to be very similar to that of the material in Indian Creek. Many of the lenses of organic siltstone may represent small ponds formed on the irregular surfaces of the mudflows.

Avalanches of volcanic mudflows alternated with deposition of silicic tuff and with erosion and deposition of gravel by torrential, heavily laden streams. Many of the well-rounded pebbles and cobbles of Paleozoic rocks in the basal coarse conglomerate and in the channel fills may have come from erosion of prevolcanic gravels.

MILL CREEK-HARRY CREEK-COOKS CREEK AREA

In the southern part of the Mount Lewis quadrangle, a wedge of Tertiary sedimentary and volcanic rocks that thickens rapidly to the south overlaps Paleozoic bedrock. Conglomerate and tuffaceous sandstone, welded tuff, bedded tuff, and basalt flows have buried a topography of considerable relief cut in the Paleozoic rocks. The northern thin edge of the wedge is a cuesta of welded tuff forming a ridge south of Mill Creek, and the southern thick part is a section of pyroclastic rocks and lava flows several thousand feet thick that extends many miles to the south of the quadrangle. In the area between Mill Creek and Wilson Pass, the Tertiary rocks are irregularly distributed, partly because of faulting and tilting and partly because they fill canyons in the underlying bedrock.

SEDIMENTARY ROCKS

The Tertiary sedimentary rocks (Ts) mapped in the Cooks Creek and Harry Creek areas form a thin discontinuous basal conglomerate beneath the volcanics and sporadic interbeds within them. The conglomerate thickens in the valleys and thins or is absent on the hills and knobs of the erosion surface cut in the Paleozoic rocks beneath. Northeast of the high knob covered by welded tuff in sec. 9, T. 29 N., R. 45 E., gravels fill a valley about 250 feet deep cut in Tertiary basalt and the Valmy Formation. Here the gravel is obviously younger than the basalt but older than the welded tuff.

Where the conglomerate rests on Paleozoic rocks it consists wholly of chips, pebbles, and cobbles of chert, quartzite, shale, and greenstone. Rock fragments are subangular to subround. Stream bedding is locally exposed in valleys of the older surface, but on the hills the equivalent stratum is commonly a scree of small angular chips of the underlying bedrock.

VOLCANIC ROCKS

Distribution and stratigraphic relations

The volcanic rocks of the Cooks Creek-Harry Creek area include welded tuff, vitric-crystal tuff, and well-bedded tuffaceous stream and lake deposits. They are exposed along the ridge between Mill and Harry Creeks; in the southwest corner of the Mount Lewis quadrangle in the valley (known locally as Red Rock Canyon) of the stream, that flows west out of the quadrangle at a point about a mile north of its southwest corner; in the area south of Horse Mountain; and at the head of Cooks Creek. This discontinuous, patchy distribution reflects a combination of postvolcanic faulting, tilting, and erosion and of relief of the erosion surface on which the volcanic rocks and underlying patches of gravel rest.

The relief of the prevolcanic erosion surface is at least 800 feet. East of Jacks Canyon, welded tuff resting on a thin scree of gravel descends to within 80 feet of the present level of Mill Creek, 800 feet below the base of the welded tuff along the ridge crest south of Mill Creek. Welded tuff, basalt, and vitric-crystal tuff are exposed in the bottom of Harry Creek valley both near the head and mouth. Along the ridge between Mill and Harry Creeks, the base of the welded tuff ranges in altitude from 6,000 to 7,500 feet, primarily owing to the underlying topography rather than to faulting or warping.

In the canyon in the southwest corner of the Mount Lewis quadrangle, volcanic rocks make up a continuous east-dipping sequence. Reconnaissance south and west of the map area indicates that these rocks are part of a sequence at least 3,000 feet thick that extends many miles to the south. In the map area, the rocks comprising the upper part of this section consist of well-bedded tuffaceous lake deposits, stream gravels, bentonitic clays, and fine vitric tuffs. To the west of the map area this sequence is underlain by at least 1,000 feet of vitric tuff, volcanic breccia, basaltic and andesitic flows, and tuffaceous conglomerate and sandstone, whose base is not exposed nearby.

The rocks in the southwest corner of the Mount Lewis quadrangle are separated from those of the Harry Creek-Cooks Creek area by several normal faults which have dropped them down an unknown amount, so that no direct correlations can be made with the rocks of the other areas. The sequence of coarse volcanic rocks to the east, however, contains many beds of gravel made up of pebbles of quartzite, chert, and greenstone. It thereby resembles the volcanics that overlap the Paleozoic rocks north of the normal faults more than it does the finer grained volcanics in the southwest corner of the Mount Lewis quadrangle. It is probable, therefore, that before the faulting, the volcanic rocks encroached northward on an upland of Paleozoic rocks to form a wedge which thickened southward.

Welded tuff

The best exposures of welded tuff are along the ridge south of Mill Creek, where from 50 to 300 feet of tuff is exposed. South of Wilson Pass a fault block of welded tuff, tilted 45° E., is at least 1,500 feet thick. Here the welded tuff is overlain by vitric crystal tuff, water-laid tuff, and thin lenses of welded tuff.

In outcrop, the welded tuff is light gray, green, pale purple, or pink. It most striking characteristic is the close packing of crystals of quartz, feldspar, and biotite which may make up as much as 70 percent of the rock. The quartz grains are smoky gray and the sanidine crystals give brilliant cleavage reflections. Along with the mineral grains are fragments of collapsed, flattened pumice, and scattered chips of chert, quartzite, and greenstone, which are most abundant near the base.

Most of the welded tuff is structureless, but locally there is a vague suggestion of a bedding, a wavy irregular banding parallel to the base, and a faint lamination, owing to alinement of biotite flakes and collapsed pumice fragments. Jointing normal to the base is common but not regular enough to form polygons.

Thin sections show that the welded tuff is a quartz latite. Euhedral to fragmentary crystals of quartz, plagioclase, sanidine, biotite, and a little hornblende are crowded together in a glassy devitrified matrix showing faint outlines of glass shards. Crystals and crystal fragments constitute 50–70 percent of the rock; among these, quartz averages about 40 percent, plagioclase 30 percent, sanidine 20 percent, and biotite with a little hornblende, about 10 percent.

Unbroken quartz grains have rounded and embayed outlines, are generally clear, and commonly show strain. Plagioclase is clear and unaltered and shows marked oscillatory zoning from about An_{50} to about An_{25}. The sanidine likewise is unaltered; its optic angle of about 25° in the orthoclase position, and indices of $\alpha=1.520$, $\beta=1.525$, and $\gamma=1.526$, indicate a composition of about 20 percent albite and 80 percent orthoclase. Biotite forms ragged books and shreds; many are bent. It is dark reddish brown, commonly rimmed by fine magnetite dust, and some is altered to chlorite. The sparse hornblende grains are dark reddish brown, with magnetite dust, and low extinction angles.

The groundmass consists of collapsed pumice fragments, glass shards, and minute crystal fragments in a matrix so much devitrified that shards are only faintly visible under oblique light. Matrix, shards, and pumice have recrystallized to form patches of fine-grained aggregates of quartz and feldspar in a matrix of chalcedony, opal, and dusty almost opaque glass. Spherulitic growths are common. Despite the intense devitrification of the groundmass, feldspar crystals are unaltered.

An analysis of the welded tuff is given in column 10, table 7; a photomicrograph of one of the freshest specimens is shown as figure 15.

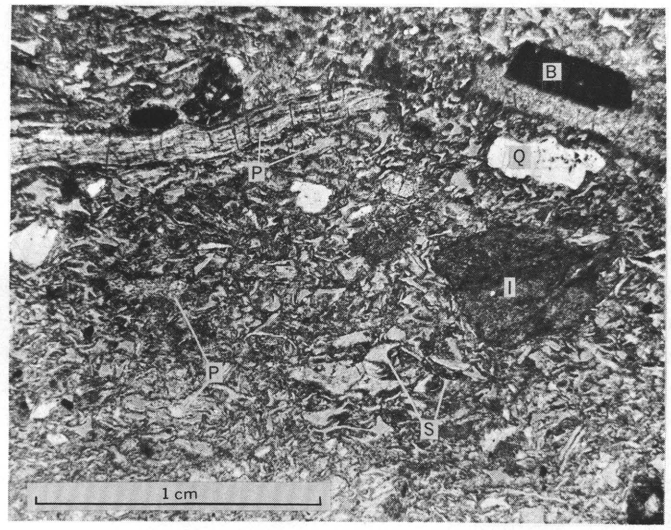


FIGURE 15.—Photomicrograph of welded tuff, showing collapsed shards. Crossed nicols. Q, quartz; B, biotite; I, inclusions of Paleozoic chert; P, pumice; S, shards.

Vitric crystal tuff

Vitric crystal tuff forms lenses beneath the welded tuff on the ridge south of Mill Creek and thin beds in the volcanic rocks at the head of Cooks Creek. In outcrop it is chalky white to pale purple, and contains white angular fragments of pumice as much as 10 cm long, black biotite flakes, and flashing sanidine crystals. The pumice, flattened parallel to the bedding, gives the rock a conspicuous foliation. In the Cooks Creek area, pebbles and cobbles from underlying gravels are incorporated near the base.

The microscope shows the same mineral species as in the welded tuff, but the crystals are less broken and make up less than one-third of the rock. The groundmass is less devitrified than that of the welded tuff, and shards, pumice fragments, and small crystal fragments are clearly visible.

Bedded tuff and tuffaceous sedimentary rocks

Bedded tuff is best exposed in the southwest corner of the Mount Lewis quadrangle and in Cooks Creek valley. In both places it forms a well-bedded sequence of white, gray, or variegated, poorly indurated sediments that weather to rounded subdued topography. Some of the bedded tuff has excellent graded bedding from a pumice-filled base to a top of very fine ash and probably represents direct ash falls from the air. Much of the bedded tuff, however, has been deposited by streams or in lakes and is a mixture of pumice; ash; broken crystals of quartz, feldspar, biotite, and hornblende; and cobbles, pebbles, and grains derived from Paleozoic rocks. Many of the bedded tuffs and tuffaceous sedimentary layers have crossbedding, cut-and-fill structures, ripple marks, and lenses of pebbles and cobbles. Some of the fine-grained lake sediments have calcium carbonate cement and have yielded a sparse fauna of fresh-water gastropods and pelecypods of probable Pliocene age (p. 88–89).

Basalt

Flows of basalt (Tb) 10 to 30 feet thick form lenses in the Tertiary volcanic rocks of the Harry Creek and Cooks Creek area. Most of these flows are in the lower part of the volcanic sequence, and either rest on the basal gravel or fill valleys in the Paleozoic rocks beneath the welded tuffs. None can be traced for more than a mile. No dikes or plugs of basalt that might have been source vents were found.

The flows weather black to reddish brown, their tops are slaggy and vesicular, and they show tachylite along the base where they rest on Paleozoic rocks. The joints normal to the base are not regular enough to describe as columnar. The only megascopic phenocrysts are olivine and pyroxene.

The microscope shows phenocrysts of olivine and pyroxene in a felty groundmass of plagioclase laths and interstitial basaltic glass, chlorophaeite, and calcite.

The olivine phenocrysts are euhedral to subhedral and average about 0.5 mm in length. Small grains of olivine also occur in the groundmass. An optic angle of about 80° suggests a composition of Fo65-Fa35. In several specimens, bowlingite, calcite, and a little antigorite have completely replaced the olivine. The pyroxene is augite, in euhedral phenocrysts as much as 2 mm long and in small grains in the groundmass. Many phenocrysts are zoned and their center, having an optic angle of 55°, is separated from an outside rim of optic angle 47° by a rather sharp discontinuity; this suggests a decline in CaO content during crystallization. Plagioclase laths in the groundmass are unaltered calcic labradorite, very slightly zoned progressively toward a more albitic plagioclase at the edge.

Brown glass, crowded with minute, red, hexagonal flakes—perhaps biotite microlites—fills most of the interstices between the labradorite laths. The glass has negative relief against balsam. The norm of this rock has 15 percent orthoclase and the normative plagioclase is less calcitic than the modal. The norm also shows 3 percent quartz. This glass must therefore be moderately rich in alkalies and silica.

Other materials in the interstices include a very pale brown to pale-green clear glass, probably chlorophaeite, and serpentine replacing chlorophaeite. There is also considerable calcite which partly encloses clear plagioclase laths and occurs side by side with unaltered chlorophaeite. The calcite apparently is not hydrothermal but crystallized in the final stages of magmatic crystallization, for the basalt is unaltered and much of the chlorophaeite is fresh.

An analysis of this rock is given in column 1, table 7.

WEST MARGIN OF THE SHOSHONE RANGE

The Tertiary rocks along the northwest edge of the Shoshone Range from the mouth of Crum Canyon to Pipe Canyon consist of rhyolite—mapped as Tertiary volcanic rocks—and andesite, probably a lava flow. These rocks are on the downthrown side of the normal faults along the edge of the range and are partly buried by Quaternary alluvium.

Rhyolite

North of the mouth of Lewis Canyon, a mass of rhyolite at least 1,500 feet thick underlies the andesite. The rhyolite is massive, without bedding, banding, or mappable foliation. It clearly underlies the andesite, but the contact is covered, and no contacts with Paleozoic rocks are exposed. The exact nature of the mass is not known; it could be a dome or, more probably, a mass of welded tuff.

In outcrop, the rock is pale purple to pale green. Small phenocrysts of feldspar and quartz and scattered chips of quartzite and chert can be seen with a hand lens.

The microscope shows phenocrysts, many broken, of quartz, sanidine, plagioclase, and biotite, in a devitrified groundmass. Quartz phenocrysts are subhedral and many have rounded and embayed outlines. Sanidine is clear; it has an optic angle of about 10° in the orthoclase position. The very subordinate plagioclase is oscillatively zoned, and ranges in composition from a center of An_{50} to a rim of An_{25} and is chiefly andesine. Biotite, in ragged books, is generally oxidized to a dark reddish brown with a rim of magnetite dust. The originally glassy groundmass is now a patchy mosaic of quartz and feldspar. Under oblique light, very faint shadows suggest that the groundmass once was either flow banded or consisted of flattened and alined shards and pumice.

The small masses of rhyolite near the mouth of Whisky Canyon are also included on the map under the general rubric of Tertiary volcanic rocks. They resemble mineralogically the rhyolite near Lewis Canyon, but they have excellent laminated flow banding and are probably flows or domes.

An analysis of this rock appears in column 11, table 7.

Andesite

A flow of andesite (Ta), its top rubbly and brecciated, crops out in a narrow band between Quaternary alluvium and underlying rhyolite just north of the mouth of Lewis Canyon. It is dark purple in outcrop and contains conspicuous white phenocrysts of plagioclase and a few black phenocrysts of hornblende.

The microscope shows phenocrysts of plagioclase, hornblende, and pyroxene in a matrix of microlites and devitrified glass. The euhedral plagioclase phenocrysts are oscillatively zoned from about An_{60} in the interior to about An_{40} along the rim. Some hornblende phenocrysts are partly altered to chlorite. Others are greenish brown, and their rims of magnetite dust and small extinction angles indicate much oxidation. Small subhedral augite grains are unaltered, but pyroxene of an older generation, which once formed large phenocrysts, has been altered to chlorite, bastite, and calcite. The groundmass consists of a felt of minute feldspar microlites, much opaque dust, shreds of chlorite, and patches of calcite.

BRECCIA PIPES

Three breccia pipes or volcanic vents are clustered along a north trend just below the crest of the Shoshone Range west of Mount Lewis. Each covers about 1 square mile, rises through Paleozoic rocks, and contains several varieties of breccia, foundered blocks of Tertiary sedimentary rocks, and intrusive igneous rocks. They are named, from south to north after the nearest canyon, the Horse Canyon pipe, the Pipe Canyon pipe, and the Rocky Canyon pipe. Exposures are good and the pipes were mapped (pl. 5) on a scale of 1:15,840 or 4 inches to 1 mile. The units described in the following sections are shown on plate 5. The general map (pl. 1) serves only to show the general location of the pipes and their relations to the other volcanic rocks in the two quadrangles.

HORSE CANYON BRECCIA PIPE

Location and geologic relations

The Horse Canyon breccia pipe lies near the head of Horse Canyon. It is elliptical, about 1 mile in east-west diameter, three-fourths of a mile in north-south diameter, and has slightly flaring sides. Horse Canyon and its tributary canyons provide a vertical section of about 1,600 feet.

The pipe cuts chert, siliceous shale, quartzite, and greenstone of the Valmy Formation except at the east end, where it cuts Antler Peak Limestone. The pipe crosses several thrust faults that dip gently east to northeast; the rocks carried on the thrust plates, although highly contorted, generally dip with the faults. The pipe transects both bedding and older structures, neither seems to have affected its shape or location.

Coarse breccia

Discontinuous patches of coarse breccia lie along the circumference of the pipe and project a few hundred feet into it. Best exposures are (1) on the topographic nose extending into the pipe from the north (hereafter called the north nose), (2) on the south side of the south fork of Horse Canyon, and (3) near the top of the nose extending into the pipe from the east (hereafter called the east nose).

Contacts of the coarse breccia with the wallrocks are generally sharp but irregular. On the north nose, the wallrocks are much fractured to distances of as much as 20 feet from the contact; the contact is mapped at the outer limit of the obviously rotated blocks. Elsewhere, the wallrocks show little fracturing even along the contact, which can be confidently located within 1 to 2 feet. Irregularities of the contact—shallow curving embayments of breccia into the unfractured country rock and projections of country rock into the breccia—have an amplitude of 5 to 50 feet, both in plan and in section.

The coarse breccia is a mixture of large and small blocks and fragments ranging from microscopic chips to masses as much as 100 feet long. Most of the large blocks are internally fractured. All fragments, of whatever size, are angular; this indicates little abrasion and hence slight transport. The fragments are chiefly of the adjacent wallrocks. Fragments of Tertiary porphyries are very scarce.

Some of the coarse breccia appears to have slumped from its original position. On the north nose, blocks of coarsely diabasic greenstone are distributed through a zone about 100 feet high and as much as 300 feet below outcrops of similar greenstone in the walls. Similarly, on the south side of the pipe, large blocks of silicified shale lie below their parental ledges which overly the chert here forming the walls of the pipe.

Fine breccia of the central part of the pipe intrudes the coarse breccia along dikes that roughly parallel the pipe walls and appear to wedge blocks of coarse breccia away from the walls.

The angularity of fragments, their great range in size, the slight mixing of rock types, the subsidence of some blocks, and the generally sharp contacts between breccia and walls all suggest that the coarse breccia was formed by landsliding or slumping. Moreover, the curving fractures, along which the fine breccia intruded, roughly parallel the walls of the pipe and resemble the sole faults or incipient sole fractures beneath landslides.

Fine breccia

Fine breccia occupies much of the inner part of the pipe, and where coarse breccia is absent, extends to the margin. It invades coarse breccia on scales ranging from through-going dikes several yards thick that isolate large masses of coarse breccia, to narrow, branching and irregular veinlets that fill fractures in large breccia blocks. (See fig. 16.) Where the dikes cut the finer grained parts of the coarse breccia their edges are irregular and incorporate some of the coarse breccia. Where the dikes cut large blocks, contacts are sharp. No dikes of fine breccia were seen to cut the wallrocks. The rough parallelism between many of the dikes and the walls of the pipe has already been noted.

FIGURE 16.—Slab showing dike of fine breccia cutting altered silicified shale in coarse breccia. From south side of the Horse Canyon pipe. Large fragment at left is of Tertiary porphyry.

The fine breccia is structurally homogeneous. Flow banding, lineation, foliation, shears, or fault zones are lacking. Joints are common but are so irregular, both individually and in pattern, as not to justify mapping.

In outcrop and hand specimen, the fine breccia consists of angular to subrounded rock fragments in a dense siliceous matrix, with subordinate quartz grains, feldspar crystals, and biotite flakes. Outcrops are white, gray, dark green, dark purple, and rust red. Areas of intensive hydrothermal alteration are light colored; red outcrops contain much pyrite and pyrrhotite.

Fragments range in diameter from microscopic to about 3 cm; a few scattered exceptions are as large as 10 cm. Near the edges of the pipe, the fine breccia contains inclusions of coarse breccia, but nearly all in the central part of the pipe has a consistently small grain size. In some narrow dikes and thin veinlets of fine breccia, the grain size is smaller than elsewhere; perhaps the larger fragments were strained out during intrusion through narrow orifices.

The fine breccia differs from the coarse in composition as well as in grain size. It contains a thorough mixture of all Paleozoic rock types nearby and also of Tertiary volcanic rocks, particularly porphyries. About one-half to two-thirds of the fragments are of Paleozoic chert, shale, quartzite, siltstone, and greenstone. The remainder are of quartz monzonite, quartz latite, and rhyolite porphyries, and quartz, plagioclase, biotite, and hornblende crystals. Near the margins of the pipe, dikes that cut the coarse breccia show more fragments of the adjacent wallrocks or of coarse breccia, but veinlets of fine breccia even as thin as 1 to 2 inches still contain Tertiary porphyry fragments and Paleozoic sedimentary rock chips foreign to the adjacent rocks. The fine breccia has been thoroughly homogenized. (See fig. 17.)

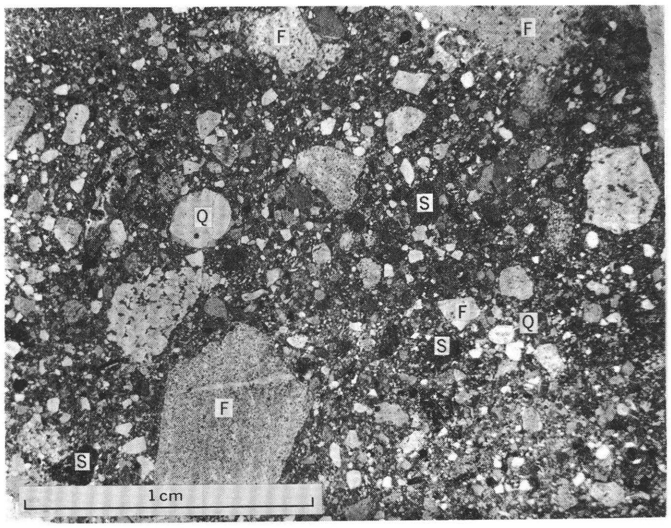


FIGURE 17.—Photomicrograph of fine breccia from the Horse Canyon pipe. Shows heterogeneity of fragments and their fine comminution. Q, quartz; F, felsite; S, siltstone. Plain light.

A careful search of more than 20 thin sections in transmitted, oblique, and reflected light failed to find shards of pumice. Many samples are so devitrified and hydrothermally altered that original shards might have become unrecognizable through recrystallization. Equally altered rocks elsewhere, however, do retain traces of shards so it seems probable that the fine breccia never contained them.

The microscope shows that the mixture of rock varieties, lack of structure, and unsorted mingling of angular and subround fragments persist down to the scale of a thin section. The crystal and rock fragments are held in a siliceous, isotropic to devitrified glassy matrix crowded with opaque dust, small rock and mineral fragments, and minute shreds and specks of chlorite, sericite, and clay. In the highly devitrified samples, the matrix is a patchy mosaic of quartz, chalcedony, feldspar, chlorite, and sericite; its exact nature is uncertain, but it is probably finely divided rock flour, silicified and recrystallized hydrothermally.

All the fine breccia has been somewhat altered hydrothermally. Chert fragments have been recrystallized to quartz aggregates, feldspars saussuritized, hornblende and biotite altered to chlorite and epidote. Patches of calcite are scattered throughout the matrix and in feldspars and ferromagnesian minerals. Many spaces between fragments are filled by a yellowish, highly birefringent, platy mineral, perhaps nontronite. Pyrite and pyrrhotite are widespread.

Near the central plug of quartz monzonite porphyry, veinlets of quartz, K-feldspar, and green hornblende cut the fine breccia. The original chlorite of the greenstone fragments has here been altered to amphibole, and diopside has replaced limestone in some chips of Valmy(?) limestone. Temperatures here were probably higher than elsewhere in the pipe, suggesting that some of the hydrothermal activity stemmed from the central plug. The alteration, however, is not clearly zonal around the central plug; fine breccia alongside some later quartz latite and rhyolite dikes has been much altered and the dikes themselves show devitrified groundmasses and altered phenocrysts.

Conglomerate

Several blocks of Tertiary conglomerate have apparently foundered in the breccia pipe. Two blocks are exposed on the north nose at altitudes of about 7,700 and 7,500 feet. Another large block covers the east nose between altitudes of about 7,000 and 8,100 feet.

The conglomerate is surrounded and intruded by, and mixed with, the fine breccia. Locally, on the north nose, the breccia sharply truncates the conglomerate. Elsewhere, irregular tongues of fine breccia infiltrate the conglomerate and isolate individual pebbles or groups of pebbles. On the east nose, the contact between breccia and conglomerate is gradational and the dashed contact on the map (pl. 5) should be interpreted as representing a broad zone in which faintly bedded conglomerate is mixed with structureless fine breccia.

The conglomerate consists of round to subangular pebbles and a few cobbles of Paleozoic sedimentary and Tertiary volcanic rocks. The blocks on the north nose show excellent bedding, marked by differences in grain size, cross bedding and scour along bedding surfaces. Attitudes vary considerably within the blocks. The conglomerate on the east nose is made up of lenses and pods of bedded pebble and cobble conglomerate and sandstone interbedded with considerably more unbedded, unsorted conglomerate. The conglomerate resembles the fine breccia in the mixture and general proportion of the different rock varieties, but contains numerous and better rounded cobbles.

These similarities between the conglomerate and the fine breccia suggest that the conglomerate was formed by reworking of the breccia by running water. Much of the conglomerate was apparently unconsolidated although some was cemented at the time that the underlying slightly younger breccia invaded and mixed with it, destroyed much of the bedding, and left remnants and confused masses of sorted sediment in a matrix of unsorted conglomerate and fine breccia.

The presence of conglomerate almost 1,000 feet below the upper exposures of the pipe and as much as 1,200 feet below the bedrock rim indicates subsidence. As the conglomerate stands lowest near the center of the pipe, more subsidence apparently occurred here than around the margin.

Quartz monzonite porphyry

A small plug of quartz monzonite porphyry covers an area of about a hundred acres in the bottom of Horse Canyon near the center of the breccia pipe. The porphyry intrudes both the fine breccia and the conglomerate along sharp contacts. Numerous apophyses cut the breccia, but the plug is itself invaded by fine breccia, especially along its borders.

The quartz monzonite porphyry is white to light gray in outcrop, showing grains of quartz and small phenocrysts of feldspar set in a finely granular matrix of quartz, feldspar, and biotite. Thin sections show that the porphyry consists of subhedral to euhedral phenocrysts of oscillatively zoned andesine, rounded and embayed phenocrysts of quartz, small books of biotite, and sparse hornblende, in a matrix of fine-grained granular K-feldspar and quartz. Samples from the edges of the plug show that angular fragments of quartz monzonite porphyry, along with many fragments of Paleozoic rocks, are included in the quartz monzonite. Most of the plug, like the surrounding fine breccia, has been much altered hydrothermally.

An analysis of a specimen from this plug is given in column 9, table 7.

Quartz latite and rhyolite porphyry dikes

Dikes of quartz latite and rhyolite porphyry cut all other rock varieties within the breccia pipe, except dikes of hornblende andesite, and extend into the surrounding Paleozoic rocks. In mapping it was not everywhere possible to distinguish a dike of quartz latite in which hydrothermal alteration had destroyed the biotite from one of rhyolite. Hence many dikes, probably including some of rhyolite porphyry and some of quartz latite porphyry, are simply mapped as quartz porphyry (pl. 5).

The dikes are discontinuous, range in width from 10 to 100 feet, trend north, and are vertical or dip steeply west. This orientation is not controlled by any obvious structure. (See fig. 18.)

The dikes show strong flow banding near their margins and dense glassy chilled margins as much as 6 inches thick. This chill zone has deep vertical grooves, resembling fault grooves or mullions, not only at the contact with the wallrocks but also within the chill zone itself; the magma must have been very viscous when intruded.

The quartz latite dikes are composed of phenocrysts of zoned andesine, embayed quartz, and biotite in a devitrified matrix. The rhyolite dikes have fewer

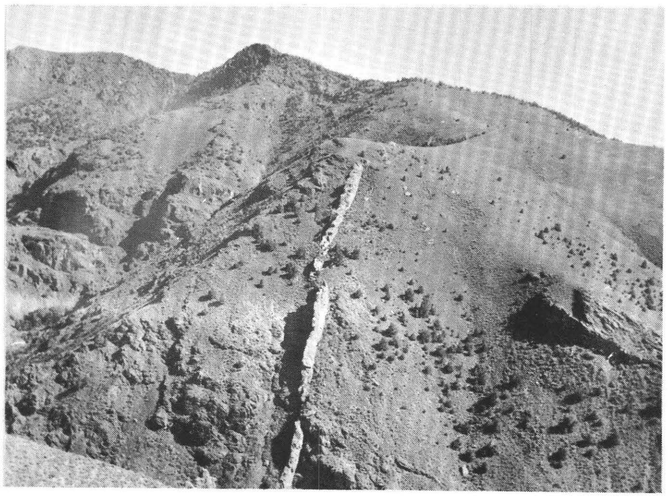


FIGURE 18.—Dike of quartz porphyry in south wall of Crippen Canyon, where it cuts Slaven Chert without regard to structure. This exposure is representative of many, both in and near the several volcanic centers and, as here, at some distance. Craggy slopes, Slaven Chert; smooth slope to upper right, sandstone of the Valmy Formation.

phenocrysts than the quartz latite—primarily albite and sanidine crystals with a very little biotite. Both varieties have undergone more or less hydrothermal alteration. Unlike the central plug of quartz monzonite porphyry, the quartz porphyry dikes are free of inclusions of fine breccia or conglomerate.

Hornblende andesite dikes

The final event in the history of the Horse Canyon breccia pipe was the intrusion of a few thin, discontinuous dikes of hornblende andesite. These cut the quartz latite dikes and occur both within and outside the pipe.

The dikes are black in outcrop. The only minerals visible under a hand lens are needles of hornblende and sparse rounded quartz grains. Thin sections show that the phenocrysts are thin laths and needles of hornblende as much as 1 mm long. In some dikes the hornblende is dark reddish brown, rimmed by opaque dust, and has an extinction angle of 5°–10°, suggesting considerable oxidation. In others, the hornblende is a light golden to orange, its $2V=80°$, and $Z \wedge c=16°$. The matrix consists of small labradorite laths, chlorite, and abundant patches of calcite. Calcite also replaces some phenocrysts whose shapes suggest pyroxene or olivine. A few rounded and embayed grains of quartz are rimmed by calcite.

Conglomerate and breccia near Horse Canyon mouth

About a mile west of the Horse Canyon breccia pipe is an area of conglomerate, coarse and fine breccia, quartz monzonite porphyry, and quartz latite porphyry—all probably representing a downfaulted remnant of the upper part of the Horse Canyon pipe.

The remnant is bounded on the north by a zone of coarse breccia containing acre-sized blocks of Antler Peak Limestone and Battle Conglomerate. The eastern contact against Ordovician rocks is a north-trending fault that dips gently westward. Tertiary conglomerate containing pebbles and cobbles of Paleozoic and Tertiary rocks and showing torrential crossbedding makes up most of the mass. On the southwest side is a small patch of fine breccia. A small body of quartz monzonite porphyry invades the conglomerate and several dikes of quartz latite porphyry cut the breccia and the conglomerate. The similarity in rock types and in their mutual relations between this small fault block and the Horse Canyon pipe is evident.

The stratigraphy and structure of the Ordovician rocks on the west side of the remnant match closely those of the Ordovician rocks high on the west side of the Horse Canyon pipe. If, in imagination, the rocks in these two areas are restored to coincidence by movement along the gently dipping north-trending fault that bounds the remnant on the east, the rocks comprising the remnant come very close to an upward extension of the flaring west edge of the Horse Canyon breccia pipe. The mass thus may be the downfaulted upper part of the main pipe. Such a reconstruction would place the original top of the pipe at a present altitude of about 8,500 feet and, would indicate a throw of about 2,000 feet on the fault.

Development of the pipe

The sequence of events leading to formation of the Horse Canyon breccia pipe appears to have been as follows: (1) formation of the fine breccia, (2) reworking of the fine breccia to form gravels and conglomerate, (3) subsidence accompanied by slumping of the steep walls to make the coarse breccia, (4) mobilization and intrusion of the fine breccia, (5) intrusion of the central plug of quartz monzonite porphyry accompanied by hydrothermal alteration, (6) intrusion of extremely viscous quartz latite and rhyolite porphyry dikes, (7) intrusion of hornblende andesite dikes.

The relations between coarse breccia, fine breccia, and conglomerate suggest an alternation of subsidence and intrusion of fine breccia. Only one cycle has been clearly preserved in the pipe, but several probably occurred. The thorough mixing of rock types, the fine comminution, and the rounding of some fragments indicate considerable turbulence and movement of the fine breccia. Inclusions of breccia in the central plug of quartz monzonite porphyry and the abundance of porphyry fragments in the fine breccia suggest repeated crystallization and brecciation as the magma worked its way upward. Landsliding and slumping of the wallrocks during subsidence would aid materially in enlarging the pipe and producing its flaring shape, and incorporation and reworking of the coarse breccia in successive cycles of subsidence and intrusion would add to the volume of fine breccia.

The cycles of subsidence and of intrusion of the fine breccia were probably controlled by changes in level and intrusive force in the central plug of quartz monzonite magma. Hydrothermal solutions or gases streaming upward from the plug would aid in mobilizing the fine breccia and produce much of the alteration so widespread in the pipe. The lack of shards and pumice argues against a tuff-forming eruption as the cause of the brecciation and intrusion.

PIPE CANYON BRECCIA PIPE

Location and geologic relations

The Pipe Canyon breccia pipe is at the head of Pipe Canyon near the Horse Canyon pipe, from which it is separated by a narrow band of quartzite, chert, and greenstone. To the northeast it joins the Rocky Canyon pipe.

It forms an irregular ellipse whose long diameter, about three-fourths of a mile, is oriented approximately northeast. The west side is steep to vertical, but the east side appears to dip west; the resultant east-west sectional shape suggests a tilted funnel. The pipe cuts across Ordovician sedimentary rocks and Paleozoic and Mesozoic thrust faults.

Most of the units mapped in the Pipe Canyon pipe are similar to those in the Horse Canyon pipe but some differ slightly.

Coarse breccia

Patches of coarse breccia are less abundant around the margins of this pipe than in the Horse Canyon pipe. The largest exposure is along the south edge where Ordovician chert and greenstone have been broken into large blocks, tilted and rotated, but not moved far from their original positions. Other outcrops of marginal coarse breccia are in the low saddle of the ridge cut by the southwest border of the pipe and on the south slope of the south fork of Rocky Canyon between the 7,400- and 7,600-foot contours.

In both these localities the coarse breccia contains a mixture of rock varieties, including Paleozoic chert, greenstone, siliceous shale, argillite, limestone, and Tertiary rhyolite, quartz latite porphyry, pumice fragments and fragments of consolidated fine breccia. Many fragments are subround; their diameters range from 3 inches to 1 foot, but a few are as much as 2 feet across. The matrix of this coarse breccia contains angular fragments of the same rocks as the large fragments and broken crystals of quartz, feldspar, and hornblende, surrounded by finely comminuted rock and by shreds and patches of chlorite and feldspar. This breccia has been thoroughly mixed, and the subrounding of many fragments indicates considerable abrasion.

Fine breccia

Fine breccia fills the southern and eastern half of the pipe. It intrudes coarse breccia and in turn is cut by pumiceous vitrophyre, rhyolite, and quartz latite.

In grain size, shape and angularity of fragments, mixture of rock types, and absence of structure, this fine breccia resembles that of Horse Canyon. There is, however, a major difference: this fine breccia contains many fragments of collapsed pumice, tuff, pisolitic tuff, and perlitic glass as well as the fragments of Paleozoic sedimentary and Tertiary intrusive rocks. Alteration and silicification have been like those in the Horse Canyon pipe.

Pumiceous vitrophyre

Several dikes of pumiceous vitrophyre—offshoots of the main mass of the Rocky Canyon vent—cut the fine breccia. The contacts between vitrophyre and fine breccia are sharp, and excellent flow banding parallels the contacts. Where the dikes pinch out, the vitrophyre is crowded with inclusions of fine breccia.

The pumiceous vitrophyre is a glassy vesicular rock of rhyolitic composition. It is described in detail in the section on the Rocky Canyon pipe.

Rhyolite

A plug of rhyolite porphyry occupies the west part of the pipe. It invades Ordovician rocks on the west and both fine and coarse breccia on the east and south. Flow banding is very faint or absent and inclusions of wallrock are sparsely scattered throughout the plug.

A swarm of rhyolite dikes cuts the plug and appears to radiate from it. Within the plug intrusive relations between dikes are very complex and no attempt was made to map the individual dikes. Clearly, though, the dikes are younger than the plug. They cut not only the plug but also each other, the fine breccia, coarse breccia, and pumiceous vitrophyre. The dikes have flow-banded margins, columnar jointing normal to their margins in many places, and contain no inclusions of Paleozoic or Tertiary rocks. Their pattern suggests that they radiate from a volcanic vent. Possibly the plug was a conduit for a vent and the radiating dikes represent renewed influxes of magma into the conduit.

The microscope shows euhedral phenocrysts of dusty unzoned albite, commonly streaked with wisps of chlorite and irregular patches of calcite. Indices suggest a composition of about Ab_{95}, but many extinction angles on sections normal to (010) are as high as 22°, higher than is customarily recognized in albite.

K-feldspar phenocrysts have the blocky outline characteristic of sanidine. Optic angles, however, range from 45° to 55° in the orthoclase position and indices of $\alpha=1.522$, $\beta=1.526$, $\gamma=1.528$, suggest an albite content of about 25 to 30 percent. These optical properties suggest orthoclase cryptoperthite. Some of the K-feldspar phenocrysts have a thin rim of albite. Some quartz phenocrysts are euhedral, with the outline of high quartz; others are rounded and embayed. Irregular clots of quartz, K-feldspar, and magnetite suggest former biotite phenocrysts.

The groundmass is highly devitrified, a patchy mosaic of quartz, feldspar, shreds of chlorite and sericite, and blebs of calcite. Spherulitic growths are common.

Both the minerals and the chemical analysis (column 12, table 7) show these albite-bearing rhyolites to be rich in soda. The abundant soda in, and the rims of albite around, some K-feldspars, and their generally altered appearance suggest soda metasomatism.

ROCKY CANYON BRECCIA PIPE

Location and geologic relations

The Rocky Canyon pipe lies at the head of Rocky Canyon and extends along the ridge between Rocky and Lewis Canyons: It joins the Pipe Canyon pipe to the south and the two are here separated more for convenience of description than because of geological significance.

The pipe covers a very irregular, triangular area of about half a square mile. One offshoot extends northwestward, another southeastward to the west wall of Lewis Canyon. The east side of the vent dips steeply west to northwest. The west and north sides are poorly exposed, and although probably steep, their attitudes are not known.

Coarse breccia

Alluvium and landslides cover the west and north margins of the vent and whether or not coarse breccia is present there is unknown. However, at the head of the north fork of Rocky Creek, several north-trending normal faults have jumbled together and apparently downdropped many fault blocks of Paleozoic and Mesozoic sedimentary rocks. This complex is truncated by pumiceous vitrophyre and also by the range-front faults west of Lewis Canyon. Part of the faulting may have accompanied formation of the vent or intrusion of the vitrophyre.

The east margin of the vent is well exposed and shows coarse breccia only in parts of the small offshoot on the west wall of Lewis Canyon where the vitrophyre invades Valmy chert and coarse Triassic(?) grit and limestone conglomerate. Blocks as much as 50 feet long of chert, greenstone, and quartzite are mixed with a finer breccia of the same rock varieties and of Triassic(?) sedimentary rocks. As the immediately adjacent wallrocks are chert, and the nearest quartzite, greenstone, and Triassic(?) rocks are on thrust plates high above the breccia outcrop, the breccia seems to have subsided.

Pumiceous vitrophyre

The main body of vitrophyre crops out between the north and south forks of Rocky Canyon. Several dikes cut the fine breccia of Pipe Canyon and a small offshoot invades chert on the west wall of Lewis Canyon. This is the only dike of vitrophyre seen to extend outside the vent.

Flow banding and foliation in the vitrophyre generally dip west and northwest, although local deviations in orientation suggest considerable turbulence. The vitrophyre seems to have risen from beneath the area of block faulting northwest of the vent. A swarm of narrow vitrophyre dikes, ranging from 1 to about 10 feet in width and thus too small to show on plate 5, sharply cuts the main mass and indicates a second eruptive episode from a northwest direction.

Contacts between vitrophyre and the breccias are generally sharp with rigorously parallel flow banding, but those with the blocks of tuffaceous sedimentary rocks next to be described are gradational through zones between 25 and 200 feet wide. The transitional zone is marked by a gradual decrease in the distinctness of the flow banding and foliation and a gradual increase in the proportion of breccia, tuff, and lava fragments included in the vitrophyre. Faint and distorted patches and lenses of crude bedding and sorting appear and gradually become so distinct that the rock is a well-bedded tuff. In many ways, these contacts resemble those between the fine breccia and the conglomerate blocks in the Horse Canyon pipe. The fact that the blocks of bedded tuff are steeply tilted, deformed, and completely surrounded by vitrophyre rules out the possibility that the gradation records reworking of the vitrophyre after emplacement. On the contrary, the vitrophyre must have engulfed comparatively unconsolidated tuff and tuffaceous sediment, stirred and destroyed the bedding, and incorporated fragments from them.

Contacts between vitrophyre and hornblende dacite block agglomerate are generally sharper than those between vitrophyre and the volcanic sedimentary rocks, perhaps because of better cementation of the agglomerate. On the north side of the south fork of Rocky Canyon, vitrophyre overlies block agglomerate along a contact marked by a rubble of fragments and by inclusions of block agglomerate in the vitrophyre.

Many chips and fragments of Paleozoic rocks, Tertiary porphyries, and glassy, tuffaceous, or pumiceous volcanic rocks are scattered throughout the vitrophyre. There are also thin pods and lenses of breccia resembling the coarse breccia of the Pipe Canyon pipe. Within the well-lineated and flow-banded vitrophyre are also ill-defined masses that grade into the vitrophyre and lack vesicles and vugs. These masses consist primarily of rock fragments like those of the fine breccia in Pipe Canyon. Probably the Rocky Canyon vent was once largely filled with breccia, both coarse and fine, much of which became incorporated in the pumiceous vitrophyre, while the remainder was carried to the surface and there deposited as tuffs and extrusive breccia.

In outcrop, the pumiceous vitrophyre is light gray to white; its texture is porcelaneous. It is hard and rings when struck by a hammer. Strung-out lenses of inclusions, the approximate alinement of flat chert chips, open lens-shaped vugs, and stretched-out, flattened, and elongated vesicles and fragments of collapsed pumice all testify to flowage. The flow structures wrap around the large inclusions and show many intricate swirls and folds.

The microscope shows that some pumiceous vitrophyre is vesicular glassy lava containing phenocrysts of quartz, sanidine, and plagioclase, and many rock inclusions. Other thin sections are of vitrophyre consisting of fragmental vesicular lava and glass. Most of the vitrophyre is crowded with small vesicles, so stretched that their thickness is a small fraction of their length. The vesicles wrap around rock inclusions and phenocrysts and form many flowage folds. (See fig. 19.)

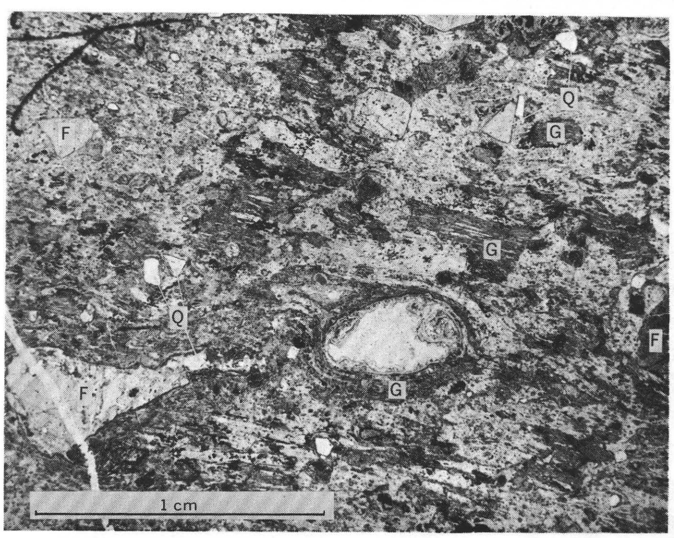


FIGURE 19.—Photomicrograph of pumiceous vitrophyre from the Rocky Canyon pipe. G, glass; Q, quartz; F, felsite. Light areas are devitrified patches. Plain light.

In addition to the small vesicles, elliptical vugs as much as 3 mm long and 1 mm thick are common. The small vesicles are deformed around the vugs. Many vugs contain a rock fragment at one end and several have fragments at each end, thereby forming a dumbbell-shaped vug. The two fragments in these dumbbell-shaped vugs clearly have formed by the fracturing, pulling apart, and rotation of an originally single fragment during flowage. The magma was apparently so viscous as not to fill the space between the two fragments of the fractured inclusion.

Much of the pumiceous vitrophyre has been altered hydrothermally, the feldspars saussuritized, and the hornblende and biotite converted to chlorite, magnetite, calcite, and quartz. The glassy vesicular groundmass has been devitrified; it shows faint granular pale-gray birefringence and contains wisps and irregular patches of a yellowish-brown birefringent mineral, perhaps an iron-bearing montmorillonite. Vesicles and vugs are filled with opal, chalcedony, quartz, or the brown clay mineral. Some rocks were devitrified in patches, more intensely near minute cracks that slightly offset some of the vesicles, perhaps by gases that followed the cracks.

The pumiceous vitrophyre erupted as a viscous magma, some remaining liquid enough to flow without fracturing but some so viscous that it fractured. The numerous vesicles and vugs and its prevalent devitrified and altered condition indicate that it was evolving much gas during its ascent. The glassy texture points to rapid cooling, which may have facilitated separation of a gas phase. On reaching the surface it may well have poured out as an expanding mass of shards and pumice in a cloud of its own gases to form a welded tuff.

Bedded tuff

The pumiceous vitrophyre has engulfed several large blocks of bedded tuff and dacite block agglomerate. The blocks have been tilted, generally eastward at angles ranging from 30° to 90°. Within the blocks are many small folds and faults too small to map on the scale of plate 5.

The tuff includes several varieties. Beds of rock fragments and ash that show crossbedding, cut-and-fill structures, and scour along bedding planes, clearly originated as stream deposits. Massive or crudely graded beds of shards and pumice probably represent ash falls directly from the air. Massive layers of coarse tuff-breccia indicate much explosive volcanism. Lenses of very fine ash, thinly laminated and crossbedded, probably represent pond or lake deposits.

Some ash beds contain spherical pellets about 5 mm in diameter. Around a nucleus of coarse fragments is a concentric shell in which the grain size decreases outward and in which shards and rock fragments are oriented concentrically. Some pellets are slightly flattened parallel to bedding and others are broken.

Similar pellets have been observed to form by accretion of ash in a moisture-laden eruption cloud (Perret, 1913, p. 612) and by "snowballing" of ash, wet with rain and melted snow, down volcanic slopes in Japan (Tanakadate, 1927, p. 165). Concentric pellets in breccia pipes in Missouri have been attributed to accretion in an atmosphere of magmatic spray during an explosive eruption (Rust, 1937, p. 62–69). The pellets in the Rocky Canyon pipe doubtless formed during explosive eruptions.

Grain size of the bedded tuff ranges from submicroscopic to cobble size. Most fragments are composed of sand, grit, pebbles, ash, and lapilli. Water transport was apparently short, as angular to subangular fragments predominate over rounded ones. About 30 percent of the sand grains, pebbles, and cobbles are of Paleozoic rocks; the rest are of Tertiary porphyries, glass shards, pumice lapilli, perlitic glass, and crystals of quartz and feldspar.

Within the tuffs are lenses of block agglomerate of hornblende dacite much like that in Indian Creek valley. Roughly spherical blocks of hornblende dacite porphyry that range from 1 to 6 feet in diameter lie in a matrix of angular fragments of dacite and fractured crystals of quartz, plagioclase, and hornblende. The dacite has been much altered and oxidized as was the agglomerate of Indian Creek.

The tuff and agglomerate clearly record explosive eruptions older than the pumiceous vitrophyre. The coarse grain size and angularity of fragments, the pellets. and the lenses of block agglomerate all suggest a nearby source. The lake beds probably record small caldera or crater lakes near the Rocky Canyon pipe.

The relation of the tuff in the Rocky Canyon pipe to the fine breccia of the Pipe Canyon breccia pipe is obscure, as the two were not found in contact. Most of the fragments of Paleozoic rocks are much the same in both. On the other hand, fragments of tuff, perlitic glass, and pelitic tuff abound in the fine breccia. Both tuff and breccia thus appear to represent mixtures of the same components, part bedrock and part volcanic. This mixing could have come about either by intrusion of unconsolidated tuff by the fine breccia, like the intrusion of conglomerate by breccia in the Horse Canyon pipe, or by several cycles of volcanic explosions, reworking by running water, and subsidence.

The blocks of tuff are as much as 300 feet below neighboring country-rock ridges and are tilted, fractured, and folded. Because the very viscous vitrophyre shows evidence of strong upward flow, it probably would have tended to carry inclusions upward rather than permit them to sink. Thus the subsidence and fracturing of the tuff probably took place before the vitrophyre was erupted and the pyroclastic rocks may have been carried upward from even lower levels within the vent. Field evidence is inconclusive as to the timing, but some subsidence must have occurred.

Rhyolite

The final intrusion in the Rocky Canyon vent was of viscous rhyolite, chiefly along the borders of the northern offshoot. The rhyolite carries no inclusions, has excellent flow banding, and consists of phenocrysts of quartz, albite, and sanidine in a devitrified matrix.

Relation to Horse Canyon pipe

The combined Pipe Canyon-Rocky Canyon complex has many parallels to the Horse Canyon pipe. It was once probably largely filled with fine breccia, some of which may have been reworked into younger sediments. Inclusion-filled, volatile-rich rhyolite invaded both the fine breccia and sediments. Intrusion of inclusion-free flow-banded dikes ended the igneous activity. The main differences between the centers are that, in the Pipe Canyon-Rocky Canyon centers, ash, pumice, and block agglomerate were wholly erupted before the vitrophyre, and here the magma was rhyolite instead of quartz monzonite porphyry as in the Horse Canyon pipe.

ORIGIN OF THE BRECCIA PIPES

Many breccia pipes and vent breccias have been studied and many mechanisms for their origin proposed. The well-known pipes of the Colorado Plateau are generally associated with basaltic or basic alkaline volcanics (Dutton, 1885, p. 164–179; Johnson, 1907, p. 322–324; Williams, 1936; Hunt, 1938, p. 68–73; Shoemaker, 1956, p. 180–183) and resemble those in the Shoshone Range only in a general way. Some of the breccias of the British Tertiary volcanic province are more nearly comparable, as several are associated with siliceous magmas along ring dikes. All these breccias are dominated by fragments of either wallrocks or other previously consolidated rocks (Johnson, 1907, p. 322–324; Williams, 1936, p. 116–120, 131–148; Hunt, 1938, p. 68–73; Shoemaker, 1956, p. 180–183; Tyrrell, 1928; Bailey and others, 1924, p. 199–210; Richey and Thomas, 1932, p. 805–811; Richey, 1940).

Most of these have been ascribed to violent explosions and subsequent infilling of the vent or crater by debris hurled into the air and by slumping of the walls. Richey (1940, p. 168) however, noted that the walls of some of the vents are shattered, wallrock fragments are only slightly displaced, and brecciated rock at depth has remained in place. Johnson (1907) and Hunt

(1938) reported similar observations in the Mount Taylor volcanic field, New Mexico.

Several differing but allied theories attribute the brecciation to various kinds of repeated or intermittent gas fluxing and explosive activity, which did not necessarily begin with a single gigantic explosion (Cloos, 1941, p. 783–789; Williams, 1929, p. 179; Hunt, 1938; Shoemaker, 1956; Lovering and Goddard, 1950; Rust, 1937, p. 71–72; Reynolds, 1954). The growth of the breccia and its containing pipe or vent is gradual. Attrition by fragments in the gas stream, in addition to the explosiveness of the gas itself, is an important agent of brecciation and of erosion of the walls.

Many of the Pando sills in Colorado terminate in lenses of breccia that contain fragments of the sill and of the surrounding strata, so that Tweto (1951, p. 527) concluded that gas expelled by chilling of the magma brecciated the rocks ahead of the advancing sills and aided their emplacement. Such a process might also assist magmatic dikes to rise through overlying rocks.

Another group of theories is based on chemical reaction between the magma, its volatiles, and the wallrock, resulting in solution and collapse. The breccia pipes of carbonate rocks in the Terlingua quicksilver district of Texas (Thompson, 1950, p. 35) and the Pilares breccia pipe of Sonora, Mexico, which is composed of mineralized, altered, and subsided andesite fragments (Locke, 1926), have both been attributed to solution stoping. The breccia pipes in the Red Mountain mining district of Colorado have been thought to be due to the upward passage of magmatic emanations through favorably jointed and fractured volcanic rocks (Burbank, 1941, p. 177–178). These emanations "caused certain chemical or volume changes along some vertical axis of greatest concentration, which impaired the strength of the rock and induced local crackling and crumbling."

Another mechanism that can cause brecciation of rock was briefly mentioned by Locke (1926, p. 446) but perhaps deserves more consideration than it has been granted heretofore. Rock bursting occurs both in quarries at the surface and in mines at depth and can produce many types of breccia on a large scale. White (1946) and Bain (1931) described rock expansion and bursting in granite and limestone quarries. The report of the Witwatersrand Rock Burst Committee (1925, p. 14–15) gives this graphic description of a rock burst in one of the deep Rand gold mines:

> without any warning a sudden shock extinguished all lights in the area affected and the footwall appears to rise and the hanging wall to descend. Props are shattered, pigstyes and packs are compressed, and quantities of rock burst from the faces of the stopes, from pillars, from dykes or other remnants of ground which have been left unworked, and from the sides and roofs of levels. Tracks are buckled and displaced and pipelines bent or broken; cracks open along fault planes and sometimes the rock along the fault is so shattered and brecciated that quantities fall into mine excavations. Often with the shock the solid face of a stope or pillar punches into and shatters the hanging wall or both of those walls in its vicinity and the falls from the hanging wall so caused may result in even more loss of life than the actual burst from the face. The shock may also shake down slabs of hanging wall at some distance from the place where the burst occurs. After a severe burst it is frequently the case that movements immediately cease, but sometimes movement of the hanging wall will continue more or less spasmodically for some hours or even days.

The immediate cause of rock bursting in quarrying or mining is excavation of the rock to give a free face, thereby causing a stress concentration. Rock at depth is under essentially isotropic stress, primarily due to load, although other factors have been cited (Bucky, 1942, p. 30). When a free face is formed the stress becomes concentrated, and if sufficient, brecciates the rock.

A process much like rock bursting has been induced in the laboratory. Bridgman (1918), in order to study the behavior of cavities in rocks under pressure, put cylinders of several minerals and varieties of rock, in which a cylindrical hole had been drilled, under hydrostatic pressures as great as 7,000 kg per cm^2. He found that the cylinders failed by a mechanism resembling rock bursting. Bridgman (1918, p. 266) summed up his results thus:

> Cavities in the materials dealt with in this paper, which may be broadly characterized by the property of brittleness, exhibit a method of failure under high compressive stresses not shown by ductile materials like metals. This method consists of shooting-off of minute fragments with considerable violence from the walls of the cavity. The frequency, and probably the velocity, of projection * * * becoming greater at higher pressures. This mode of disintegration is shown both by rocks and by single crystals; in rocks the splinters show no relation to the boundaries between chemically homogenous parts of the mixture, and in crystals there is no obvious connection with the crystalline symmetry.

The depth at which rock bursting begins in mines differs from place to place and from mine to mine. In the Rand gold mines bursting becomes a serious problem at 2,000 to 3,000 feet (Witwatersand Rock Burst Comm., 1925, p. 38); in the Lake Superior copper mines, it begins at about 1,000 feet and intensifies with depth; it is a constant hazard at 6,000 feet or greater depths (Crane, 1929, p. 10).

Rock bursting is most violent in the brittlest rocks: in the Rand, the massive quartzite of the hanging wall (Witwatersand Rock Burst Comm., 1925, p. 21), and in the Lake Superior region, the hard, strong, and brittle traprock. Crane (1929, p. 6) concluded as a rule of thumb that the harder a rock is to drill, the more it is subject to rock bursting.

Not all rock bursts are as violent as the one described; the rock may brecciate only along a single face or throughout extremely large volumes. Near Johannesburg, a rock mass estimated to weigh 500 million tons broke up in sections during a movement at the surface of only 2 or 3 inches and at the 14th level of about 9 inches (Witwatersand Rock Burst Comm., 1925, p. 13). Crane (1929, p. 37) reported that rock bursting in one Lake Superior mine produced cracks in the surface as much as a mile away. Rock bursting also causes spitting or spalling of small fragments from the face, some as far as 12 feet.

Rock bursts produce breccias of a wide variety. Some are incoherent masses of angular fragments of all sizes from dust to blocks weighing many tons. Others consist of small fragments described as the size of a man's fist or of road metal. One observer reported complete disintegration of a hanging wall "so that it had little more cohesion that sand" (Witwatersrand Rock Burst Comm., 1925, p. 12, 14). Rock bursts surround some tunnels and stopes with zones of breccia; some stopes cave upward to an arch, leaving far more broken rock than was originally mined, a process utilized in the block-caving methods of mining.

In the Shoshone Range, no particular breccia can be identified as a product of rock bursting, although the breccia dike offshoots from the small crescent-shaped plug of granodiorite on Bullion Mountain (p. 51) may be, as they do not seem related to faulting. Volcanic activity might, however, produce free faces at depth. Intrusion or extrusion elsewhere could drop the magma level in a chamber, or free faces might form along fractures made by magmatic or gaseous pressure or be left in the empty conduit after a gas eruption. Furthermore, if gas under high pressure rapidly escaped up a pipe and broke through to the surface, the pressure on the walls would be reduced as the pressure energy of the gas was converted to kinetic energy, as in a Venturi tube. The wallrocks might thereupon burst into and eventually choke the gas stream. In the Shoshone Range the rocks on the upper plate of the Roberts thrust are largely brittle chert, siliceous shale, sandstone, and quartzite. Intermittent explosions and many intrusions took place and material subsided in the breccia pipes. Conditions must have favored rock bursting, but no particular breccia can be proved a product of it and in any event it could not alone have formed the pipes; eruptive and explosive activity such as was briefly reviewed above could have made all the breccias.

The pipes were conduits for rising plugs of magma. The breccias formed above these rising plugs and were invaded and mixed with them. Alteration and vesiculation indicate abundant magmatic volatiles. The primary factor in forming the pipes was the rise of a plug of volatile-rich magma.

The abundant tuff and agglomerate among the extrusive rocks, both in the Rocky Canyon vent and in the area as a whole, show that explosive eruptions were common and doubtless contributed to making the breccias.

Intermittent intrusion and eruption apparently alternated with subsidence. Subsidence is demonstrable in the breccia pipes, and the intrusive relations, thorough homogenization, and fine comminution of the fine breccia all suggest several cycles.

The breccia pipes of the Shoshone Range were probably formed by a two-way pump action, impelled by intermittent intrusion of magma and evolution of eruptive volatiles. Intrusion of magma, rapid evolution of gases, and explosive eruptions formed breccia on the upstroke of the metaphorical pump. Landsliding, slumping, collapse stoping, and rock bursting made more breccia on the downstroke. A mass of breccia, becoming finer and more mixed with each impulse, thus formed ahead of the rising magma, was intruded by it, and in turn became intrusive itself. Eventually this carapace of breccia above the magma broke through to the surface where it was reworked by running water, mixed with pyroclastic debris of previous eruptions, and then during renewed upward movement, invaded or engulfed blocks of pyroclastic rocks and landslide debris which had foundered during episodes of subsidence. The underlying magma invaded this complex of breccia and reached the surface as pumiceous vitrophyre, perhaps there to form a welded tuff. Magmatic activity closed with intrusion of viscous acidic magma, probably drained of its volatiles by previous explosions.

QUARTZ LATITE BRECCIA OF MOUNT LEWIS

DISTRIBUTION AND GEOLOGIC RELATIONS

"an immense outburst of gray and green dacite. It is the grandest and most elevated body of dacite to be found within the limits of the Fortieth Parallel Survey." Thus Hague (1877, p. 620), visiting the area as a member of the Fortieth Parallel Survey, described the intrusive quartz latite breccia of Mount Lewis, which in his day was more aptly named Shoshone Peak. (See fig. 20.)

The breccia makes up the summit of Mount Lewis, extends eastward in two spurs separated by a deep valley, forms several small exposures in Indian Creek valley and a small remnant on the spur west of the summit. Its original extent is unknown; but as no other outcrops are found in the two quadrangles, it probably was not widespread. As the geologic map and sections (pl. 5) show, the breccia rests on Paleozoic

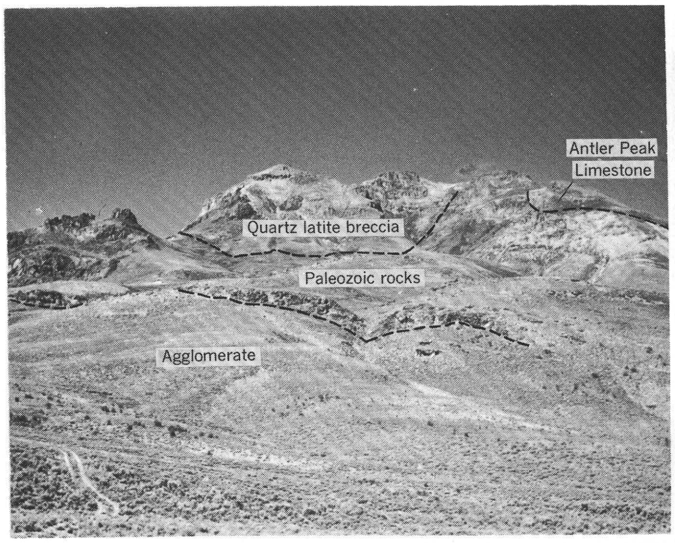


FIGURE 20.—Mount Lewis, from a point near Maysville Summit. Block agglomerate of hornblende dacite in foreground; Paleozoic rocks above and quartz latite breccia of Mount Lewis on left and middle summits; dip slope on right summit is Antler Peak Limestone above the Havingdon Peak fault.

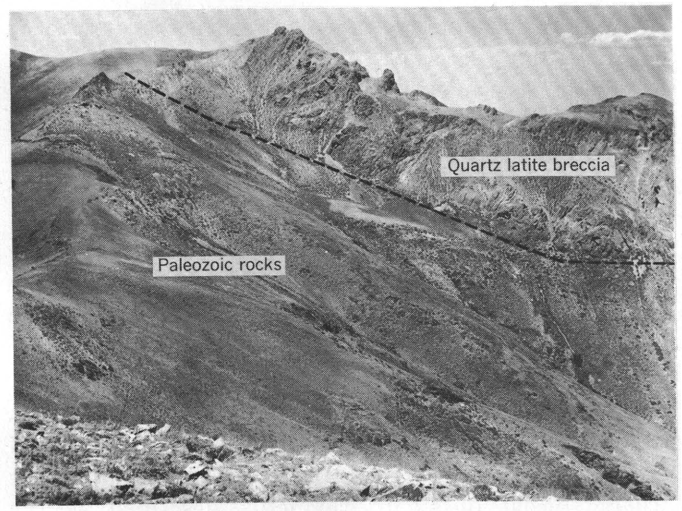


FIGURE 21.—Looking northeast toward the southeast spur of Mount Lewis. Intrusive breccia forms scree-covered upper slopes, terminating downward at the right in massive outcrops. Smooth slopes below are on Paleozoic bedrock, above the Kattenhorn fault.

rocks or a coarse Tertiary breccia and thus appears to be extrusive. Considerable evidence, however, points to the conclusion that the quartz latite breccia was intruded along the unconformity between Paleozoic rocks and the overlying Tertiary extrusive rocks, the latter, except for a small remnant west of the summit of Mount Lewis having since been removed by erosion.

The breccia rests on the Valmy Formation and Battle Conglomerate along a steep contact between the summit and Indian Creek valley. This contact is exposed in small gullies on the south side of the southeast spur of Mount Lewis, in the valley between the southeast and northeast spurs, and along the northeast spur. In all these localities the breccia clearly rests on the older rocks; no dikes or other bodies of breccia intruding the older rocks from below were seen. (See fig. 21.)

In many exposures, the underlying rock, particularly where it is chert, is brecciated to a depth as great as 10 feet, and the adjacent breccia contains more abundant bedrock fragments than elsewhere. In the bottom of the gulch that dissects the breccia on the east, knobs of Battle Conglomerate and breccia of Valmy chert protrude through the quartz latite breccia. The quartz latite breccia also both overlies and includes blocks of Battle Conglomerate, chert breccia, and of a second very coarse breccia which contains round cobbles and boulders derived from the Valmy Formation, the Battle Conglomerate, hornblende dacite, and granodiorite. Some of this younger breccia has a much contorted tuffaceous and clayey matrix; it is probably a mudflow, perhaps in part volcanic, and its content of dacite and granodiorite clearly implies its Tertiary age.

A probable remnant of the upper contact of the quartz latite breccia with the Tertiary rocks, beneath which it was intruded, is exposed on the west spur of Mount Lewis. On this spur, Tertiary extrusive rocks overlie the quartz latite breccia, and both extrusives and breccia are cut off below by a low-dipping fault, locally having a few patches of tectonic breccia and locally buried beneath landslide deposits on the west side of the spur. The wedge of Tertiary rocks above this fault and in a subsidiary fault block within it, consists of a jumble of large and small blocks of quartz latite breccia, finely laminated tuffaceous lake beds (some of which have ripple marks and laminate crossbedding), rhyolite tuff, coarse mudflow breccia, stream-bedded arkose, and dacite block agglomerate. This wedge appears to be a large slump block underlain by a shallow, curving sole fault.

Along the northwest and west sides of the west spur of Mount Lewis, the stratigraphic succession of the Tertiary rocks in this wedge is not disturbed. From the top downward, the sequence is as follows:

The uppermost rock, the dacite breccia, consists of angular blocks of banded hornblende dacite as much as 5 feet long in a matrix of smaller blocks, broken crystals, and glass fragments. Hornblende and biotite are black from oxidation, and plagioclase is saussuritized. The dacite breccia rests disconformably on stream-bedded tuffaceous and arkosic sandstone and grit beds, which in turn conformably overlie a coarse mudflow that consists of large balls of mudstone and boulders of tuffaceous sandstone and conglomerate held in a matrix of dark-red mudstone and contorted streaks of sand.

The rock body beneath the mudflow and above the quartz latite breccia is made up of angular to round fragments of Paleozoic rocks, Tertiary porphyries, tuffaceous mudstone, feldspar, quartz, hornblende, and biotite. In the upper part, this material has crude bedding marked by lenses and pods of pebbles and cobbles but lower down, as the quartz latite breccia is approached, the pods and lenses become more and more discontinuous and contorted. Some seem to be broken off. Irregular masses of quartz latite breccia having gradational diffuse boundaries are mixed with the unsorted fragmental material and with remnants of the pods and lenses.

The contact between this rock body and the quartz latite breccia is gradational and characterized by a decrease in the number and roundness of fragments of Paleozoic rocks and a gradual change in color from the pale green of the overlying fragmental sedimentary material to the dark purple of the quartz latite breccia. In a few places the color change takes place along a sharp contact that has marked irregularities several feet in amplitude. The quartz latite breccia immediately below the contact has no foliation or sheeting and contains more fragments of Paleozoic rocks and of Tertiary porphyries than is common in the mass as a whole.

The quartz latite breccia apparently invaded relatively unconsolidated tuffaceous sediments and mixed with the lowermost bed of sediments, destroying or disturbing much of the bedding and incorporating fragments from it. At the same time it left irregular masses of welded breccia within the disturbed overlying beds.

On the southeast spur of Mount Lewis, the quartz latite breccia is in sharp contact with a plug of banded hornblende dacite. The contact dips very steeply east and transects the foliation and platy jointing of the breccia. On a gross scale the flow banding in the hornblende dacite parallels the contact; these relations suggest that the dacite intruded the breccia. Details of the contact oppose this interpretation, however. Locally the breccia cuts the flow banding of the dacite in many places, and along the contact, fragments of hornblende dacite are included in the quartz latite breccia. These details show the quartz latite breccia is younger than the hornblende dacite.

No source of the quartz latite breccia can be confidently identified. Difference in lithology and in the time relations between the breccia and the dacite plug argue against the plug being the source. A feeder pipe could well underlie the breccia itself on Mount Lewis, but evidence is lacking. The breccia in many respects resembles lithologically the fine breccia of the Horse Canyon breccia pipe, and it or any other of the three pipes may have been the source.

BRECCIA DIKES

Throughout the mass of quartz latite breccia are steeply dipping to vertical dikes of breccia. They cut the main mass along sharp contacts, are 6 inches to 3 feet wide, and rarely extend more than 20 to 30 feet. Most consist of breccia like the enclosing mass except for a slightly smaller average grain size. A few have scattered rounded pebbles and cobbles of Paleozoic rocks, and some are composed of angular fragments of chert and greenstone probably derived from the underlying brecciated rocks of the Valmy. Most of the dikes are more oxidized than the surrounding breccia and a few are red from hematite. These dikes probably were emplaced by steam streaming upward through cracks in the breccia.

STRUCTURE

Several sets of joints that cut the mass into irregular columns occur in places along the contacts of the welded breccia. The most consistent set strikes north and dips 75° to 80° W., somewhat steeper than the normal to the basal contact. Another conspicuous set strikes east and is vertical. Less well developed sets strike northeast and northwest and are nearly vertical. These joints tend to fade out in the central parts of the welded breccia and appear related to cooling of the breccia along its contacts. The columnar pattern in the upper parts of the mass suggests that the original upper contact was not far above the present surface.

Near and roughly parallel to the upper surface of the welded breccia is a widespread conspicuous sheeting, along which the rock splits into lenticular slabs ranging from 10 inches to 2 feet thick. The sheeting closely parallels the foliation and, like it, dips east from 5° to 20° more steeply than the basal contact; it steepens near the contact with the hornblende dacite plug.

Foliation formed by the parallel orientation of flat chips of chert and shale, flattened blebs of green glassy porphyry, biotite and chlorite flakes, and small elliptical vugs is conspicuous through most of the quartz latite breccia. Near the base it is less well developed than above and approximately parallels the contact with Paleozoic rocks; higher up it dips more steeply and in many places roughly parallels the present upper surface, except on very steep cliffs.

The facts that the sheeting and foliation commonly parallel the upper surface and that the joints stand normal to it suggest that, where undissected, the present upper surface may approximately parallel the original roof of the breccia. If so, the original shape of the mass was that of a tongue thickening westward—perhaps as a laccolithic wedge.

INCLUSIONS

Abundant inclusions of Paleozoic rocks and of

Tertiary porphyries foreign to the breccia are distributed throughout, more abundantly near the base and upper surface than in the middle. They make up 2 to 3 percent of the total volume of the rock. Most are angular fragments of chert, quartzite, and greenstone in the 1- to 5-mm size range, but a few are well-rounded pebbles and cobbles, chiefly of granodiorite and quartz monzonite (presumably from the Granite Mountain stock), and hornblende dacite porphyry. The angular fragments are distributed rather evenly, without local clots of high concentration, and seem thoroughly mixed with the breccia. Although the rounded cobbles and pebbles are also randomly mixed, they are commonly in clots and thin lenses that suggest masses of stream gravel, especially near the upper surface.

The abundant fragments near the base were partly incorporated from the underlying brecciated Valmy chert and greenstone. The abundant rounded cobbles and pebbles of Tertiary rock near the upper surface were doubtless incorporated from the overlying rocks. The distribution of fragments throughout the mass, however, would demand extremely turbulent intrusion. All the inclusions were assumed to have come from the adjacent underlying and overlying rocks. More likely the parental magma, like those of the quartz monzonite porphyry plug in the Horse Canyon breccia pipe and the vitrophyre of Rocky Canyon, contained many inclusions before intrusion.

PETROGRAPHY

The quartz latite breccia is a hard well-cemented rock that rings when struck with a hammer. It is pale purple to pale green; where highly altered it may be chalky white. The surface texture is glassy to porcelaneous. Inclusions of quartzite, chert, greenstone, and Tertiary porphyries, and crystals of quartz, biotite, hornblende, and feldspar can be seen with a hand lens.

The microscope shows fragments of quartz latite porphyry and glass, crystals of plagioclase, hornblende, biotite, and quartz, and inclusions of sedimentary rocks. These are surrounded by an isotropic to faintly birefringent cloudy siliceous matrix. The breccia is somewhat altered hydrothermally.

The fragments of quartz latite porphyry range from 2 mm to about 10 mm in length; the median is about 3 mm. Most are angular and about equidimensional, but many near the base of the breccia are flattened and elongated. The quartz latite of these fragments consists of euhedral to fragmentary crystals of oscillatively zoned andesine, rounded and embayed quartz, biotite, and horneblende, in a partly devitrified glassy groundmass. Some of this groundmass has perlitic cracks, some contains vesicles, and some is structureless. (See fig. 22.)

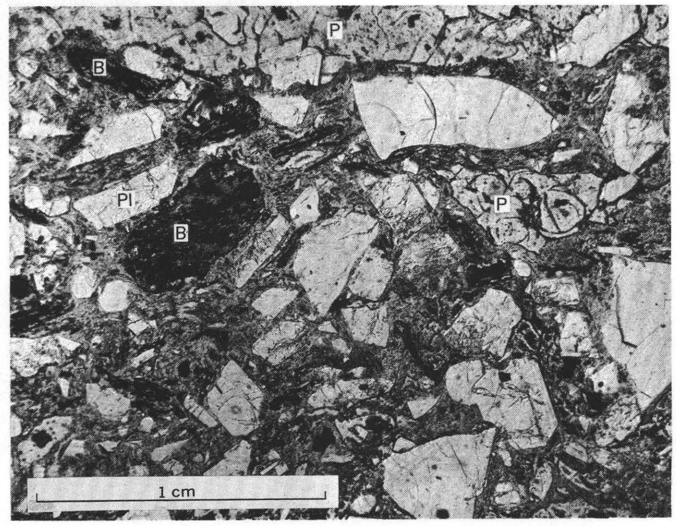


FIGURE 22.—Photomicrograph of quartz latite breccia from Mount Lewis. Along top is a large fragment of perlitic glass. P, perlite; Pl, plagioclase; B, biotite. Plain light.

The broken crystals in the breccia are of the same minerals and in the same relative abundance as in the porphyry. As broken crystals occur both in the porphyry fragments and as discrete crystals in the breccia, there were at least two episodes of brecciation. Angular to subround inclusions of Paleozoic rocks range from 1 to 6 mm in diameter, both in the porphyry clasts and in the surrounding breccia. Thus some were incorporated in the magma before the final brecciation.

All this broken material is surrounded by a siliceous devitrified matrix of small glass fragments, some resembling shards, much opaque dustlike material, and minute shreds and flakes of chlorite and clay minerals. This matrix resembles that of the fine breccias in the Horse Canyon and Pipe Canyon breccia pipes and like them has been hydrothermally silicified and recrystallized.

In the most intensely altered rock the feldspars have been altered to albite, clay, saussurite, and calcite. Some of the plagioclase crystals contain thin veins and small patches of K-feldspar. Hornblende and biotite have been replaced by chlorite, calcite, epidote, and magnetite. Some biotite has altered to quartz, K-feldspar, and chlorite. The matrix is devitrified and contains much calcite. In these much-altered rocks, the contacts between devitrified matrix of the breccia and the devitrified groundmass of the porphyry fragments are obscure.

In the least altered rocks, particularly near the base of the breccia on the northeast spur of Mount Lewis, the plagioclase and many crystals of hornblende and biotite are fresh. The perlitic porphyry fragments, however, are rich in irregular patches of a fibrous pale-olive to yellowish-green mineral, colloform, with radiating fibers or plates, a birefringence of about 0.025, parallel extinction, and a faint pleochroism. It is probably nontronite. The nontronite(?), along with chlorite, also forms curving films in the perlitic cracks, but is absent from the matrix. These nontronite(?)-bearing fragments form conspicuous strings of green pods parallel to the foliation.

A chemical analysis of slightly altered quartz latite breccia from Mount Lewis is given in column 8, table 7. Although the rock contains no K-feldspar, the norm shows 18 percent orthoclase. Clearly most of the potassa must be in the glassy matrix and groundmass of the porphyry fragments, although some is in the biotite and in sparse veins and patches of K-feldspar that replace some of the plagioclase.

ORIGIN

Field relations suggest that the quartz latite breccia of Mount Lewis was intruded as a laccolithic tongue approximately along the contact between Tertiary extrusive volcanic rocks and Paleozoic rocks. Hay (1954) described intrusive tuff breccias in the Absaroka Range that have jostled, truncated, and deformed overlying beds and locally brecciated the underlying ones. He concluded that the tuff breccias were originally extrusive and became intrusive by some mechanism, perhaps of a landslide nature, not clearly evident. Krauskopf (1948, p. 723 and fig. 13) described a lava flow near Parícutin, Mexico, that dived beneath the weakly consolidated tuffs, thus changing from extrusive to intrusive and back again. The lack of obviously intrusive bodies resembling the Mount Lewis breccia in this area makes such an assumed origin uncertain.

Durrell and Curtis studied intrusive breccias in California and each proposed a different mechanism of brecciation, either one or both of which may be applicable to the breccia at Mount Lewis. Curtis (1954, p. 469–470) suggested that vesiculation

> so increases the viscosity that the magma can no longer adjust internally by flow with sufficient rapidity to keep pace with the inexorably moving lava from behind. Large joints or fractures develop in the vesiculated but still plastic magma along which differential movements may cause slight dilation. For a brief instant, confining pressure drops to almost zero along dilated joints, and gases in vesicles immediately adjacent to the surfaces of joints, under confining pressure at least equal to that of the column of magma above, expand so rapidly in the direction of reduction of pressure that they cause spalling of the viscous lava along both joint surfaces. In response to very slight additional movement, continued spalling occurs. Thus, at almost the instant of the first fracture, the mass may become brecciated to an astonishing degree.

Durrell (1944, p. 269–270) proposed a slightly different mechanism for autobrecciation.

> The dike magma which was in an advanced stage of crystallization must have been rapidly cooled on being injected in small bodies into cold and wet wall rocks at a shallow level. Rapid chilling caused further crystallization and concentration of the volatile constituents in the ever-diminishing liquid residue. Thereby, pressures in excess of the confining pressure may develop. In the case of dikes 2 to 8 the magma continued to rise after brecciation and therefore probably during brecciation, causing rapid lowering of the external pressure. When rapid cooling and lowering of external pressure occur, violent escape of the volatiles is expected. That this should cause disruption of the nearly crystallized magma is clearly indicated by the structures of dike 1.

Durrell noted that the dikes are highly altered and that the following minerals had formed: residual albite, chlorite, epidote, serpentine or a chloritic mineral closely resembling bowlingite, and much clay of the montmorillonite type. He thought the alteration was caused by volatiles expelled from the andesite.

Either of these theories might account for the breccia of the Mount Lewis body. The abundant glass and particularly the perlitic porphyry fragments indicate rapid chilling. Evolving volatiles vesiculated some of the rock and formed the shardlike glass fragments. Residual iron-rich volatiles trapped in the glass may have altered some of the perlitic fragments to nontronite(?) and chlorite. The dikes of breccia and the widespread hydrothermal alteration and oxidation record the passage of much steam. Perhaps the Tertiary sediments were wet with ground water which not only would cause rapid chilling but also would supply volatiles to aid in the brecciation and alteration.

DACITE PLUG ON SOUTHEAST SPUR OF MOUNT LEWIS

GEOLOGIC RELATIONS

The plug of dacite on the southeast spur of Mount Lewis has the shape of an oval funnel tilted to the east. The long diameter is about half a mile, the short one about a quarter. The flow banding, which parallels the east wall, dips about 45° W.; that along the west wall was probably nearly vertical before intrusion of the breccia.

The plug cuts the Valmy Formation on the north, east, and south. A zone of breccia along the contact consists of angular fragments of the adjacent wallrock as much as 3 feet long.

The dacite has conspicuous, intricately folded flow banding. Locally one set of flow bands cuts another set. The banding conforms to the margins of the plug but is vertical in the center.

Near the east end of the plug is a small core of coarse block agglomerate and breccia. Angular to round blocks of dacite are set in a matrix of smaller dacite fragments and broken crystals of quartz, plagioclase, hornblende, and biotite, along with a few chips and pieces of Paleozoic rocks. This core resembles the block agglomerate in Indian Creek valley. Dikes of glassy dacite that merge with the surrounding dacite cut the core. The flow banding of the enclosing dacite swerves around the core and conforms in many places to the irregular outline of blocks forming the edge of the agglomerate. Apparently the core is not a central vent blasted through the dacite but rather an inclusion within it.

On the small knob outlined by the 8,400-foot contour is a small patch of breccia probably formed by steam explosions after the dacite had solidified. Angular blocks of flow-banded dacite have been jostled and tilted, but banding, almost continuous from one block to another, suggests that the blocks have not moved far. A red, very fine grained breccia fills cracks in the blocks and forms their matrix. This breccia consists of angular to round dacite fragments, ranging from about 5 mm to dust size, enclosed in a red, iron-stained siliceous matrix. Calcite and hematite are abundant; and the hornblende and biotite fragments in the breccia are black from oxidation.

PETROGRAPHY

In hand specimen the dacite shows phenocrysts of white feldspar about 1 to 2 mm long, black flakes of biotite, needles of hornblende, and a few quartz grains, set in a pale grayish-purple matrix of porcelaneous texture. Alinement of biotite and hornblende, thin streaks of fine dacite breccia, and contorted laminae of slightly differing colors contribute to the flow banding.

Under the microscope the feldspar phenocrysts are seen to be oscillatively zoned andesine, mostly saussuritized and veined by albite. Biotite and hornblende have been completely altered to chlorite, calcite, and magnetite. Fine magnetite dust rims many crystals. Quartz grains are rounded, embayed, and many are fractured. The groundmass is almost opaque from dusty small black flakes and fibers and contains irregular patches of calcite and chlorite. A few narrow streaks of brecciated groundmass and crystals are visible in some slides. Apparently the rock was brecciated during intrusion, perhaps because of shearing stresses in a very viscous magma.

RHYOLITE PLUG

GEOLOGIC RELATIONS

A rhyolite plug alongside the dacite plug is separated from it by only a thin septum of quartzite and chert breccia. (See fig. 23.) Its shape is elongated oval, about half a mile long. Like the dacite plug it has a rim of coarse breccia and a core of breccia surrounded

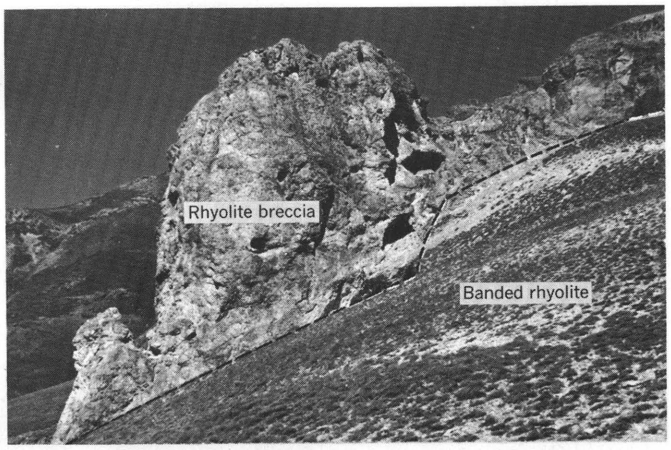


FIGURE 23.—Rhyolite plug (about 200 ft high) on the southeast spur of Mount Lewis. Core of breccia with banded rhyolite in foreground.

by banded porphyry. The flow banding roughly parallels the margins of the plug and also conforms to the core. (See fig. 24.) Locally, however, the banding has been broken and fragments are incorporated in the core.

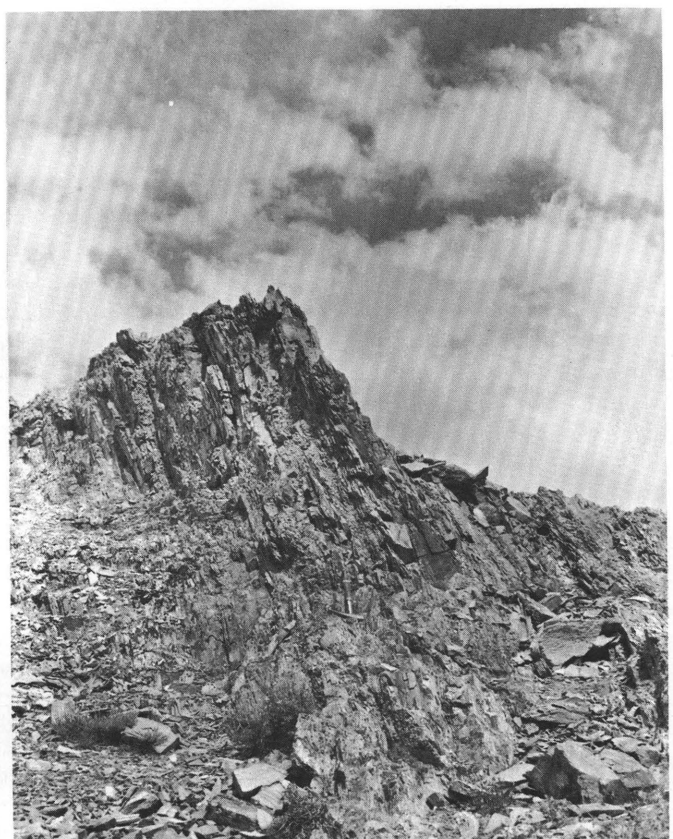


FIGURE 24.—Flow-banded rhyolite in plug on the southeast spur of Mount Lewis.

The core contains angular to rounded fragments of pebble, cobble, and boulder size, not only of rhyolite but also of dacite, quartz latite, quartz monzonite, and the Ordovician country rocks. Some of the rounded blocks are 10 feet in diameter. Most of the foreign fragments are confined to the central part of the core; the outer parts are primarily rhyolite breccia. A few thin dikes of rhyolite cut the core.

The rhyolite has more streaks and lenses of breccia than the dacite plug. Particularly near the margins the rhyolite is brecciated in lenses, streaks, and pods parallel to flow banding. This kind of breccia consists wholly of small angular fragments of rhyolite, with feldspar and quartz crystals, set in a red siliceous matrix, and probably formed by the shearing of a very viscous magma during intrusion.

PETROGRAPHY

The rhyolite is chalky to creamy white; yellow, red, or pale-pink flow banding gives the rock the appearance of ribbon candy. Small phenocrysts of quartz and feldspar in a porcelaneous groundmass, commonly iron stained, can be seen in hand specimen. Inclusions of Paleozoic rocks are common near the edges of the plug where the rhyolite is much brecciated, but are rare in the central part of the plug except in the breccia core.

The microscope shows that most phenocrysts are of hydrothermally altered sanidine and rounded and embayed subhedral quartz; many are broken. A few samples contain subordinate phenocrysts of albite. Sparse clots of magnetite and chlorite suggest that the rhyolite once contained biotite. The groundmass has been completely devitrified; in some rocks it consists primarily of spherulitic growths of very low birefringence, in others of a fine-grained mosaic of quartz, feldspar, and fine shreds of sericite and chlorite.

Chemical analysis indicates that this rhyolite, the most siliceous rock analyzed, is a potash rhyolite rather than a soda rhyolite like that of the Pipe Canyon breccia pipe (column 13, table 7). Phenocrysts of quartz and sanidine together make up less than 5 percent of the rock, so most of the silica, potash, and alumina are in the groundmass.

ORIGIN OF THE RHYOLITE AND DACITE PLUGS

The upward flaring and flow banding of the two plugs and the glassiness of their rocks suggest that they may be the roots of domes. They resemble many domes described by Williams (1932, p. 146). The cores of agglomerate and breccia are inclusions enveloped in the banded dacite and rhyolite and are therefore older than these. The plugs were probably once vents for explosive eruptions, such as those that contributed the dacite block agglomerate and acidic tuffs of Indian Creek valley. They were once completely filled by the agglomerate and breccia now preserved only in the cores. Viscous magma invaded and engulfed the agglomerate and breccia and pushed the fragments upward. Although not enough of the original plugs remains for certainty, possibly the cores are roots of crumbling spires of viscous magma and breccia like those that form the blocky, brecciated tops and talus-covered flanks of many domes.

DIKES

DISTRIBUTION AND GEOLOGIC RELATIONS

Dikes and small intrusive pods or lenses abound in the Mount Lewis and Crescent Valley quadrangles. One group near the Granite Mountain stock is perhaps associated with it; others are clustered near the small stock of granodiorite at Tenabo. The most conspicuous dikes form the swarm in the western part of the range between Trout Creek and the mouth of Lewis Canyon. (See fig. 18.) These trend north, most are vertical, and many are as much as a mile long. They cut the breccia pipes and in turn are cut by the Basin-Range faults along the west side of the range. The concentration of the dikes of course must be related to some underlying source of magma, perhaps the same that fed the breccia pipes. Broad zones of slightly hornfelsed and recrystallized chert, siliceous shale, and quartzite in the general area of the dike swarm suggest such an underlying source of heat. No control for the orientation of the dikes is apparent in the structure of the Paleozoic rocks they intrude; they do not parallel any known faults or major joint system. The normal faults to the west truncate the dike swarm and are clearly younger.

The long curving Trout Creek fault is followed almost continuously for several miles from the end, on the ridge west of Goat Peak, by a dike of quartz latite porphyry. For some unknown reason dikes are much fewer south of this fault than north of it.

The dikes clearly formed by injection rather than replacement. Contacts are sharp, borders are chilled, and strata pushed apart. Where the dikes cut chert, siliceous shale, or quartzite the wallrocks have been little altered or recrystallized. Where the dikes cut limy shale or limestone, some calcite apparently recrystallized but such metamorphic minerals as tremolite, diopside, or phlogopite are absent.

PETROGRAPHY

Most of the dikes are of intermediate composition—granodiorite and quartz monzonite porphyry, dacite and quartz latite porphyry. Many are of rhyolite, particularly near the breccia pipes, and a few are of pyroxene quartz diorite, andesite, and basalt. As it was not always practicable in the field to distinguish between the various kinds of intermediate dikes, many have been lumped as quartz porphyry (Tqp) on the geologic map (pl. 1). Some rhyolite dikes also have been included under this symbol.

Intermediate dikes

The dikes classed as intermediate include some of granodiorite, quartz monzonite, dacite, and quartz latite porphyry. The dikes of quartz latite and quartz monzonite porphyry contain slightly more biotite and more K-feldspar than those of granodiorite and dacite porphyry. The only essential difference between the quartz monzonite-granodiorite porphyries and the quartz latite-dacite porphyries is in the crystallinity of the groundmass. The intermediate rocks are all so closely allied that they can justifiably be described together.

All the intermediate dikes in the quadrangles have the same mineral species, even including rare grains of allanite. In mineralogy they resemble the granodiorite of the Granite Mountain stock, the breccia of Mount Lewis, the dacite plug on Mount Lewis, and the quartz monzonite in the Horse Canyon breccia pipe. Plagioclase phenocrysts are oscillatively zoned andesine; quartz phenocrysts are rounded and embayed; hornblende and biotite are the ferromagnesian minerals. Phenocrysts of K-feldspar are rare. One dike in Horse Canyon has large perthitic phenocrysts of K-feldspar that enclose plagioclase, quartz, hornblende, and biotite and partly replace some plagioclase in precisely the same manner as the large K-feldspar phenocrysts in the quartz monzonite porphyry facies of the Granite Mountain stock. In the other dikes K-feldspar generally forms either a granular crystalline intergrowth with quartz in the groundmass or a fuzzy patchwork with quartz in a devitrified glassy groundmass. As in the Granite Mountain stock, K-feldspar clearly crystallized late.

An analysis of a quartz latite porphyry from Horse Canyon is given in column 9, table 7.

Rhyolite

Rhyolite dikes are most abundant near the breccia pipes. A few cut the granodiorite stock at Tenabo. The rhyolites generally are flow banded and many are mineralized with pyrite and pyrrhotite. Mineralogic descriptions and chemical analyses of representative potash and soda rhyolites have already been given. Potash and soda rhyolites were not distinguished in the field but most of the dikes sampled are soda rich and contain abundant albite phenocrysts.

Pyroxene-quartz diorite

A small pod and a dike of pyroxene-quartz diorite porphyry cut Ordovician rocks in Lewis Canyon. A dike near the south contact of the Granite Mountain stock also is of pyroxene-quartz diorite. The mineral composition of these dikes suggests that their parental magma may have assimilated carbonate rocks of the eastern facies.

In the dike in Lewis Canyon, plagioclase phenocrysts are euhedral, with marked oscillatory zoning, and are more calcic than those of the intermediate rocks—entire crystals average about An_{50}, cores are as calcic as An_{58}. Some have a thin rim of even more calcic plagioclase full of minute inclusions of pyroxene. Quartz phenocrysts are less abundant than in the intermediate rocks, are rounded and embayed, and have borders of small grains of pyroxene and calcite. Pyroxene forms phenocrysts as much as 1 mm long and small grains in the groundmass. It is diopsidic augite—optic angle=54°, and $Z \wedge c=40°$. There are also ghosts of olivine(?) and hypersthene(?) phenocrysts, now filled with chlorite and calcite. The groundmass consists of small laths of labradorite, a little interstitial quartz and K-feldspar, and patches of calcite and chlorite.

The rims of calcic plagioclase, the abundant calcite, and the diopsidic pyroxene all suggest that lime and carbon dioxide were added to the magma. The fresh feldspars and pyroxenes indicate that the calcite is probably not hydrothermal. The carbonate rocks of the lower plate of the Roberts thrust are the most likely source of both carbonate and lime.

An analysis of this rock is given in column 3, table 7.

The pyroxene-quartz diorite porphyry dike south of the Granite Mountain stock also contains phenocrysts of labradorite and of diopsidic augite as well as clots of pyroxene, actinolite, and axinite—perhaps recrystallized inclusions of carbonate rocks from the lower plate.

Hornblende andesite

Hornblende andesite dikes have already been described in connection with the Horse Canyon breccia pipe. Very few were found except close to the breccia pipes. One from the crest of the ridge west of Goat Peak, the least altered example found, was analyzed (column 2, table 7). The hornblende andesite dikes in the Horse Canyon breccia pipe cut the late quartz latite and rhyolite dikes and are thus probably the youngest intrusive rocks in the Mount Lewis quadrangle.

EAST FLANK OF THE SHOSHONE RANGE

The east side of the range lacks bordering normal faults like those on the south, west, and northwest sides and its slope is governed primarily by the eastward tilting of the range as a whole. Three units are recognized among the Tertiary rocks. The oldest, presumably of the same general age as the other intermediate to rhyolitic extrusive rocks, consists of a few patches of rhyolite welded tuff resting unconformably on Paleozoic rocks. A thick gravel in turn rests unconformably on the welded tuff and Paleozoic rocks, and is overlain conformably by flows of basaltic andesite that make up a prominent cuesta, shown on the map as the Mal Pais. The basaltic andesite flows may be of late Pliocene or early Pleistocene age.

RHYOLITE WELDED TUFF

Small patches of rhyolite welded tuff are scattered along the east flank of the Shoshone Range from Corral Canyon nearly to the south edge of the map area. Most of these bodies are interbedded with gravels but some rest unconformably on the Paleozoic rocks. These masses doubtless represent erosional remnants of formerly more extensive bodies such as occur south of Red Rock Creek on the west flank of the range or in the northern Toiyabe Range and Cortez Mountains in the Cortez quadrangle. Wherever the relation between the tuff and the widespread basaltic andesite nearby can be determined, the tuff underlies the andesite. Across Crescent Valley to the southeast, however, similar rhyolitic rocks overlie the basaltic andesite on the dip slope on the east side of the Cortez Mountains.

The welded tuff is gray, pale green, or purple. Some exposures show foliation or sheeting; others are massive with vertical, but not columnar, jointing. Locally the base of the tuff is very dense and glassy. Phenocrysts of quartz and sanidine, and oxidized biotite are visible in hand specimen. In thin section the rock shows euhedral phenocrysts of quartz, sanidine, and oxidized biotite in a devitrified matrix of shards and collapsed pumice fragments.

GRAVEL

Late Tertiary gravel (Ts) fills a large valley along Corral Creek to a depth of at least 800 feet. The valley is cut in Paleozoic rocks and appears to have had much the same trend as the present Corral Creek. Gravel rests unconformably on a few small patches of rhyolite welded tuff south of the mouth of Mud Spring Gulch and also forms a few lenses between the flows of basaltic andesite lava that overlie the main valley fill of gravel. Some rhyolitic tuff is also interbedded with the gravel in Corral Canyon, and between Tub Spring Gulch and Indian Creek.

The gravel slumps readily and so is poorly exposed. Well-rounded to subangular pebbles and cobbles (of Paleozoic rocks, granodiorite from the Granite Mountain stock, and Tertiary volcanic rocks), very coarse bedding, and a few lenses of sand all clearly indicate a stream deposit, perhaps originally an alluvial fan.

BASALTIC ANDESITE

In the northwestern part of the Crescent Valley quadrangle, flows of basaltic andesite (Tba on pl. 1) crop out on the east side of Slaven Canyon and form a prominent cuesta, the Mal Pais, which is tilted about 5° E. (See fig. 25.) Two normal faults, apparently extensions of the Corral Canyon fault that cuts the Granite Peak stock, border the cuesta on the west. On the ridge east of Slaven Canyon the flows rest unconformably on rocks of the Valmy Formation. East of the normal faults, the flows rest both on rocks of the Valmy and on the gravel. The flows dip under the Quaternary alluvium of Crescent Valley.

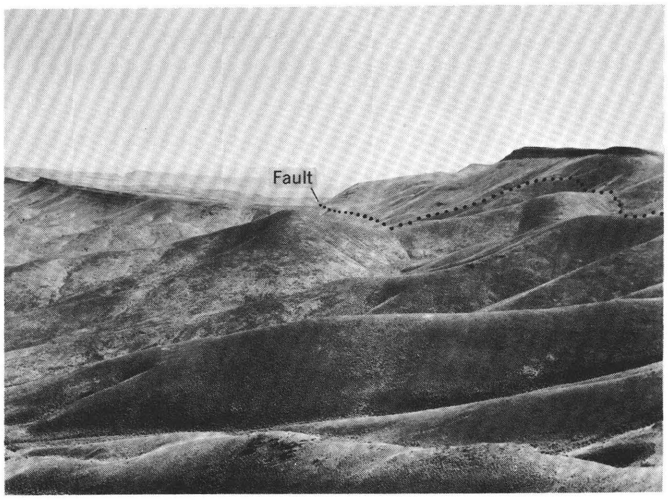


FIGURE 25.—Cuesta of basaltic andesite on east side of Slaven Canyon, looking northeast. A normal fault, downthrown to the northwest, has offset the lava sheet so that the rocks at the top right are the same flows as form the cuesta on the left.

The basaltic andesite is about 200 feet thick at Corral Canyon and thickens to at least 950 feet at the north edge of the quadrangle. Because the flows form the dip slope of the cuesta and strike approximately parallel to the trend of the Shoshone Range, this increase in thickness reflects a lower prelava topography toward the north rather than merely later tilting and subsequent erosion. The flows apparently overlapped southwestward against the Shoshone Range, burying the piedmont gravels and then the Paleozoic bedrock. To what extent the lavas originally buried the range is not known.

Dikes that might have been source fissures for the basaltic andesite are absent in the Shoshone Range. In the small part of the Cortez Mountains included in the southeast corner of the Crescent Valley quadrangle there is a swarm of dolerite dikes whose mineralogy is much like that of the basaltic andesite. These may well have been feeders for the flows, though of course the feeders may be buried beneath alluvium in Crescent Valley.

In Fire Creek a small area of brightly colored siliceous sinter, opal, and chalcedony and a surrounding zone of much-altered lavas indicate that hot-spring activity followed deposition of the lavas. Hot springs with much sinter, opal, chalcedony, and sulfur are presently active on Hot Springs Point, where similar basaltic andesite flows also rest unconformably on rocks of the Valmy Formation. The Beowawe geysers (Nolan and Anderson, 1934) lie about 6 miles northeast, on the strike of the cuesta of the Mal Pais, and the fault that is inferred to control their hydrology is the continuation of a branch of the Corral Creek fault.

Individual basaltic andesite flows range in thickness from 5 to 30 feet and generally can be traced several miles. Some have several sets of joints normal to their bases, but regular polygonal columns are absent. A few flows have very platy jointing and excellent flow banding parallel to their bases. The platy flows contain few vesicles but otherwise have the same minerals and textures as the jointed ones.

The nonplaty flows are very vesicular. At the base, vesicles are small, relatively sparse, and generally highly distorted in an irregular swirling pattern. The grain size of plagioclase microlites averages about 0.02 mm. The middle parts of the flows tend to have numerous large vesicles, some as long as 6 mm, mostly very much flattened and elongated parallel to the base of the flow. Some vesicles have completely collapsed and remain only as short streaks of chalcedony. Plagioclase laths average about 0.15 mm in length. The upper parts of the flows have many large vesicles, but these become fewer, smaller, and less distorted toward the top. In the upper 6 inches or so, the vesicles are small, very irregular, and accompanied by ropy structures and angular inclusions of dense, nonvesicular lava.

Several flows apparently moved like a tractor tread. In these, upper and lower zones of breccia, each about 10 feet thick, enclose a middle zone of vesicular lava 15 to 20 feet thick. The breccia consists of slabs from 2 to 6 inches thick and from 2 to 10 feet long, made up of glassy, platy, nonvesicular chilled lava. Some of these slabs are bent like a shallow bowl and others have a very shallow S-shape in cross section. Generally they parallel the flow tops. The slabs are in a matrix of vesicular lava. Contacts between lava and slabs look sharp in hand specimen but in thin section appear welded. The top of the flow apparently chilled and crusted over. Continued movement of the lava beneath broke the chilled top into thin slabs, which then mixed with the lava and were carried over the nose of the moving lava and buried beneath the flow like the caterpillar treads on a tractor.

The basaltic andesite is black to dark gray where fresh and weathers reddish brown. The only minerals visible with a hand lens are a few minute scattered phenocrysts of plagioclase and pyroxene. Most vesicles are unfilled, but many have a thin lining of chalcedony.

The microscope shows that the basaltic andesite flows have an intersertal or an intergranular texture and are monotonously alike in mineralogy. The main minerals are plagioclase, pyroxene, and magnetite. A few flows have a very little interstitial quartz. Plagioclase forms euhedral laths, generally of microlite size, but as much as 2 mm long in the centers of some of the thick flows. Where flow banding is present, the plagioclase laths are alined parallel to the banding. Otherwise, they form a feltlike network. They are unaltered and commonly show two or three albite twins, some gradationally zoned from about An_{65} in the interior to An_{55} near the edge. A few slides contain one or two large euhedral phenocrysts of labradorite of the same composition as the small laths but more clearly and sharply zoned.

Pyroxene is in small anhedral to subhedral grains. Most is augite, whose optic angle ranges from 44° to 55°, is slightly zoned, and has a more calcic rim. In a few slides, grains of augite as large as 1 mm contain a kernal of pigeonite. The contact between augite and pigeonite is abrupt but not so sharp as to appear as a definite line under the microscope.

Magnetite forms both euhedral crystals, averaging about 0.5 to 1 mm in diameter, and fine dust in the interstitial glass. Partial enclosure of the corners of some of the euhedral grains by augite or plagioclase suggests that the magnetite began to crystallize first.

The interstitial glass is colorless to pale brown; its refractive index is about 1.51. Much is slightly devitrified. Many vesicles have a thin lining of glass and a fringing growth of tridymite, opal, and chalcedony.

DIKES IN THE CORTEZ MOUNTAINS

It has been mentioned (p. 83) that the swarm of north-northwest-trending dikes on the steep frontal scarp of the Cortez Mountains may represent feeders for the basaltic andesite flood lavas of the Mal Pais and of the dip slope of the Cortez Mountains southeast of this area. Single dikes are as much as a mile long and nearly 800 feet thick and occupy much of the Cortez Mountains in the Crescent Valley quadrangle. They cut some quartz latite dikes near the ridge crest, and thus are younger than some of the siliceous volcanics, just as is the basaltic andesite. This is not a close bracket of age, however, as siliceous volcanic plugs cut the basaltic andesite farther south on the east slope of the Cortez Mountains.

The dikes are of basalt, now considerably altered. Stubby laths of labradorite are intergrown with subhedral to anhedral grains of subcalcic augite to give an intergranular texture. There are a few phenocrysts of plagioclase and euhedral augite. Small irregular grains of quartz are probably of hydrothermal origin. Pyroxene grains are rimmed with opaque oxides, and most of the plagioclase laths are veined by albite and replaced by irregular patches of calcite and abundant dusty saussurite.

HYDROTHERMAL ALTERATION OF THE IGNEOUS ROCK

The rock descriptions make it clear the hydrothermal alteration of the igneous rocks has been widespread in the Shoshone Range. The most intense alteration and mineralization occur near and in the breccia pipes and near concentrations of small dikes and plugs, particularly in the breccia pipes of Rocky, Pipe, and Horse Canyons, along the range front near the mouth of Lewis Canyon, in the Maysville Summit and Hilltop areas, and near Tenabo and Gold Acres. The main areas of relatively unaltered igneous rock are the Granite Mountain stock (although some of its satellitic dikes are altered) and the Tertiary volcanic rocks in the valleys of Harry and Cooks Creeks. Of approximately 400 thin sections, a representative sampling of the igneous rocks of the Shoshone Range, at least three-fourths show some alteration and one-half have been extensively recrystallized. The close areal correlation of alteration with the breccia pipes, dikes, and plugs indicates that much of the hydrothermal activity probably accompanied or closely followed the volcanic activity.

The hydrothermal solutions primarily added CO_2 and H_2O to the rocks. Calcite has partly replaced plagioclase, hornblende, and biotite in most of the altered rocks. In groundmasses it forms irregular blebs and patches. Most of the necessary CaO was released during alteration of the plagioclase. Some andesine crystals are extensively veined by albite and many have patches of residual albite and streaks and blebs of clay. Epidote, clinozoisite, sericite, and prehnite also commonly replace plagioclase. Hornblende and biotite have been altered to chlorite, magnetite, ilmenite, quartz, calcite, epidote, and in the case of biotite, to K-feldspar.

K-feldspar was little affected; most shows only a slight dusting of sericite. In some dikes containing large phenocrysts of K-feldspar, the surrounding matrix is highly altered, but the phenocrysts and the minerals enclosed in them are clear except for thin marginal zones of sericite, calcite, and chlorite. This suggests that the K-feldspar was remarkably stable despite the extensive alteration of the surrounding rock.

Hydrothermal solutions have also altered the Paleozoic sedimentary rocks. Normally dark cherts, argillites, and siliceous shales have been bleached to a white, light-gray, or creamy color except for much red iron staining along bedding planes and joints. Locally there are thin veins of quartz and small knots and scattered crystals of pyrite and pyrrhotite.

CHEMICAL COMPOSITION

The chemical compositions of representative Tertiary igneous rocks, both intrusive and extrusive, altered and unaltered, are given in table 7, and plotted in figure 26, showing the variation in the weight percentage of the principal oxides as silica content varies. Spectrographic analyses of minor elements are listed in grams per metric ton in table 8.

TABLE 7.—*Chemical analyses and norms of Tertiary igneous rocks, Mount Lewis and Crescent Valley quadrangles, Nevada*

[Analyses 4, 5, 6, 8, 9 by L. D. Trumbull; all others by L. N. Tarrant]

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO_2	52.98	53.75	56.32	62.81	64.12	64.96	67.61	67.98	69.29	71.31	73.13	76.84	79.18	55.89
Al_2O_3	15.65	15.75	14.36	16.21	15.52	16.96	15.48	14.62	16.77	14.57	12.51	12.40	11.51	14.70
Fe_2O_3	2.39	2.20	.84	2.01	2.14	.88	.72	1.56	.56	1.50	.54	.31	.28	2.90
FeO	5.58	5.31	5.29	2.15	1.87	1.62	2.73	1.43	.49	.76	.18	.45	.24	7.37
MgO	4.93	5.62	7.64	2.78	1.78	3.18	1.73	.99	.65	.46	.20	.01	.07	3.98
CaO	7.70	6.85	5.91	4.68	2.56	4.42	3.65	2.83	3.03	2.12	1.57	.35	.10	7.39
Na_2O	3.06	3.58	1.94	2.78	4.69	3.29	3.14	3.67	3.90	3.50	1.74	3.93	.49	3.11
K_2O	2.50	2.28	2.74	2.59	3.13	.92	3.31	3.06	3.71	4.43	5.37	4.60	5.87	1.84
H_2O^{-1}	.42	.13	.29	.74	.39	.53	.13	.37	.15	.31	1.54	.14	.49	.48
H_2O^{+1}	.89	2.00	3.01	2.20	1.61	2.53	.73	1.24	.65	.31	3.24	.38	1.17	.34
TiO_2	1.57	1.36	.58	.52	.50	.64	1.45	.41	.41	.33	.10	.04	.04	1.36
CO_2	1.34	.24	.55	.01	1.16	.02	-------	1.15	.03	.00	.01	.20	.03	.02
P_2O_5	.57	.56	.15	.19	.17	.20	.16	-------	.19	.13	.03	.01	.01	.46
S	.00	.02	.03	.01	-------	.00	.01	.02	.02	.02	.01	-------	.06	------
MnO	.14	.12	.12	.09	.08	.03	.07	.06	.02	.05	.01	.02	.01	.16
BaO	.03	.07	.12	.06	.13	.06	.03	.11	.11	.07	.04	.08	.22	.03
F^{-1}	-------	-------	-------	.03	.03	-------	-------	.05	.03	-------	-------	-------	-------	------
Less O	-------	.01	.02	+.01	-------	-------	.01	.03	-------	-------	-------	-------	.03	------
Total	99.75	99.83	99.87	99.86	99.88	100.24	100.94	99.52	100.01	99.87	100.22	99.76	99.74	100.03

TABLE 7.—*Chemical analyses and norms of Tertiary igneous rocks, Mount Lewis and Crescent Valley quadrangles, Nevada*—Continued

[Analyses 4, 5, 6, 8, 9 by L. D. Trumbull; all others by L. N. Tarrant]

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
C.I.P.W. norms														
Q	3. 18	2. 70	8. 46	21. 78	16. 50	27. 90	24. 90	27. 18	25. 38	29. 40	38. 88	35. 10	53. 34	9. 36
or	15. 01	13. 34	16. 12	15. 57	18. 35	5. 56	19. 46	18. 35	21. 68	26. 13	31. 69	27. 24	35. 03	10. 56
ab	25. 68	30. 39	16. 24	23. 38	39. 82	27. 77	26. 72	30. 92	33. 01	29. 34	14. 64	33. 54	4. 19	26. 20
an	21. 68	18. 35	22. 52	22. 80	11. 95	21. 13	17. 24	13. 62	14. 46	9. 73	7. 78	1. 67	. 56	20. 85
wo	5. 45	4. 99	2. 78	-------	. 12	-------	-------	-------	-------	-------	-------	-------	-------	6. 61
en	12. 30	14. 10	19. 10	7. 00	4. 50	8. 00	4. 30	2. 50	1. 60	1. 10	. 66	-------	-------	8. 60
fs	5. 94	5. 68	8. 05	1. 58	1. 06	1. 06	3. 83	. 66	-------	-------	-------	. 66	. 13	9. 11
il	3. 04	2. 58	1. 22	. 91	. 91	1. 22	. 76	. 76	. 76	. 61	. 15	-------	-------	2. 58
mt	3. 48	3. 25	1. 16	3. 02	3. 02	1. 39	. 93	2. 32	. 46	1. 62	. 46	. 46	. 46	4. 18
hm	-------	-------	-------	-------	-------	-------	-------	-------	. 32	. 32	. 16	-------	-------	-------
ap	1. 34	1. 34	-------	. 34	. 34	. 34	. 34	. 34	. 34	. 34	-------	-------	-------	1. 34
C	-------	-------	-------	. 31	-------	2. 86	. 41	. 41	1. 12	. 51	1. 02	. 31	4. 08	-------
Niggli values														
qz	−3. 8	−5. 2	19. 8	70	55. 4	96. 4	100. 4	117. 6	111. 8	139. 8	214	231. 2	300	12. 8
k	. 35	. 29	. 48	. 38	. 30	. 16	. 41	. 36	. 38	. 46	. 67	. 43	. 89	. 28

DESCRIPTION OF ANALYZED SAMPLES

1. Basalt from small knob near the mouth of Harry Creek, sec. 2, T. 28 N., R. 44 E.
2. Hornblende andesite from dike on Goat Ridge, sec. 28, T. 29 N., R. 45 E.
3. Pyroxene-quartz diorite porphyry from dike in Lewis Canyon, sec. 26, T. 30 N., R. 45 E.
4. Dacite tuff breccia, south end of Tertiary area in Indian Creek valley, sec. 17, T. 29 N., R. 46 E.
5. Dacite porphyry of the block agglomerate, highly altered and calcitized. Head of Dean Canyon, sec. 36, T. 29 N., R. 45 E.
6. Quartz diorite porphyry, slightly altered; sec. 3, T. 29 N., R. 46 E.
7. Granodiorite, porphyritic, from Granite Mountain stock, sec. 11, T. 29 N., R. 46 E.
8. Quartz latite breccia, slightly altered, from Mount Lewis; sec. 12, T. 29 N., R. 45 E.
9. Quartz latite porphyry from dike in Horse Canyon, sec. 4, T. 30 N., R. 45 E.
10. Quartz latite welded tuff, hilltop south of Mill Creek, sec. 6, T. 28 N., R. 45 E.
11. Rhyolite, from low hill north of mouth of Lewis Canyon, sec. 14, T. 30 N., R. 45 E.
12. Soda rhyolite, Pipe Canyon breccia pipe, sec. 34, T. 30 N., R. 45 E.
13. Rhyolite, plug on southeast flank of Mount Lewis; sec. 7, T. 29 N., R. 46 E.
14. Basaltic andesite, east of Slaven Canyon, sec. 7, T. 30 N., R. 47 E.

NOTE.—The variation diagram of figure 26 takes into consideration analyses 6 and 7, even though they represent rocks of an earlier cycle than most of the others; their general chemistry suggests close similarity of the magmas. Analysis 14 has been included in this diagram also, despite the probable considerably younger age of the rocks it represents. However, it seems chemically so different—in its relatively lower K_2O considering its silica content—as to suggest a different magma type as the source.

TABLE 8.—*Semiquantitative spectrographic analyses, in grams per metric ton, of trace elements of the intrusive and volcanic rocks of Miocene and Pliocene and earlier ages*

[Analyst, Paul R. Barnett. Localities given in table 7]

	1	2	3	4	5	6	7	8	9	10	11	12	13	Average igneous rock [1]
B	20	-------	-------	10	-------	-------	-------	-------	-------	10	-------	-------	30	3
Ba	1, 000	1, 000	2, 000	700	2, 000	1, 000	1, 000	1, 000	1, 000	2, 000	500	1, 000	2, 000	250
Be	-------	-------	-------	1	1	1	1	1	1	2	1	1	4	2
Co	30	30	40	10	9	6	9	9	-------	-------	-------	-------	-------	23
Cr	100	100	200	100	60	60	30	30	4	5	1	3	3	200
Cu	20	40	30	9	6	20	80	80	8	5	3	4	4	70
Ga	20	20	10	20	20	20	10	10	20	10	8	9	10	15
La	100	100	-------	50	80	-------	90	90	90	100	-------	-------	-------	19. 6
Nb	30	-------	-------	8	10	10	30	30	6	40	20	40	20	24
Ni	70	80	90	20	5	10	7	7	-------	3	-------	-------	-------	80
Pb	20	20	50	20	20	5	30	30	20	30	30	50	30	16
Sc	30	20	30	20	20	20	10	10	6	6	3	4	2	5?
Sr	1, 000	1, 000	700	900	800	700	800	800	900	700	400	300	100	300
V	200	200	200	90	90	100	90	90	40	40	6	-------	-------	150
Y	60	60	40	40	40	30	40	40	20	40	20	30	10	7. 4
Yb	4	4	3	2	3	3	3	3	1	3	2	4	2	2. 7?
Zn	200	200	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	-------	51
Zr	100	100	60	60	100	60	70	70	100	90	40	60	70	220

[1] Data from Green (1953, table 2).

Looked for but not found: Ag, As, Au, Bi, Cb, Cd, Ge, In, Mo, Pt, Sn, Ta, Th, Tl, U, W.

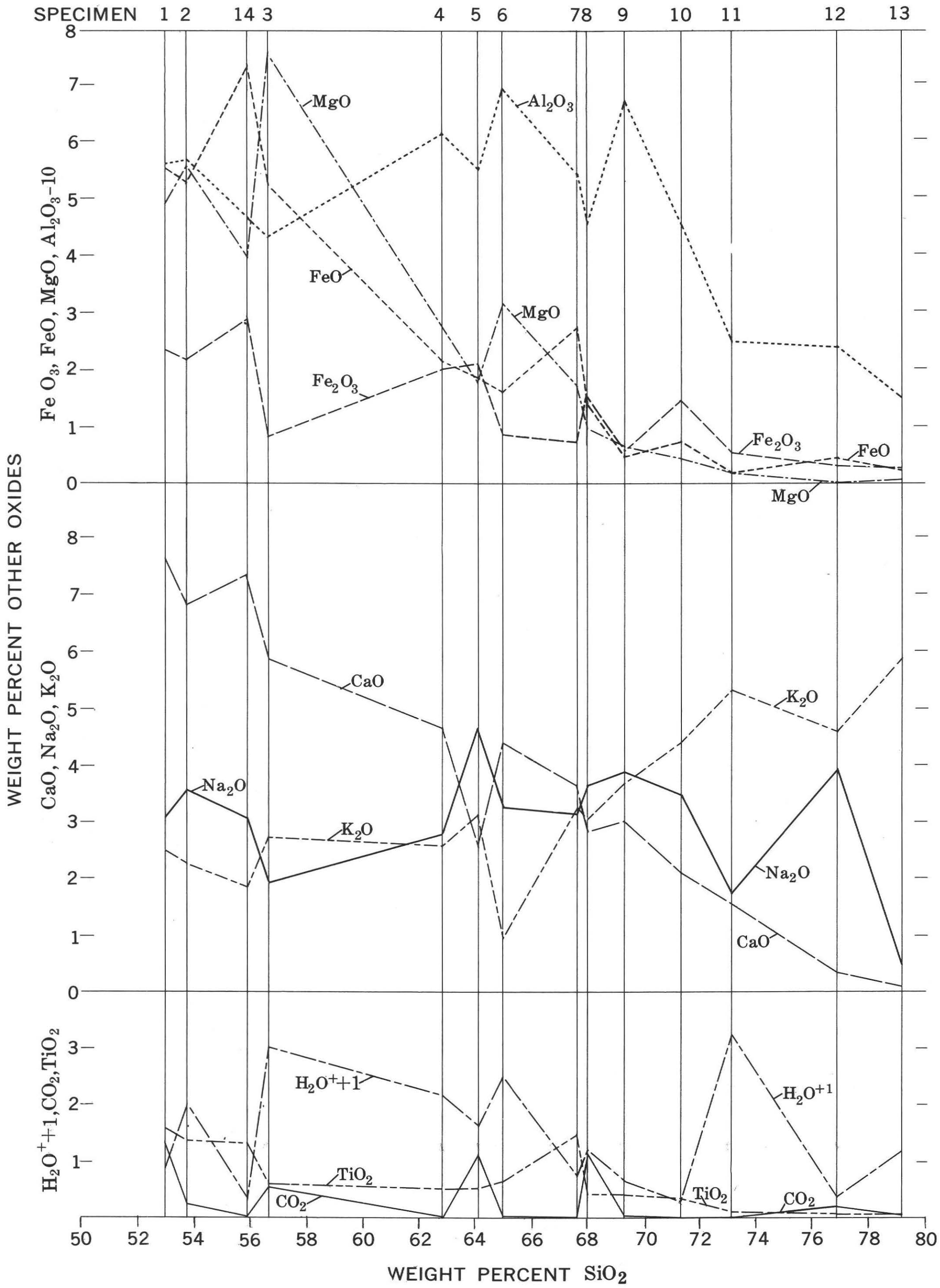


FIGURE 26.—Variation diagram, showing relations of other major component oxides to silica, in the Tertiary igneous rocks of the Mount Lewis and Crescent Valley quadrangles.

The most persistent trends among the main oxides, interrupted only by departures in the altered rocks, are a rise in K_2O, particularly in the siliceous rocks, and a decline in CaO. MgO and $FeO+Fe_2O_3$ show a precipitous decline among the intermediate rocks, but there is a marked variation in the ratio of FeO to Fe_2O_3. The ferric iron exceeds the ferrous primarily in the altered rocks and in the extrusive welded tuffs. Na_2O has a very irregular trend, probably in large part because of differences in alteration. Among the minor elements V, Cr, Ni, Co generally decline as silica increases. Sr declines slightly and Ba, Zr, and Cu show considerable variation.

Much of the scatter can be attributed to hydrothermal alteration. In analyses 5 and 8, the rise of Na_2O and decrease in CaO reflect albitization of plagioclase, and the relatively high content of CO_2 and H_2O results from hydrothermal calcite and chlorite. In the rock of analysis 6, the plagioclase shows only slight veining by albite, so that loss of CaO has been slight; but most of the hornblende and biotite have altered to chlorite so that the water content is high.

Among the minor elements, strontium and barium are most abundant. Strontium closely parallels CaO in falling off with increasing silica; this suggests that strontium is substituting for calcium in the plagioclase. Both barium and strontium are two to three times as abundant as in other intermediate to acid igneous rocks as reported by Rankama and Sahama (1950, p. 472–473, 476). The consistence of the trend of strontium suggests that it was not introduced during hydrothermal alteration. Barium has a considerable scatter, perhaps more apparent than real, as the analyses are given only to the nearest hundred parts per million. Barium commonly substitutes for potassium in K-feldspar and biotite (Rankama and Sahama, 1950, p. 472–473). Although Ba parallels the trend of K_2O, the correlation is not as close as that between Sr and CaO. The richness in barium suggests that the Tertiary igneous activity of the Shoshone Range is the source of the many deposits of barite in the general area.

On a regional scale, the Tertiary igneous rocks of the Shoshone Range are intermediate in potassium content between those of the eastern Basin and Range province and the Sierra Nevada to the west. Several geologists have commented on the abundance of Tertiary igneous rocks of intermediate composition in the Basin and Range province that are relatively rich in potassa (Lindgren, Graton, and Gordon, 1910, p. 36; Butler, 1920, p. 89; Nolan, 1935, p. 50; Merriam and Anderson, 1942, p. 1723–1725). Merriam and Anderson plotted the Niggli k ratios of many Tertiary igneous rocks from California, western Nevada, and the rest of the Basin and Range province. The rocks of the Shoshone Range are in this respect about comparable with rocks from western Nevada and intermediate between those of California and of the eastern Basin Ranges.

AGE OF THE VOLCANIC AND ASSOCIATED ROCKS

A few vertebrate bones, shells of fresh-water invertebrates, and fragments of plants are the only identified fossils found in the Tertiary rocks of the Mount Lewis and Crescent Valley quadrangles. The vertebrate fossils were collected from tuffaceous silty sandstone 250 feet west of the southwest corner of the Mount Lewis quadrangle. G. E. Lewis of the U.S. Geological Survey identified them as:

Merychippus sp., metapodials II, III, and IV (fragments) with some corresponding phalanges.
Camelid cf. *Procamelus* sp., front part of proximal end of metacarpals III–IV.

Mr. Lewis stated that the age of these is "probably late Miocene, although early Pliocene is possible," and that the fragments are similar to Merriam's "Barstow."

The vertebrate fragments came from a thick sequence, which extends many miles to the south, of fine ash, tuffaceous sandstone and siltstone, pebble conglomerate, and lake beds. Van Houten (oral commun., 1955) considered these beds to be part of his "vitric tuff unit," which he (1956, p. 2814–2819) correlated with the middle of the Humboldt Formation and with the Truckee Formation of Late Miocene to early and middle Pliocene age.

Not more than 100 feet stratigraphically lower and 3,500 feet farther north, a coquina of fresh-water clams and snails was found. These fossils were examined by Dwight W. Taylor of the U.S. Geological Survey, who reported as follows:

USGS Cenozoic locality 20901 (F–154). Lander County, Nev., Mount Lewis quad., 600 ft W., 3,500 ft N., of SW cor. of quadrangle. Collected by J. Gilluly, 1954.
Sphaerium n. sp.
Bulimnea cf. *B. megasoma* (Say)
Planorbidae indet.

In the Nevada area the known range of this assemblage is late Miocene and early Pliocene. This is the characteristic suite of mollusks found in northern Nevada, southern Idaho, and northwestern Utah in upper Tertiary rocks often included in the Salt Lake Formation. It does not occur as high as middle Pliocene rocks and may not range throughout all lower Pliocene rocks.

A single fresh-water clamshell was found by E. M. Shoemaker in float from the tuff about 3,000 feet farther northeast. The horizon is probably a few hundred feet stratigraphically higher than either the vertebrates or the other mollusks. This position would thus tend to confirm Mr. Taylor's tentative opinion for he reported as follows:

USGS Cenozoic locality 20899 (F–63), Lander County, Nev., Mount Lewis quad., 3,500 ft W., 1,200 ft N. of SE cor. sec. 26, T. 28 N., R. 44 E., collected by E. M. Shoemaker, 1952.
Sphaerium n. sp.

The species does not contribute much useful stratigraphic information because it is not known from other localities. The most closely related species are middle Pliocene or younger, and in the Nevada area the known early Pliocene or Miocene species are clearly distinct from this one. The *Sphaerium* of locality 20901 nearby is also distinct, and a little more likely to be older than younger, but this is not at all certain.

Best guess of the age is middle Pliocene or younger.

Putting the inferences from the vertebrate and invertebrate fossils together, an early Pliocene age for the tuffs in the southwest corner of the Mount Lewis quadrangle seems most likely. From admittedly cursory reconnaissance in the adjacent Mount Moses quadrangle, it seems that the fossiliferous beds are high, perhaps nearly at the top of the volcanic accumulation.

Plant fossils found in claystone lenses in the block agglomerate of Indian Creek valley were identified by the late R. W. Brown of the U.S. Geological Survey as:

Sequoia affinis Lesquereux
Pinus sp.
Fragments of dicotyledonous leaves.
Acer sp.
Prunus sp.

For the first three, Brown suggested an age of late Eocene to late Tertiary and for the last two the latter half of the Tertiary.

The rocks of Indian Creek valley that contain the plant remains are in a small fault block isolated by surrounding Paleozoic rocks from the beds in which the vertebrate fossils were found.

Because the extrusive rocks occur in local isolated areas and fossils are very sparse, no stratigraphy for the Tertiary System as a whole can be established. It is assumed that all the extrusive rocks of Tertiary age date from the late Miocene to early Pliocene except the basaltic andesite flows of the Mal Pais. These are considered late Pliocene because they rest unconformably on the presumed upper Miocene to lower Pliocene rocks.

The youngest rocks cut by the breccia pipes are of Early Triassic(?) age. The breccia pipes are considered as late Miocene to early Pliocene because their rocks are closely like the intermediate volcanics. The dikes and plugs that cut the breccia pipes and the siliceous plugs on the southeast flank of Mount Lewis probably all belong to the same general episode. It has already been pointed out that the intrusive rocks are of two distinct ages, as not only are boulders of granodiorite from the Granite Mountain stock in the conglomerate beneath the volcanics in Indian Creek valley, but also quartz latite tuff rests unconformably directly on the granodiorite east of Tub Spring gulch. Erosion had therefore exposed this granodiorite prior to the later Tertiary surficial volcanism.

QUATERNARY SYSTEM

Much of the Mount Lewis quadrangle and more than half of the Crescent Valley quadrangle are covered by Quaternary deposits: colluvial, alluvial, lacustrine, and eolian. This survey was primarily concerned with bedrock geology and little attention was given these materials. The several varieties of Quaternary deposits are not generally distinguished one from the other on plate 1 and indeed little of the colluvium and talus that mask considerable areas is mapped at all. The following comments may serve to outline the broad features of the Quaternary deposits, but a map of the surficial geology was not attempted and much additional work could profitably be directed toward it.

COLLUVIUM

A discontinuous veneer of colluvium is widespread in the mountainous parts of the quadrangle. Certain areas, such as the north slopes of Mount Lewis, the hills near the head of Indian Creek, the slopes of Horse Mountain, and many other places, have great sheets of talus streaming down their slopes. Many of these sheets are so large as to require mapping on the scale of plate 1. But wherever they could be reasonably ignored, they were, and the bedrock geology was projected across them. Their detailed mapping would have rendered an already complex bedrock map almost illegible. In addition to the talus, disrupted fragments and weakly coherent rubble interrupt the outcrops—especially those of shale, sandstone, and argillite.

ALLUVIUM

Alluvial deposits are the most widespread geologic bodies in the area. Both the Reese River Valley and Crescent Valley are almost completely masked by stream gravels and sand, and broad terraces extend into the range along Cooks and Elder Creeks and Slaven Canyon. Other wide terraces, separated from the valley fill by rock gorges, are found along Feris, Indian, Corral, and Rock Creeks, well toward their headwaters. (See fig. 27.)

The materials composing these alluvial deposits are all locally derived. Most near and in the mountains are poorly sorted gravels with boulders as much as 2 feet in diameter. Downstream they grade to sand and even, toward the center of Crescent Valley, to silt. Sorting is generally poor and there are sporadic interbeds of mudflow material—largely masses of nonsorted gravel, silt, sand and boulders, some several feet across—exposed in stream cutbanks for a mile or more from the mountain front. Few of these deposits are cemented, at least at the surface, though some pits 20 feet deep show some caliche cement.

Alluvial deposits of two ages are readily distinguishable along most of the mountain valleys. The older deposits form terraces that stand at heights as great as 50 feet or as little as 6 or 8 feet above the alluvial plains composed of the younger stream gravels. In general, these terraces are considerably wider than the modern flood plains. Locally the modern flood plains have very recently been themselves dissected, and there are thus two sets of terraces above a very narrow present-day stream channel.

Valleyward from the mountain fronts the terraces rapidly become lower and less definite and toward the

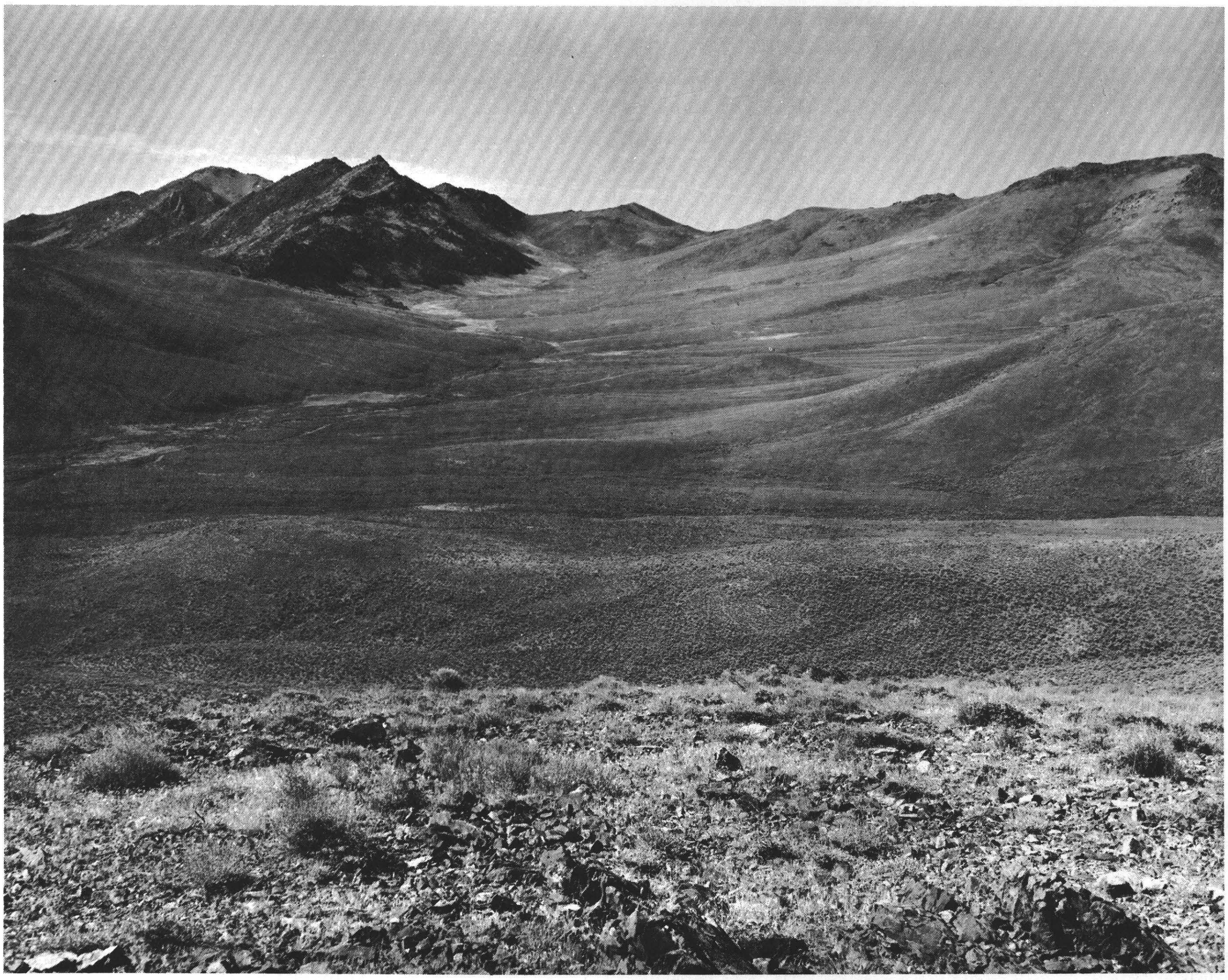


FIGURE 27.—View looking south, up The Park, toward the head of Corral Canyon, showing alluvial cones in the depression west of the scarp of the Corral Canyon fault. Granite Mountain at the left center.

axes of Crescent and Reese River Valleys they are no longer distinguishable. No attempt was made to distinguish older and younger alluvium on plate 1.

PLAYA AND LACUSTRINE DEPOSITS

Playa clays and silts are exposed in the southeastern part of Crescent Valley, southeast of the Dean Ranch, and are being added to during wet seasons. They probably are more abundant at depth, for the region was much more humid during parts of the Pleistocene than it now is and the nearby Grass Valley and Carico Lake Valley both held considerable lakes at times. Doubtless Crescent Valley, too, had a permanent lake or expanded playa during the Wisconsin and earlier glaciations. Such a lake, if present, was shallow and had shore features so feebly developed that they have since become masked by encroaching fans. Excavations along the road from the Dean Ranch to Cortez and, just south of the map area, along the branch road to Mill Canyon (Cortez Mountains), both show lacustrine clays overlain by thin alluvial veneers. Alkali playas also extend northward as far as the valley floor west of Hot Springs Point.

DUNES

Windblown silt and sand has accumulated in small dunes a few feet high to the northeast of the playas in Crescent Valley. These dunes are largely anchored by vegetation and are now growing only slowly.

VOLUME OF QUATERNARY SEDIMENTS

The local relief of the bedrock surface of the Shoshone Range is about 2,000 to 3,000 feet. On the east flank of the range, this irregular surface has been tilted gently southeastward at an angle, measured by the cuesta slope of the basaltic flows, of roughly 4° or 5°. The slope evidently increases or continues unbroken in the same direction for several miles out into Crescent Valley, for there are no inliers of bedrock exposed in the gravel-fan cover beyond a zone about a mile wide along the mountain base. Only at Hot Springs Point and along the foot of the Cortez Range does bedrock again appear—in each locality beyond an obvious fault. Evidently the alluvial fill in Crescent Valley must be several thousand feet thick in order to conceal the originally irregular upper surface of the bedrock over so wide an area.

In an effort to obtain more information on this point, Messrs. Donald Plouff and Samuel Stewart, of the U.S. Geological Survey, measured the value of gravity at 50 stations in and near Crescent Valley, in the Crescent Valley and Cortez quadrangles. Mr. Plouff's complete discussion is in the section "Gravity survey of Crescent Valley." It suffices to point out here that, on the assumption of a very considerable hypothetical density difference of 0.4 between valley fill and bedrock, there is a minimum thickness of 8,000 feet of valley fill near the middle of Crescent Valley. (See fig. 52.)

A thickness of valley fill of 1 to 2 miles along the deepest part of Crescent Valley is therefore practically demonstrated. How much of this thickness should be attributed to the Quaternary and how much to the Tertiary is impossible to determine from available data. The sedimentary parts of the Tertiary section include much poorly coherent and highly porous material which cannot differ greatly in density from the Quaternary gravels and sands. Similarly, much of the welded tuff of the Tertiary is of comparable density. Accordingly, the Quaternary part of the thick valley deposits may be many thousands of feet thick or merely a few hundred—it is at present impossible to tell.

STRUCTURAL GEOLOGY

FIELD CRITERIA OF THRUSTS

The map (pl. 1) exhibits many thrusts whose existence is obvious from the transection of lithologic members of the upper plate along surfaces whose intersection with the topography can only indicate relatively consistent dips and extensive continuity across the strike. Faults such as the Bateman fault along the north wall of Bateman Canyon were recognized by these criteria as thrusts before we had fossil evidence to show the footwall to be Devonian and the hanging wall Silurian and Ordovician.

But many features mapped as thrusts on plate 1 are by no means as readily demonstrable. Some appear to lie nearly parallel with the bedding for long distances, and others, though transecting the bedding beneath, lie nearly parallel to the beds of the hanging wall and thereby form map patterns that resemble those of unconformities as well as thrusts. It is indeed possible that one or two of the contacts of Elder Sandstone on the Valmy or of Slaven Chert on the Elder that are here mapped as thrusts represent instead only minor shearing along an unconformity that brings widely differing lithologies in contact. This is certainly not true, however, of the vast majority of the thrusts mapped on plate 1; they are mechanical contacts as demonstrated by one or more of the following criteria:

1. Fossil evidence, showing either that older beds lie on younger or that notably younger beds lie on older, without the intervention of beds of intermediate age known to be represented elsewhere in the map area.
2. Geometric evidence, showing either (a) sporadic transection of hanging-wall beds (despite general parallelism to the boundary); (b) breccia lenses of quartzite, chert or other resistant rocks strewn discontinuously along the contact; or (c) obviously dragged beds in the footwall block. Such breccia masses are commonly cemented by silica and stand out conspicuously on the hillsides and can readily be traced even though beds above and below appear nearly concordant. When so traced, within a few hundred feet or at most a mile or two, the breccia horizon eventually is seen to cut across the bedding of footwall, hanging wall, or both. As noted above, enough of these features have been demonstrated on fossil evidence as thrusts so that we are relatively confident that almost all that are so mapped have been correctly diagnosed.

METHODS AND PRECISION OF ANALYSIS

The fundamental data used in the structural analysis attempted here are of course those compiled on the geologic map (pl. 1). The geologic sections (pl. 2) illustrate our interpretation of the geometric relations of the many structural units at depth. In drawing these sections, the structural units have been projected as far as their continuity in surface extent and plunge seemed to justify. It has seemed best, in this attempt, to err rather in the direction of too great, rather than of too little extrapolation. We thereby show our interpretation of the probable sequence of thrust sheets (their topology) at depth even though we recognize the quantitative (geometric) uncertainties. The reader will note, for example, that near the surface the sections illustrate detail essentially equivalent to that shown on the map. In general the uncertainties of the interpretations increase, perhaps exponentially, with distance of projection.

Obviously, in absence of drilling, the smaller details of structure cannot be confidently projected very far. We have therefore used conventions to suggest the varying degrees of our confidence in our interpretations—near the outcrop the maximum detail as derived from the map is shown in patterns like those of the map. Where we feel that a structural unit may be projected but that the details of its makeup are more doubtful, we have shown abrupt terminations, first of members patterns and then of formation patterns. Finally we have projected the facies of the pre-Pennsylvanian rocks as units—upper plate contrasted with lower plate of the Roberts thrust. Another way of expressing this convention is to say that we are far more confident of fault extensions than we are of the details of the lithologic composition of the individual fault slices. We believe the map pattern supports this view.

No attempt has been made to project individual formations of the eastern facies to depth. The reasons are fairly clear from the map. Not only are these formations exposed over areas too small to permit confident interpretations of major fold plunges, but they are cut by faults less systematically arranged than those of the upper plate of the Roberts fault.

Projections have been continued, where possible, above the present land surface. Naturally this extrapolation as well as the extrapolation to great depths cannot be considered as much better than diagrammatic. Perhaps the Roberts fault in areas such as that near Mount Lewis lies at double the depth shown on sections such as those from *C–C′* to *G–G′*. Perhaps it lies at somewhat shallower depth than shown, though error in this direction seems to us likely to be much less than in the other. The general relations shown in these and other sections are nevertheless considered qualitatively supported by the areal pattern, dips, and plunges recorded on the map.

Exposed plunges of major thrust sheets have encouraged us to project individual formations to great depths in some areas; on the other hand we have been forced to generalize at much shallower depths in areas where plunges are unfavorable to extrapolation.

GENERAL FEATURES OF THE STRUCTURE

Because of the contrast in facies between carbonate and clastic strata, a mere description of the stratigraphy has demanded some discussion of the major structural features of the Mount Lewis-Crescent Valley area. This section of the text has the object of analyzing these and other features to clarify their geometric and chronologic relations, and, as far as possible, the genetic inferences to be drawn from them. In order to introduce the necessarily complex discussion of the structural evolution of the area, a summary of our interpretations is presented here, rather categorically. Detail of evidence and inferences from them are presented in subsequent sections. Supporting evidence is presented in appropriate sections of the text.

The structure of the northern Shoshone Range records orogenic deformation of at least four different ages. The first two involved great folds and thrusts; the last two, chiefly normal faulting and block tilting.

The earliest notable deformation seems to have taken place in Early Mississippian to Early Pennsylvanian time. This was when the Roberts thrust and Trout Creek transcurrent fault formed (see p. 42), along with considerable folding both of the main thrust surface and of the rocks above and below it. This folding involved higher subsidiary thrust scales, such as the isoclinally folded Kattenhorn thrust, wholly within the upper plate, as well as the Roberts thrust itself. Notable among the faults in the upper plate (see pl. 6) are the Bateman, Hilltop, Crum Canyon, Kattenhorn, Chert Ridge, Crippen Canyon, and Dean Mine thrusts to the north of the Trout Creek fault, and the Harry Creek, Cooks Divide, Mill Creek, Goat Ridge, North Fork, Shoshone Summit, Feris Creek, Utah Mine, Elder, Greystone, and Indian Creek thrusts to the south. The Slaven Canyon fault is represented at the surface both north and south of the Trout Creek trend. (See pl. 6.) The deformation that produced these structures was the Antler orogeny of Roberts (1949).

Erosion greatly reduced the relief that must have been produced during the Antler orogeny and this was followed by the deposition, in presumably relatively nearby areas, of Battle Conglomerate, Antler Peak Limestone and, after at least minor disturbances, the China Mountain(?) Formation.

The second epoch of strong deformation followed. This orogeny doubtless folded still further the already contorted upper plate of the Roberts thrust, with its many thrust scales. It superposed still higher thrust plates upon them—plates that include Battle, Antler Peak, China Mountain(?), Havallah, and Harmony Formations, as well as more plates of pre-Carboniferous formations identical with those involved in the older Roberts thrust. Thrusts confidently attributed to this episode include the Whisky Canyon, Havingdon Peak, and Pipe Canyon group (pl. 6; see p. 50).

The age of this deformation cannot locally be more closely bracketed than post-Lower Triassic(?) to pre-Miocene. Erosional gaps and conglomerates and other clastic rocks at several horizons of the Middle and Upper Triassic sections in the Golconda (Ferguson, Roberts, and Muller, 1952); Winnemucca (Ferguson, Muller, and Roberts, 1951a); Mount Tobin (Muller, Ferguson, and Roberts, 1951); and Mount Moses (Ferguson, Muller, and Roberts, 1951b) quadrangles imply considerable orogenic disturbance nearby throughout the Triassic. The presence of two adjacent but highly contrasting facies of Triassic rocks in those quadrangles demands that the Golconda thrust, which separates them, is a major thrust of post-Triassic age. The authors cited refer the Golconda thrust, which telescoped the Triassic facies, to Jurassic time.

This assignment was doubtless influenced by the fact that strong orogenic movements have been demonstrated in the general region at two distinct times during the Jurassic: Early Jurassic (Liassic) in the Pilot and Gabbs Valley Ranges (Ferguson and Muller, 1949, p. 13) and Late Jurassic—the Nevada orogeny—in the Sierra Nevada (Blackwelder, 1914, p. 643–645). In the Eureka district the Newark Canyon Formation, of Early Cretaceous age, unconformably overlaps highly deformed Permian rocks but is itself involved in notable folds (Nolan, Merriam, and Williams, 1956, p. 68–70). Thus we have evidence of post-Early Cretaceous deformation in the general region as well as the several older episodes. The absence in the Shoshone Range of strata whose ages are intermediate between Early Triassic(?) and Miocene, makes the age of the post-Early Triassic(?) thrusting here a matter of conjecture. To avoid premature designation as either Early Jurassic, Nevadan, Middle Cretaceous or Laramide, the deformation is referred to in this report simply as the Lewis orogeny, because of structures assigned to it on Mount Lewis (p. 123).

It is noteworthy that no plutonic or volcanic activity has yet been recognized as related to either the Antler or the Lewis orogeny, although they are responsible for most of the large tectonic features of the area.

The third epoch of deformation is not readily separated from the fourth, or Basin Range faulting. But the plutonic and volcanic activity of Tertiary time was accompanied by some notable faults, many of which became sites of igneous injection. The local tilting and shouldering aside of the country rocks did not, perhaps, involve major orogeny, but it certainly did connote marked crustal unrest. The faults of the volcanic episode, so far as they can be recognized, all seem to be normal and doubtless most of them reflect adjustments of the rocks above to the movements within a magma chamber.

The latest deformation is the Basin Range faulting. At many other places in the Basin and Range province these faults have been shown to have begun their activity far back in Tertiary time, but no record of this faulting has been recognized here older than the basaltic andesite of probable Pliocene age, forming the Mal Pais on the east flank of the Shoshone Range north of Mud Spring Gulch. The northwest front of the Shoshone Range shows a complex of Basin Range faults all downthrown toward the Reese River Valley. The aggregate displacement must be several thousand feet. A similarly great fault separates Crescent Valley from the Cortez Mountains. Both these fault systems have been active recently enough so that scarps in the youngest alluvial cones are well preserved. The only considerable faults of this epoch recognized within the range are the relatively low-dipping fault near the mouth of Horse Canyon (which is thought to have dropped the higher parts of the Horse Canyon breccia pipe down toward the mountain front) and the Corral Canyon fault. The Corral Canyon fault has a throw of a few hundred feet at the north edge of the area and west of Granite Mountain, but must die out within a mile or two to the southwest.

ANTLER OROGENY, THE ROBERTS THRUST

DISCOVERY AND TYPE LOCALITY

The existence of the Roberts thrust (originally called the Roberts Mountain thrust) was first suggested by Edwin Kirk (1933, p. 31–32). Although he did not actually observe the fault, the close juxtaposition of extremely contrasting facies of Ordovician rocks in the Roberts Mountains suggested to him the possibility of telescoping by thrusting. The fault was soon demonstrated by the mapping and stratigraphic studies of Merriam and Anderson in 1939 (Merriam and Anderson, 1942). The type locality of the thrust is in the Roberts Mountains, about 20 miles southeast of the map area.

The name Roberts thrust, as used in this report, is restricted to the fault that separates the carbonate (eastern) facies from the siliceous (western) facies. Faults that do not have this relation are not considered as parts of the Roberts, even though many were doubtless formed at the same time or even are branches from the main thrust.

REGIONAL RELATIONS

The Roberts thrust was demonstrated in the Roberts Mountains by the superposition of the Vinini Formation of Ordovician age upon rocks as young as Late Devonian. Two large and several small windows there expose the thrust (fig. 28). In the Roberts Mountains the rocks beneath the thrust belong to the same facies as the classic section at Eureka, and consist, except for the Eureka Quartzite, almost wholly of carbonate rocks. They include the Pogonip Group sensu lato (including Cambrian and Ordovician rocks), Eureka Quartzite (Middle Ordovician), Hanson Creek Dolomite (Late Ordovician), Roberts Mountains Formation and Lone Mountain Dolomite (Silurian), Nevada and Devils Gate Limestones (Devonian). Over these rocks on a warped thrust surface lies the Vinini Formation of Early and Middle Ordovician age. The Vinini greatly

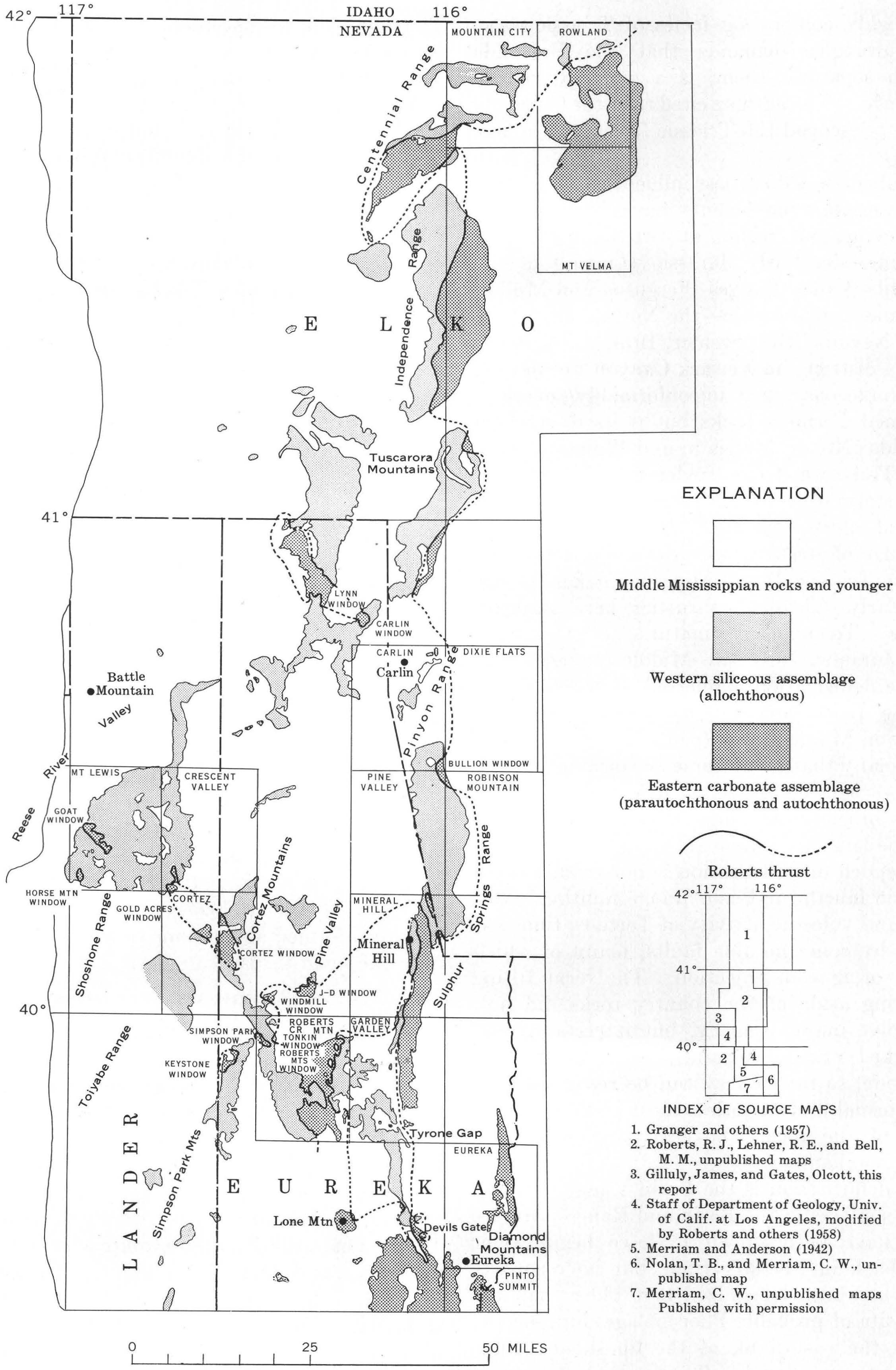


FIGURE 28.—Regional relations of the Roberts thrust (modified from fig. 4 of paper by Roberts and others, 1958).

resembles parts of the Valmy Formation of the Shoshone Range, in consisting largely of chert, shale, quartzite, and sandstone with subordinate greenstones, but differs in being generally finer grained. There are relatively far fewer quartzites in the Vinini than in the Valmy and a notably higher proportion of shale, some of which is highly organic. Greenstone, though abundant in the Valmy, is not a major constituent of the Vinini. No Late Ordovician fossils have been reported from the Vinini; the Valmy is in part Late Ordovician. The two formations nevertheless are clearly of the same facies and probably originally formed a continuous body; they differ dramatically from the chronologically equivalent formations beneath the thrust.

Relations generally similar to those in the Roberts Mountains have since become widely recognized in north-central Nevada. In the Cortez quadrangle (see fig. 28) a large window in the thrust exposes Hamburg Dolomite, Eureka Quartzite, Hanson Creek Formation, Roberts Mountains Limestone, and Devonian limestone, with some Pilot Shale, all beneath a clear-cut thrust whose upper plate consists of slices of Vinini, Valmy, Elder Sandstone, Slaven Chert, and a thick siliceous mudstone of Silurian age (Gilluly, 1954; Gilluly and Masursky, 1965). Although differing in minor degree from the classic Eureka section (a possible facies change having thickened the Roberts Mountains Limestone at the expense of the Lone Mountain Dolomite), the rocks beneath the thrust clearly belong to the eastern carbonate facies. The rocks of the upper plate as clearly represent the sharply contrasting western siliceous facies. They differ from the rocks of the upper plate in the Roberts Creek Mountains only in including rocks of Silurian and Devonian ages and in having representatives of both the Valmy and Vinini Formations. There is no suggestion of transition of either the siliceous or the carbonate facies toward the other; except for the Eureka Quartzite there are virtually no siliceous rocks in the lower plate and almost no carbonate rocks in the upper plate.

Reconnaissance mapping by Ralph Roberts and Robert Lehner of the U.S. Geological Survey has shown that in the Lynn window, in northern Eureka County (fig. 28), the lower plate of the thrust includes Hamburg(?) Dolomite, Eureka Quartzite, Hanson Creek Formation, Roberts Mountains Formation, and limestone of Devonian age, overridden by rocks referred to the Vinini Formation and shale whose fossils show Silurian affinities (Roberts and others, 1958, p. 2830, 2832–2833, 2835).

Roberts and Lehner have mapped a thrust between the carbonate and siliceous facies from the southern Independence Range to the Idaho border in the Rowland quadrangle, where it passes beneath volcanic rocks of the Snake River volcanic field (Roberts and others, 1958, fig. 4 and p. 2820).

The Carlin window (fig. 28) exposes only Roberts Mountains Formation beneath the Vinini (Roberts and others, 1958, fig. 3, p. 2820, 2834; J. Fred Smith, Jr., oral commun., 1959).

The area shown in figure 28 as the "Bullion window" exposes eastern facies rocks belonging to the Pogonip Group, the Eureka Quartzite, the Hanson Creek Formation, the Lone Mountain Dolomite, and the Nevada and Devils Gate Limestones. The siltstone that overlies these rocks is, however, not of western facies, even though it resembles the Vinini lithologically and was originally considered in reconnaissance as part of that formation (Roberts and others, 1958, fig. 4). It is of Mississippian age, and though perhaps thrust over the eastern facies rocks to form a window, is much more likely in normal depositional contact with them (J. Fred Smith, Jr., and K. B. Ketner, oral commun., 1960). Whether it is disconformable or allochthonous, there is here no evidence for the Roberts thrust, and figure 28 is insofar in error.

In the Mineral Hill area, the eastern facies formations represented are the Eureka Quartzite, Hanson Creek Formation, Roberts Mountains Limestone, Lone Mountain Dolomite, and the Nevada Formation, over which are thrust siliceous beds of Silurian (C. A. Nelson, oral commun., 1954) and Devonian (Carlisle and Nelson, 1955) ages, as well as unmistakable quartzite of the Valmy (our observation). The Devonian beds of the upper plate here include much chert, chert conglomerate, black shale, quartzite, and a considerable amount of limestone conglomerate and clastic limestone. They thus represent a possible transition between eastern and western facies. If this thrust is the Roberts, as it appears to be, this Devonian section is the only one that strongly suggests a transition between the facies of the upper and lower plates, though the increased silt content of the Roberts Mountains Limestone in the Mill Creek and Horse Mountain windows in the Shoshone Range points also toward a facies change that lessens the contrast between upper and lower plates.

Between the Roberts Mountains and Cortez lie four windows: the J–D Ranch, Windmill, Tonkin and Simpson Park, and just to the southwest of the last, the Keystone. (See fig. 28.) All these were discovered by Roberts and Lehner during their reconnaissance survey of Eureka County. None have been mapped in detail but all contain limestones of Silurian and Devonian age and are overridden by siliceous beds of Ordovician (Vinini?) and Silurian age.

To the south of the Roberts Mountains the thrust has been inferred to extend through the Antelope Range and at least as far as the Toquima Range, 20 miles east of Austin (C. W. Merriam, oral commun., 1950). These localities are not shown on figure 28, but are indicated on figure 1. There are suggestions in abrupt facies contrasts that the Roberts thrust or one with similar relations extends north as far as central Idaho and southwest at least to the Inyo Range.

In summary, a thrust reasonably correlated with the type Roberts thrust has been identified for a strike length of nearly 180 miles (lat 42° N. to 39°25′ N.) and may extend much farther. It is exposed in windows for a width transverse to the strike of at least 55 miles (Horse Mountain window, Mount Lewis quadrangle to Mineral Hill). Wherever exposed in this large area, a surface of thrusting separates a carbonate sequence beneath from a siliceous sequence above; we regard it as a single thrust.

A case can be made, though it is not compelling and is not here adopted, for considering the displacement still greater (Roberts and others, 1958, p. 2851.) This would require correlation of the Roberts thrust with the Adelaide thrust of the Sonoma Range far to the west, but the facies contrasts on the Adelaide thrust do not resemble those of the Roberts and a connection between them remains unproved. We feel that the Roberts thrust has not yet been identified west of the Reese River Valley.

STRUCTURAL SIGNIFICANCE OF THE ROBERTS THRUST

Although local features along the contacts of eastern and western facies throughout this area clearly result from mechanical disruption, the question may nonetheless be raised as to whether it is necessary to interpret the locally demonstrated faults as all connected beneath the surface and hence as forming parts of a single structure. It is perhaps conceivable that orogeny might cause localized shearing at a facies boundary throughout the area and that the present "windows" are really only reefs and shallows of carbonate (eastern facies) deposits, surrounded, or nearly so, by parautochthonous siliceous (western facies) deposits of deeper water facies. On this hypothesis the present distribution of facies differs only slightly from their depositional pattern.

Although this possibility must be considered, there are many reasons for rejecting it and interpreting each of the geometrically demonstrable local shear surfaces that separate the two facies as a part of a single surface of translation upon which the rocks of the western facies have moved eastward for a minimum distance of 55 miles.

Consider, for example, the fact that the Eureka Quartzite and Hanson Creek Formation maintain their essentially unchanged lithologies and thicknesses from Eureka to the Goat window and thence, to the Lynn window. Neither formation shows any suggestion of transition toward the lithology of the virtually contemporaneous Valmy or Vinini Formations (though both of these contain individual beds of quartzite quite as pure as that of the Eureka). It seems incredible that at each of the isolated windows and at the main eastern outcrop of the Roberts thrust a relatively small mechanical disruption should follow an original facies change so precisely as to leave no siliceous tongues in the contemporaneous carbonate facies and no carbonate tongues in the siliceous.

To this improbability is added the fact that in many windows the clean separation of facies is not restricted to Ordovician rocks but takes place in precisely the same sense between rocks of Silurian age and also in rocks of Devonian age (with the one exception mentioned at Mineral Hill). Because no Cambrian rocks have yet been proved present in the upper plate of the Roberts thrust, the probable but not quite certain admixture of greenstone in the Cambrian Shwin Formation of the Goat window is not an exception to this generalization. We know nothing of the facies changes in rocks of Middle Cambrian age that can directly aid in evaluating the extent of movement on the Roberts fault.

In the light of the consistent facies separation involving rocks deposited at times in the long interval from Early Ordovician to Late Devonian, it seems impossible to assume that the eastern facies was deposited in its presently exposed complex pattern in a sedimentary province elsewhere dominated by siliceous sediments, also in roughly their present patterns. Pyroclastics and pillow lavas of andesitic composition are common in the Valmy and present, in less volume, in the Vinini. No traces of such rocks or of fragmental materials derived from them have been found anywhere in the carbonate facies of comparable age. The siliceous volcanics of the Elder Sandstone are quite unrepresented in the carbonate facies of Silurian age.

A further argument that the facies contrast represents actual translation of one facies over the other for long distances is found in the consistency of each facies within itself. The Ordovician section of the carbonate facies is essentially the same throughout the broad area of its exposure here discussed—except for the overlap of the Eureka already discussed. The very existence of the systematic overlap northwestward from Eureka to Cortez (p. 11) argues for continuity of the facies involved. It is completely inexplicable if the Cortez carbonate facies were not continuous with that of the Roberts Creek Mountains and then with that of Eureka. If these carbonate areas really were isolated from each other by the siliceous facies presently separating them there would be no reason at all for such a systematic overlap.

The Silurian rocks of the carbonate facies are also essentially similar except for changes from dolomite to limestone toward the north and west. So, too, with the Devonian. Yet there are considerable secular differences within the carbonate facies: through the Eureka Quartzite, the dolomite and limestone beds of the Hanson Creek Formation, the thin bedded limestone of the Roberts Mountains Formation, to the thicker bedded limestones of the Devonian. To assume that both carbonate and siliceous facies are parautochthonous over the large area of north-central Nevada illustrated in figure 28 is to assume that parallel facies changes went on through geologic time in precisely the same sequence and with precisely identical isolated patterns while facies changes equally drastic but quite independent of them were proceeding over the same region on a regional scale.

There is also a structural argument against the interpretation of the rocks of the siliceous facies as parautochthonous. The thrusts within the siliceous facies involve demonstrable transport measured in miles (see the Bateman fault, the Utah Mine, and the Kattenhorn faults on pls. 1 and 6 for examples chosen almost at random). Such transport is quite in keeping with a large translation of the whole siliceous facies with respect to the carbonate facies but seems altogether inconsistent with an assumption that the facies boundary is one of only minor displacement.

The discussion of the preceding pages seems in a way to belabor the obvious. It seems nonetheless necessary, for several of our colleagues, unfamiliar in detail with the implications of the facies contrasts, have suggested original depositional patterns as an explanation of the facies distribution in the region. Although the necessity of appeal to major thrusting has already been pointed out by many writers (Merriam and Anderson, 1942, p. 1702–1704; Nolan, Merriam, and Williams, 1956, p. 34; Roberts and others, 1958, p. 2851), the great travel implied by this interpretation of the Roberts thrust as a single movement horizon naturally demands consideration of all conceivable alternatives before it is accepted. We have been able to find no satisfactory alternative, however.

THE DATE OF THE ROBERTS THRUST

The dating of the Roberts thrust was based for many years only upon indirect evidence. Hague (1892, p. 165, 177, 203) inferred that the conglomerate he considered as belonging to the Upper Coal Measures in the Eureka area was a record of uplift and erosion to the west, but as the Roberts fault had not been recognized, of course neither he nor Ferguson (1924, p. 37), who found Permian rocks overlapping deformed Ordovician beds at Manhattan, connected this orogeny with the formation of the Roberts thrust. Nolan (1928) inferred the presence of a late Paleozoic geanticline extending north-northeastward across Nevada and later support for this interpretation was found by Roberts (1951) and Dott (1955). That the Roberts thrust was formed as part of this late Paleozoic orogeny was, however, for a long time unproved and there are suggestions, indeed, that movements on the thrust may include some considerably younger than this.

As reviewed above, the Roberts thrust cuts rocks as young as the Pilot Shale and so is clearly post-Late Devonian. In the Carlin area conglomeratic beds of Early Mississippian age, containing large boulders of both western facies siliceous rocks and eastern facies carbonates, overlap the fault and both of its bordering facies (J. Fred Smith, Jr., and K. B. Ketner, oral commun., 1957). This seems definitely to date the thrust as Early Mississippian in the Carlin area. In the Eureka area the orogenic sediments began to form only in Late Mississippian time, when the lower, conglomeratic part of the Diamond Peak Formation was deposited as intertongues with the Chainman Shale (Nolan, Merriam, and Williams, 1956, p. 58; Dott, 1955, p. 2268).

It is perhaps not surprising that a structure so large as the Roberts thrust should have been active at different times in different segments, as the interpretations of the preceding paragraph imply. We know of several younger orogenic disturbances in the general region (p. 50), and it is not at all unlikely that the fault has been several times rejuvenated in different segments.

In the area near Tyrone gap, Garden Valley quadrangle (fig. 28), the Vinini is unconformably overlain by limestone, sandstone, shale, and conglomerate of the Garden Valley Formation (Nolan, Merriam, and Williams, 1956, p. 67–68) of Permian age (Wolfcamp or Leonard). The Garden Valley Formation is highly deformed and partly overturned. No closely similar rocks of this age are known farther east—the nearest, the Carbon Ridge Formation, is much finer grained and richer in carbonate beds. Accordingly, Nolan, Merriam, and Williams considered that the Garden Valley Formation has been transported, during the Roberts thrusting, along with the Vinini Formation on which it lies, in order to bring it into its present proximity with the Carbon Ridge.

The Garden Valley and Carbon Ridge Formations do indeed present marked facies contrasts, as was widely recognized even before they had been formally named. The extremely rapid facies changes in similar beds such as the Tonka Formation of Dott (1955, p. 2226) and the Diamond Peak and Chainman Shale of the Eureka area (Nolan, Merriam, and Williams, 1956, p. 57–58), however, led Dott to question the necessity of fault telescoping between Garden Valley and Carbon Ridge (Dott, 1955, p. 2274). Even though the postulated telescoping near Eureka may have occurred in post-Leonard time, it seems not to negate the evidence at Carlin that the major movement on the Roberts thrust took place in Early Mississippian time. There is ample evidence in the Antler Peak quadrangle (Roberts, 1951) of pre-Pennsylvanian major thrusting and it seems likely that the Roberts thrust belongs to the same general epoch.

ROBERTS THRUST IN THE NORTHERN SHOSHONE RANGE

In the Mount Lewis and Crescent Valley quadrangles our direct knowledge of the relations of the Roberts thrust is derived from its exposures in half a dozen windows, large and small. In the following section we first describe the fault as it is exposed in these windows, with only so much discussion of the geology above and below as seems necessary to this purpose. We then discuss the geology of the lower plate and finally, of the upper plate. Although it seems more than likely that all the structures have been somewhat modified by the post-China Mountain(?) Lewis orogeny, we consider the structures below the Whisky Canyon fault to be mainly products of the Antler orogeny; they are treated as wholly of this epoch. The structures associated with the Whisky Canyon fault and those at higher tectonic levels are described under the heading "The Lewis orogeny."

GOLD ACRES WINDOW

Location.—In the area of this report the most easterly window exposing the Roberts thrust is that in which the mining settlement of Gold Acres is situated, on the west side of Crescent Valley at the south edge of the map area. Here the lower plate is exposed over an area of about a square mile.

Local features of the fault.—The eastern side of the window is buried by alluvium which conceals the fault; probably the large outcrops of carbonate rocks of early and middle Paleozoic age at Cortez, across Crescent Valley to the southeast, are a part of the same window, whose continuity is interrupted by the valley alluvium and the fault bounding the Cortez and Toiyabe ranges. To the northeast, the Roberts fault must be concealed beneath the alluvial fill of the arroyo that runs south-southeast through the eastern half of secs. 30 and 19, T. 28 N., R. 47 E. It is readily traced southwestward from this arroyo in sec. 19 where Devonian limestone and Pilot Shale in the footwall are overridden by greenstone of the Valmy Formation. The exposed fault is a crush zone that dips northward, northwestward, and westward at angles of 20° to 40°. Most of it is 2 to 3 feet thick, but locally the breccia is as much as 80 feet thick. The crush zone is composed almost wholly of chert, greenstone, and quartzite derived from the upper plate. Individual blocks form lenses a few inches to a few score feet long and from a few inches to a few tens of feet thick, held in a powdered matrix. None of these rocks show any signs of crystalloblastic metamorphism.

The Gold Acres gold mine is developed in a crush zone along the Roberts fault, marked by slickensides, drag folds, and breccia. (See p. 134.) Here the footwall formation is the black carbonaceous limestone of the Roberts Mountains Limestone. It is locally mildly metamorphosed and contains metacrysts of undetermined "contact" silicates, probably tremolite. No comparable metamorphic minerals were observed in the overlying Elder and Slaven Formations, but they are common in the limestones of Devonian age. In a most general way, the rocks of the upper plate tend to parallel the fault, but this probably represents local drag rather than regional parallelism, for different units, representing Valmy, Elder, and Slaven Formations, overlie the limestones of the window as the fault is followed southward. (See pl. 2, geologic section *12–12'*.)

GOAT WINDOW

The Goat window, which is on the slope west and southwest of Goat Peak, is the largest in the area and offers the most westerly known exposures of many formations of the eastern carbonate facies. This window is about 5 miles long and ranges in width from ½ to nearly 2 miles, extending from the North Fork of Mill Creek to the mountain front at Hancock and Trout Creek canyons.

Local features of the fault.—The geometric form of the surface of the Roberts fault is extremely complex. It can be seen on plate 1 and in figure 29 that the fault is fairly regular and simple along the south side of the window, where it dips southward and southwestward at angles of 20° to 50°. Southeastward from the canyon in which the Shwin Ranch lies the fault brings Valmy sandstone, chert, and greenstone over phyllites of the Shwin Formation, with only locally a few small horses of Hanson Creek dolomite and Roberts Mountains Limestone between them.

West of Shwin Canyon the footwall of the fault is formed by the Hanson Creek dolomite, with locally small scales of Roberts Mountains Limestone, whereas the hanging wall is much broken and slivered Valmy sandstone, greenstone, chert, quartzite, and shale. Northwest of the center of sec. 29, T. 29 N., R. 45 E.,

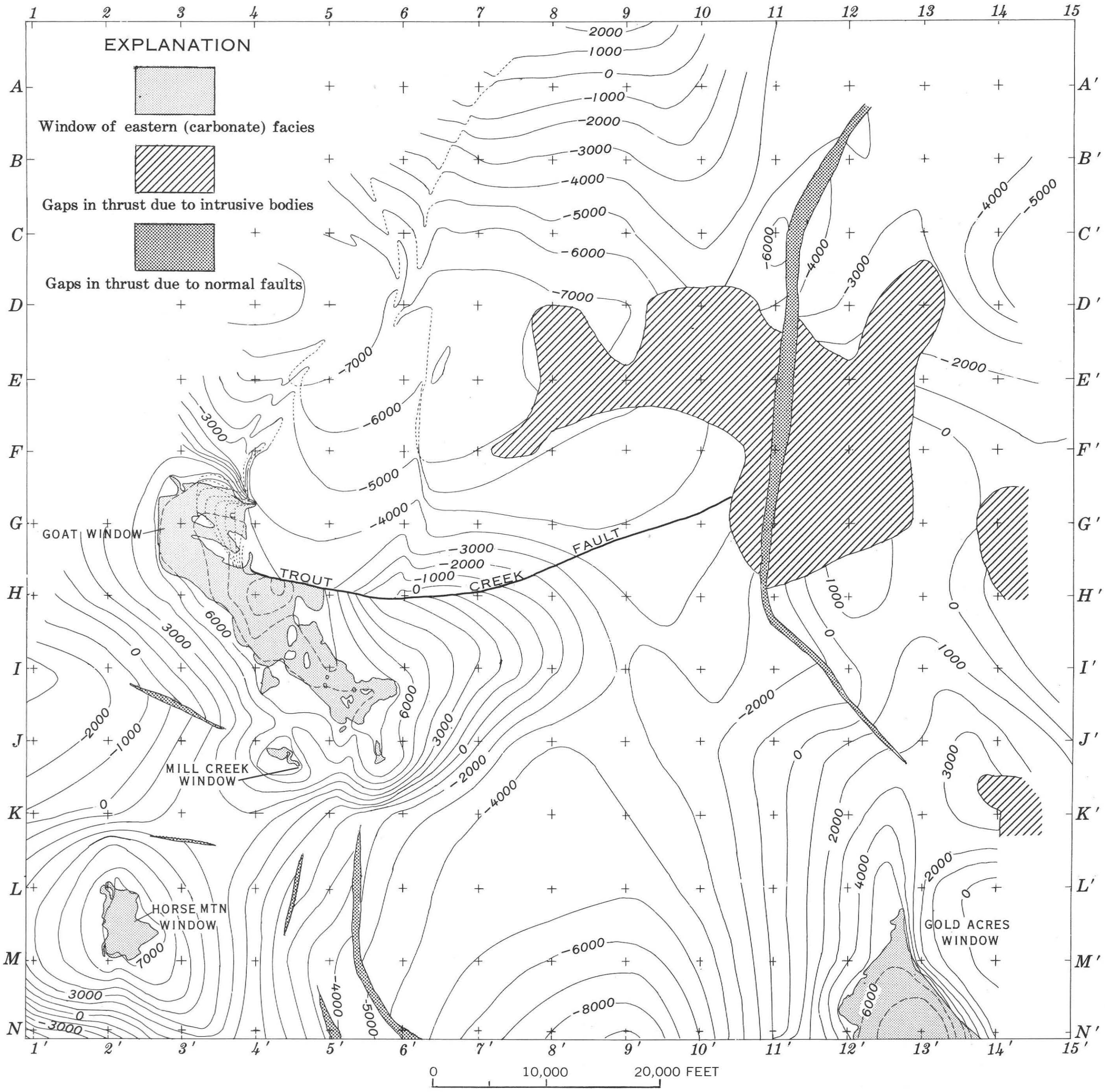


FIGURE 29.—Structure contours on the Roberts thrust as estimated from surface outcrops and projections along the plunges of the folds. Contours dotted where fault is overturned, dashed where eroded away. Contour interval 1,000 feet. Datum is mean sea level.

the fault nearly parallels the structure of the Hanson Creek in its footwall. Although buried by alluvium, the fault must curve rather sharply northward in the south half of sec. 19, T. 29 N., R. 45 E., and thence trend almost due north across the low spurs on either wall of Hancock Canyon, just below the steep mountain front in the west half of the same section. The fault here must dip about 40° or even more steeply, judging from the topography. Here the Roberts Mountains Limestone and Prospect Mountain Quartzite form the footwall for half a mile, then, to the north, the Eldorado Dolomite, with slivers of Shwin and Roberts Mountains. A wedge of Shwin Formation lies as a horse between these rocks and the Valmy, along the thrust which must here be concealed beneath the terrace gravels to the west.

Still farther north, where the thrust emerges from the terrace gravel, it must either dip steeply west or be dropped from sight, along this segment, by the Basin Range fault which, farther north, drops Valmy against Valmy. The block of Valmy sandstone beneath this high-angle fault to the northeast rests on the Shwin Formation. The fault trace between them curves east, northeast, and finally northwest over the spur south of the ranch house at the mouth of Trout Creek Canyon, in a pattern that indicates a dip of about 25° NW. The wedge of Valmy sandstone forms a horse between the low-angle thrust and the high-angle Basin Range fault.

Perhaps the most interesting segment of the Roberts fault in the Goat window extends east from a point near the mouth of Trout Creek. At the canyon mouth the fault dips gently northward, but as it is followed eastward its outcrop rises in the north canyon wall and its dip gradually steepens. In a quarter of a mile the dip becomes 60° N., and 500 feet farther east it is nearly vertical (fig. 30). Here the fault has Slaven Chert on the hanging (north) wall and Shwin on the footwall. A branch fault beneath it has comparable steepening and the two join at a point about a mile above the canyon mouth.

Beginning about half a mile west of the fault junction, a wedge of Valmy Formation lies between the Slaven Chert of the western facies and the Shwin Formation of the eastern facies and widens eastward. The Roberts thrust south of this wedge of Valmy continues eastward as a near-vertical fault for about a quarter of a mile from the junction with the footwall fault and

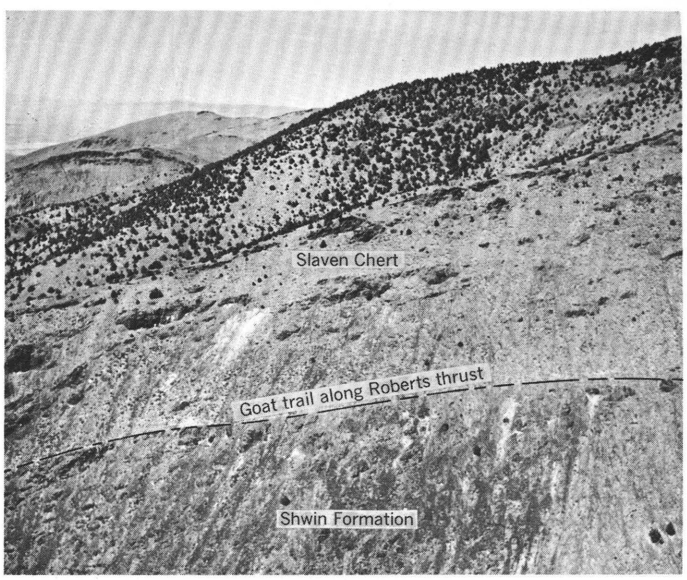


FIGURE 30.—View of north wall of Trout Creek canyon about a quarter of a mile above the mouth. The goat trail, making a light band across the canyon wall about a third of the way up, has been worn on the breccia zone marking the Roberts thrust, which here dips steeply away from the observer. The chert ledges above it are in the Slaven Chert and, as is common with other thrusts in the range, lie roughly parallel to the thrust beneath. Beds in the Shwin Formation below are much contorted.

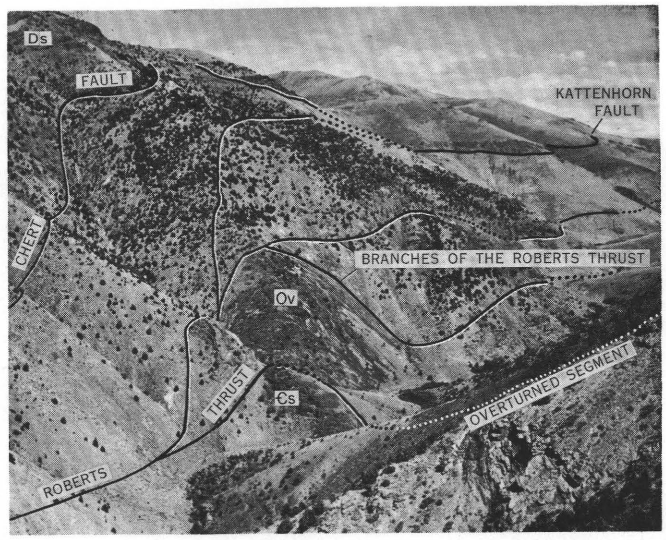


FIGURE 31.—View looking east-northeastward from the south wall of Trout Creek canyon showing hairpin bend (overturn) in the Roberts thrust; also several branch faults and the tectonically much higher Kattenhorn fault. Ds, Slaven Chert; Ov, Valmy Formation; Ꞓs, Shwin Formation.

then turns, in a hairpin curve with a radius of less than 300 feet, through an arc of 180°, still remaining nearly vertical. (See pl. 1 and figs. 29 and 31.) For about 200 yards it maintains a westward course about parallel to its northern segment (distorted by foreshortening in fig. 31.) Then its trace turns abruptly southward, crosses Trout Creek, and goes up the steep cliffs on the south wall of the canyon. Here the fault is well exposed; it dips steeply westward and has the Shwin Formation on the hanging wall and the Valmy below. In other words, the Roberts fault has been overturned so that the rocks of the window have come to overlie those of the "frame."

This relation persists almost to the crest of the ridge but at a point about 2,000 feet north of the crest a thrust plate or "semi-klippe" cuts off and overlies the overturned segment of the Roberts thrust. The Valmy in this block overlies the folded thrust, and both the carbonate facies to the west, and the siliceous facies to the east. This "semi-klippe" of chert of the Valmy forms a U-shaped remnant high on the western face of the ridge above Hancock Canyon. (See fig. 32.)

The Roberts Mountains Limestone and Shwin Formation of the eastern facies emerge from beneath the klippe of Valmy just north of the crest of the ridge and rest on chert and sandstone of the Valmy. The Roberts thrust here dips steeply west. Thus the overturn of the Roberts thrust persists until the structure is cut off at the ridge crest by dikes and a small stock of granodiorite of Tertiary age.

In the area now occupied by this stock, the fault must have again been warped sharply into a vertical position and a nearly east strike, for immediately east

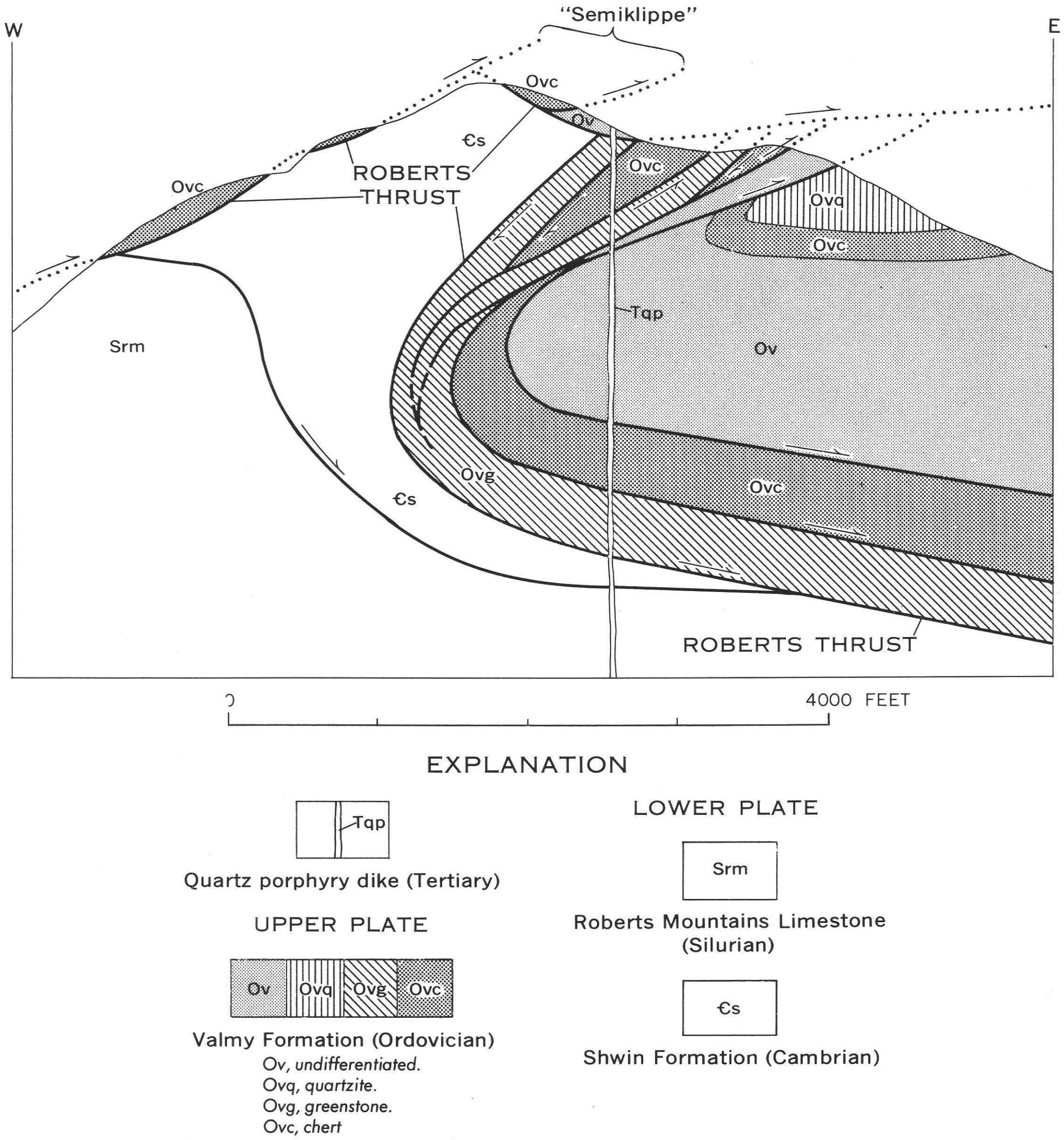


FIGURE 32.—Approximate east-west section through a point about 1,500 feet north of the granodiorite stock on the crest west of Goat Peak, in sec. 20, T. 29 N., R. 45 E., illustrates the warped and overturned Roberts thrust, higher branches broken off as the overturn developed, and an "apple seed" wedge ejected from the tightly squeezed synform in the upper plate of the thrust. Lower part diagrammatic as plunges were ignored; the thrust probably lies much deeper. (See section *G–G′*, pl. 2, which is along a line not far to the north.)

of the granodiorite contact the chert of the Valmy on the north is separated from the Roberts Mountains Limestone to the south by an east-trending vertical fault—the west end of the Trout Creek fault (pl. 6 and fig. 35). This relation persists for about 3,000 feet, with some small slivers of Shwin and Hanson Creek on the "window" side of the fault and with sandstone of the Valmy, instead of chert, on the "frame" side. The Trout Creek fault (and several branches of it) is here occupied by quartz porphyry dikes of Tertiary age, ranging from 50 to 300 feet in thickness. These dikes, with branches, continue eastward with minor interruptions for about 3½ miles and mark the course of the Trout Creek fault toward the east. As just noted, the west end of the Trout Creek fault might be considered as a vertical segment of the Roberts thrust.

The Roberts thrust again appears, south of the dike marking the Trout Creek fault, on the east side of the deep ravine in sec. 28, T. 29 N., R. 45 E. Within a few feet of the dike the fault is exposed as a gently east dipping surface that separates the Roberts Mountains Limestone beneath from the overlying quartzite and chert of the Valmy. The fault trace rises westward up the ridge, with a sliver of Shwin intervening between it and the Roberts Mountains Limestone. At the ridge crest, at the head of the canyon above the Shwin ranchhouse, the fault dips very gently eastward, with a sliver of Hanson Creek dolomite between the Roberts Mountains Limestone and the sandstone of the Valmy.

High on the east slope of this canyon the fault trace drops steeply—perhaps along an obscure normal fault, as there is an outlying klippe along the same line to the south—but then swings eastward and climbs 300 feet higher. Possibly a synformal warp rather than a fault determines this sag in the contact, because a higher branch of the Roberts fault cuts down eastward in a smooth plane to transect the silver of sandstone of the Valmy and bring a straight ledge of chert into contact with the underlying Shwin Formation. This chert ledge, the quartzite ledge above it, and the Roberts thrust, all maintain nearly rigid parallelism for about a mile along the south side of the Mill Creek-Trout Creek divide, though one small sliver of quartzite of the Valmy on the crest of the spur west of Goat Canyon seems to have been caught as a horse in the fault. The main thrust farther east cuts across the chert, quartzite, and still higher thrust scales and descends to the North Fork of Mill Creek near the west edge of sec. 35, T. 29 N., R. 45 E. At the creek it must dip rather steeply eastward, but the fault rises southward along a trace that accords with a fairly gentle southeasterly dip, almost to the crest of the divide between North and Middle Forks of Mill Creek, whence it trends west to North Fork with a gentle southerly dip.

Klippes within the Goat window.—The large klippe of Valmy on the west face of the ridge west of Goat Peak has already been mentioned in describing the overturned segment of the Roberts thrust. Around the south, west, and north sides of this mass, Valmy chert and sandstone rest in fault contact on the Shwin formation of the window. But the east edge of the mass has overridden the overturned segment of the Roberts thrust, doubtless because of movement on the main fault after the overturning had taken place, as happened with the segment of the Chert fault in Trout Creek. (See fig. 32.) East of this overstep, of course, Valmy of the detached block rests on Valmy of the "frame" of the window so that the detached block, while a klippe, is not properly a klippe of the Roberts thrust as here restricted, for it does not everywhere rest on the rocks of the lower plate of that thrust. Although the shear plane beneath it is continuous along the eastern side, it is a klippe of a subsidiary thrust and elsewhere a klippe of the Roberts thrust proper. The somewhat unusual relation of the block is illustrated in the section of figure 32. It is inferred that movement on the east-dipping segment of the Roberts thrust ceased during the folding of the thrust surface as the sliding was transferred successively to higher and higher fault branches.

As the overturning continued, a wedge of chert and greenstone was squeezed out (like an apple seed between fingers) from the tight synformal bend of the thrust. The whole overturned structure was cut by renewed movement on still higher thrust planes. The relation is significant in proving that the overturning of the Roberts thrust took place during the thrusting.

The klippe shown in section (pl. 2) and on plate 1 is succeeded, farther west down the steep face of the range, by three other klippes; these serve to control the configuration of the thrust shown in figures 29 and 32.

A klippe of Valmy Formation occurs on both walls of the canyon about a mile above the fork at the Shwin Ranch, and a whole series of very small ones extends southeastward from here to the North Fork of Mill Creek. It is thus apparent that we have fairly good control of the fault attitude on the south and west sides of the ridge.

Minor windows near the Goat window.—There are three very small windows near the major Goat window. (See fig. 33.) These are: (1) in the valley of Goat Canyon, (2) in the small tributary of Mill Creek between North Fork and Middle Fork, and (3) in Middle Fork, about half a mile above its mouth. Two of these expose the Roberts Mountains Limestone, all expose the Shwin Formation and are separated only by thin residues of the upper plate from the main Goat window.

Assignment of the Shwin Formation to the lower plate.—Because the structural relations, rather than merely its lithologic composition, should be determinative of the assignment of the Shwin Formation to either the upper or lower plate of the Roberts thrust, the question has been deferred to this point, rather than being discussed in the section on stratigraphy.

The stratigraphic arguments for considering the Shwin part of the eastern facies are several: (1) much of it is very similar lithologically and paleontologically to parts of the Secret Canyon Shale and Geddes Limestone at Eureka, and the weathered surfaces closely resemble those of the Secret Canyon and Geddes; (2) the Shwin here is closely associated with dolomite resembling the Eldorado Dolomite of the eastern facies, just as are the Geddes and Secret Canyon to the southeast; (3) the northwestward onlap of the Eureka Quartzite from Eureka to Cortez (see p. 10) makes reasonable the close association of Eureka Quartzite here with rocks resembling Geddes and Secret Canyon.

The close association of volcanic rocks with carbonate rocks of the eastern facies is anomalous, and volcanics

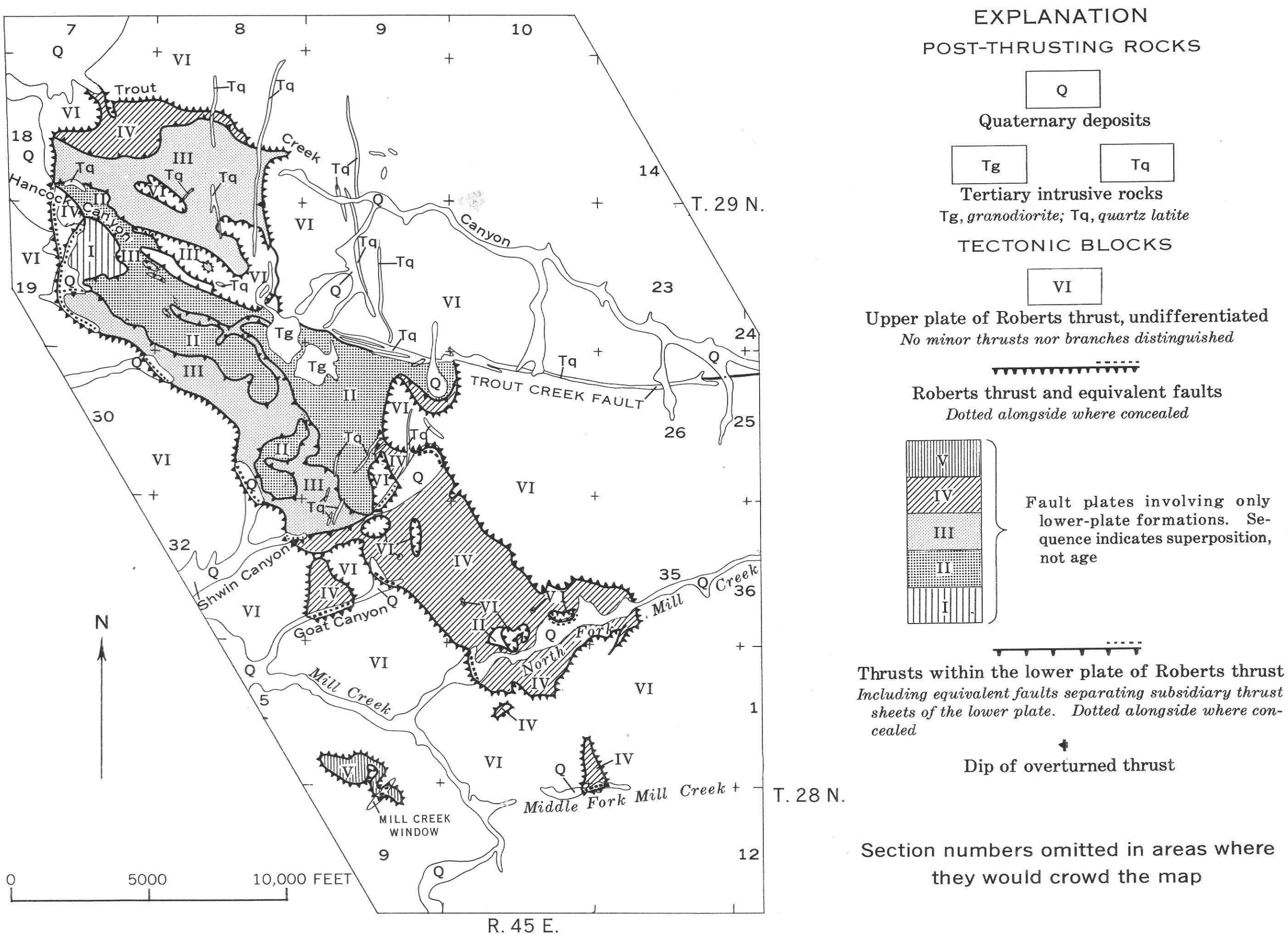


FIGURE 33.—Larger tectonic units of the lower plate of the Roberts thrust in the Goat, Mill Creek, Middle Fork of Mill Creek, and Goat Canyon windows. The fault blocks are numbered from the bottom, in the order of their superposition.

in themselves indeed suggest a western facies. This consideration would carry more weight if we knew of Middle Cambrian volcanic rocks definitely of the western facies anywhere in the general region. If, as we think most likely, the volcanics do in fact interfinger with the carbonates so as properly to be included in it, the Shwin Formation is the only Middle Cambrian formation having a volcanic component thus far known in Nevada. As all the Paleozoic section seems to change from carbonates in the southeast toward siliceous and volcanic facies in a northwest direction, it is not surprising to find volcanics appearing in the Middle Cambrian part of the section. There is no a priori reason though, why the facies boundary should have remained fixed throughout Paleozoic time and thus no reason why the volcanic-bearing Shwin Formation should not be parautochthonous in the Shoshone Range, even though nearby volcanic-bearing rocks of Ordovician and Silurian age are not. Furthermore, the limestone component of the Shwin would be as anomalous in a western facies formation as a volcanic component in the eastern facies. Accordingly we regard the volcanic component of the Shwin as irrelevant to its assignment to one plate or the other of the Roberts thrust. The matter must be decided on the balance of structural as well as stratigraphic evidence.

The structural evidence also is not absolutely unambiguous but it, like the stratigraphic, seems on balance to favor an assignment to the lower plate. The Shwin is everywhere intimately associated with rocks of admitted eastern facies. On the spur north of Hancock Canyon it lies almost parallel to the underlying Eldorado Dolomite (though the contact is a fault) and is overlain by a fault plate of Roberts Mountains Limestone. Small klippes of Roberts Mountains Limestone and Hanson Creek dolomite overlie the Shwin just east and west of the North Fork of Mill Creek and intervene between the Shwin and the smooth basal surface of the overriding plate of Valmy Formation. We consider this smooth surface the Roberts thrust.

In the small window on the Middle Fork of Mill Creek the Shwin is again associated with Roberts Mountains Limestone. Nowhere is the Shwin separated by fault slices of Valmy, Slaven, or Elder—definite western facies formations—from other formations here considered parts of the lower plate.

On balance, we feel that the structural associations of Shwin and its lithologic and paleontologic affinities all favor—even though they do not prove—the assignment of the formation to the lower plate (carbonate or eastern facies) of the Roberts thrust. If it were included with the Valmy in the "upper" plate of the thrust, anomalous slivers of it would be found far down in the "lower" plate at Hancock Canyon and equally anomalous slivers of the "lower" plate would be found well up in the "upper" plate in the North Fork of Mill Creek. Assignment to the lower plate seems to us to conform to the weight of admittedly inconclusive evidence and to lead to the simplest possible pattern for the Roberts thrust.

MILL CREEK WINDOW

The window here referred to as the Mill Creek window is only a little farther from the Goat window than the three minor ones previously mentioned.

This window is small—probably less than 60 acres—on the south side of Mill Creek about half a mile southwest of the junction of North Fork with the main stream. It is noteworthy for two features: (1) the bounding fault crosscuts almost at right angles at least three fault scales of the upper plate of the Roberts thrust within about a thousand feet (not shown in fig. 33 but clearly seen on pl. 1), in such a way as to suggest an injection "diapir" structure, and (2) the striking lithologic contrast of the Roberts Mountains Limestone in this window with that exposed only a little over a mile to the northeast in the Goat window.

The fault which bounds the eastern facies in this window and is therefore called the Roberts fault, is irregular but stands at a high angle along its entire exposed course, except perhaps along the west and northwest sides where exposures are poor. It is especially clear, in the walls of the small canyon that crosses the eastern part of the window, that the Roberts Mountains Limestone forms practically an injected body that transgresses a thrust sliver of chert, one of quartzite, one of greenstone (all of the Valmy Formation) and partly across a higher one of Slaven Chert, all in the upper plate of the Roberts thrust. The Slaven scale forms nearly three-fourths of a ring around the mass of Roberts Mountains Limestone. The structure strongly resembles a diapir intrusion of the salt-dome type. It obviously cannot be due to doming of the fault slices, for the fault scales that the Roberts Mountains body crosscuts are not themselves domed as is the bounding fault surface. The Roberts Mountains Limestone acts here as an injected body.

That the salt-dome analogy is not unreasonable may be inferred from the lithology and structure of the Roberts Mountains Limestone in the window. This is a body of very thinly laminated, fissile, silty, and shaly limestone. It is much more silty and less carbonaceous than is the Roberts Mountains Limestone of either the Goat or Gold Acres window and weathers rusty brown rather than dark gray. The fine lamination, that permits splitting slabs 8 inches across and one-fourth inch or less thick, is, however, present in both localities. Here it prevails throughout the area and is uninterrupted by massive ledges such as crop out to the north in the Goat window. A rock of this kind is highly plastic, as is clear from the contortion and shearing it reveals along the bounding faults. It seems probable that this window really represents injection by flowage of these plastic rocks into a pressure "low" that was perhaps conditioned by folding or by bridging effects of more competent higher thrust scales. Such an interpretation is of course not demonstrable, but if the structure were formed wholly from folding of the Roberts thrust, such folding must have been on an extremely tight pattern, with plunges of more than 60° in opposite directions within less than half a mile. A subjacent laccolith is ruled out by the failure of other thrust slivers to show comparable doming; a diapir intrusion seems more likely.

HORSE MOUNTAIN WINDOW

The Horse Mountain window is low on the western slopes of Horse Mountain, near the southwest corner of the Mount Lewis quadrangle.

The Roberts thrust bounding this window is warped into a low arch in east-west section (see sections *L–L′* and *M–M′*, pl. 2) and a relatively steeper one in north-south section (sections *1–1′* and *2–2′*, pl. 2). Along the western border of the window for the southern half mile a series of thin fault slivers of Hanson Creek dolomite alternate with chert of the Valmy Formation. Farther north the fault is somewhat more clean cut, with chert of the Valmy overlying Hanson Creek dolomite and Roberts Mountains Limestone. At the north end a warp in the fault permits a long tongue of chert of the upper plate to be preserved on the Roberts Mountains Limestone. The northern, eastern, and southern borders of the window are clean cut and almost free from breccia or horses in the fault.

THE FORM OF THE ROBERTS THRUST SURFACE

The description of the several windows has indicated many of the exposed features of the Roberts thrust, from which its local geometrical from can be deduced. From projections of the structures of the upper plate, illustrated on plate 2 and described on pages 119 to 122, it has been possible to outline with some consistency and probability what must be the approximate shape of the thrust in the southern part of the map area; the shape to the north of the Trout Creek fault is far less certain. (See fig. 29.)

The structure contours are, of course, only approximate and cannot be considered accurate—perhaps by a wide margin—except at points controlled by outcrops. A glance at the sections of plate 2 shows that many thrust sheets within the upper plate must cut off against the Roberts thrust. The fault therefore probably lies generally at lower elevations than would be inferred from a mechanical projection of the higher thrust plates. In this sense, the contour map of figure 29 is probably a "maximum altitude" map of the thrust; in reality the fault may lie at considerably lower levels over much of the area.

With these limitations in mind, it is nevertheless probable that the general form of the fault surface is much as it is shown in figure 29. To the south of the Trout Creek fault, the Roberts thrust lies generally above sea level, except in the synform that plunges southward in the south-central part of the area. To the north of the Trout Creek fault, the Roberts must plunge quickly to depths at least as great as in the southern synform; it remains below sea level over most if not all of this area. Although the Trout Creek fault is wholly in the upper plate of the Roberts thrust for the eastern 4 to 5 miles of its length (see sections *8–8′*, *9–9′*, and *10–10′*, pl. 2), it offsets the thrust itself toward its western extremity (sections *4–4′*, *5–5′*, and *6–6′*, pl. 2). The contours of figure 29 reflect this influence: almost certainly the location of the Trout Creek fault has been determined by the concurrent flexure in the Roberts thrust.

A point of uncertain significance is brought out by the admittedly crude diagram of figure 29: It is apparent that the two western folds in the thrust surface—those revealed in the Horse Mountain and Goat windows—tend to be alined along axes trending about N. 35°–40° W. Possibly the Gold Acres window trends more nearly parallel to these than is suggested on the map. A similar trend is suggested in many of the more easterly windows shown in figure 28, although, in absence of altitude control, the patterns—elongate in this direction—shown by many of these windows might be due to topographic accidents rather than to structural trends. The fact that most of the range-bounding faults trend east of north makes probable the interpretation that the north-northwest trends are real and significant in the regional tectonic pattern.

This point has been emphasized by Roberts (1960), who suggested that the upwarping that exposes the windows from Lone Mountain to Goat window has recurred at several times. The argument for recurrence seems not compelling, however. Evidence has been cited by Roberts (1960) for exposure of the windows of the Monitor Range and Lone Mountain soon after the thrusting and we consider the evidence very strong that both the Goat and Cortez-Gold Acres windows were upfolded concurrently with the thrusting. (See p. 106-108.) There is no particular association of intrusives with upwarps on the Roberts thrust and therefore no reason to consider the upwarps as rejuvenated in Mesozoic or Tertiary time. The association of economic mineralization with the windows, referred to by Roberts (1960) may, if real, merely be another example of the widely recognized ready susceptibility of carbonate rocks to ore mineralization.

Another feature illustrated by figure 29 is the highly contorted form of much of the thrust surface as revealed by the elevations seen in the window borders and klippe outliers. Some of the klippes may of course represent subordinate branch faults from the main surface of movement, such as are seen along the west side of the Horse Mountain window, but others without doubt are due to small-scale folds in the main thrust surface itself. The possibility that the form of the facies contact in the Mill Creek window is due to diapir injection has already been discussed; the shape of the contours of figure 29 is at least consistent with this possibility.

The most striking of the folds is conspicuous in figure 29 in the area north of the abrupt western termination of the Trout Creek transcurrent fault. Here the structure contours, except for the 8,000-foot contour, are all dashed, indicating that the higher overlies the lower and the thrust is overturned.

This overturned segment of the Roberts fault seems to offer important clues, both as to the time of the folding and as to the mechanism of fault movement; it is accordingly discussed here in some detail.

OVERTURN IN THE ROBERTS THRUST

As described and shown on plate 2 (section *G–G′*), and figures 29, 31, 32, and 34, the Roberts thrust is overturned and dips westward throughout its north-south course on the steep cliffs that form the south wall of Trout Creek. Here, through a horizontal distance of nearly a mile and a vertical distance of more than 3,000 feet, wherever the Roberts thrust is not concealed by a sliver from its upper plate, rocks of the eastern facies are revealed in its hanging wall and those of the western facies in its footwall—the rocks of the window rest on those of the frame.

The significant features are shown on figure 34*F* and our interpretation of their development is indicated in the sequence figures 34*A* to *F*.

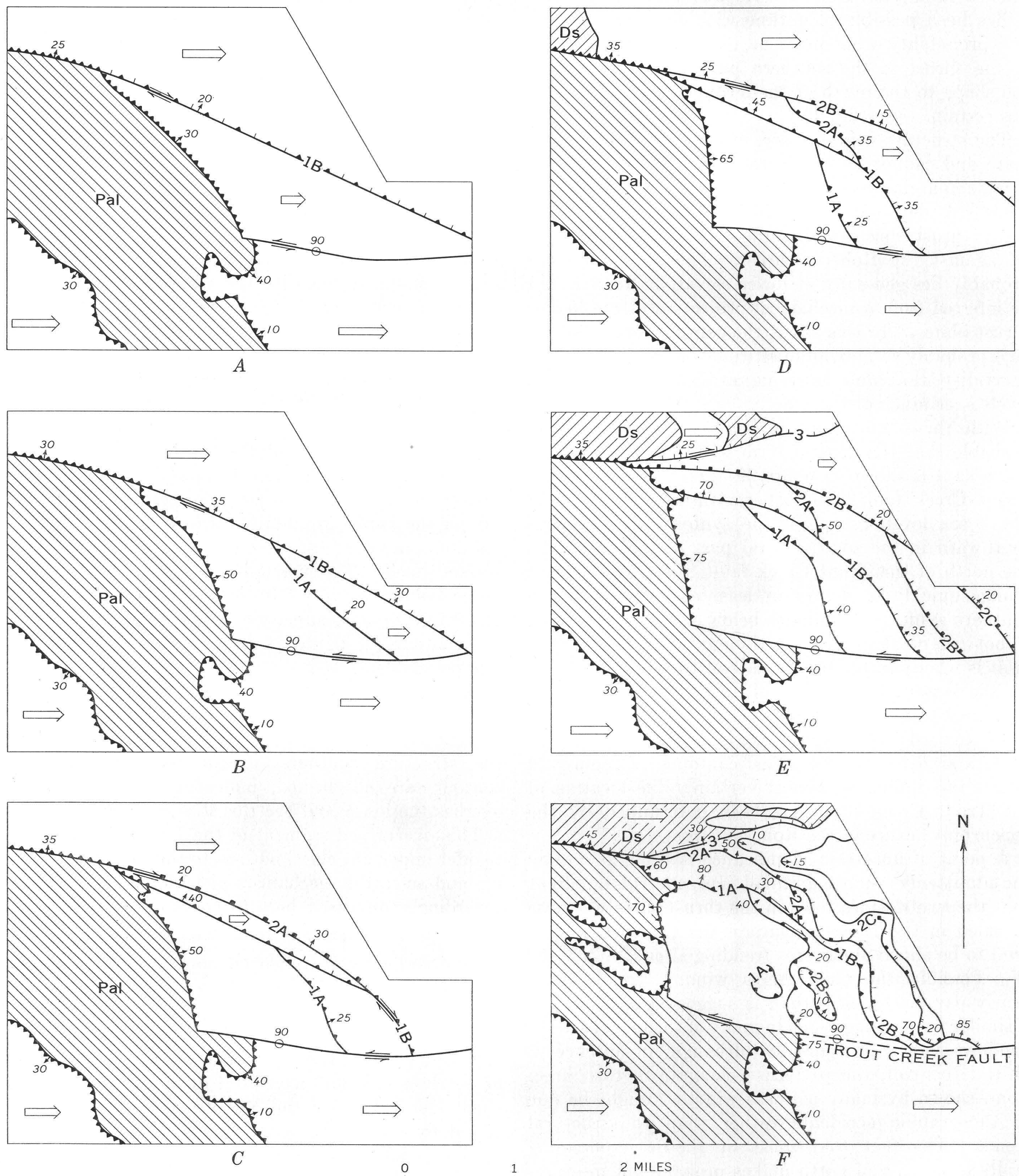
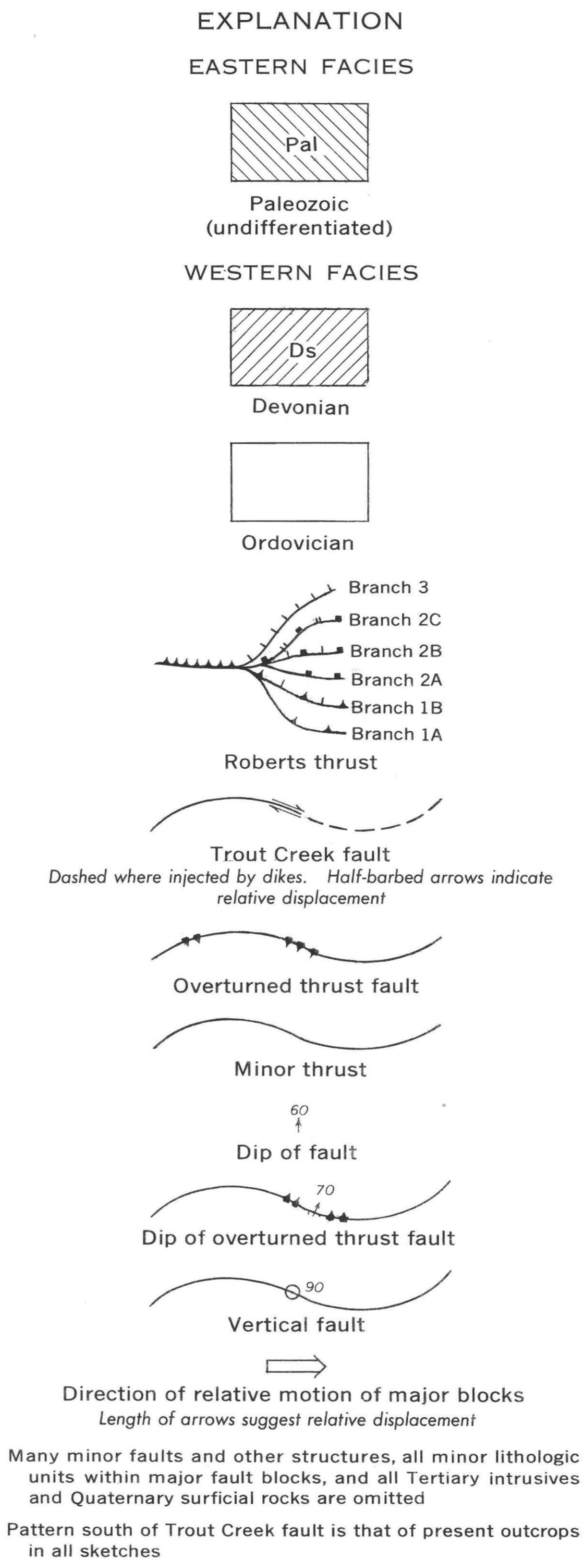


FIGURE 34.—Structural diagram of the Trout Creek-Goat Peak area, showing overturn in the Roberts thrust, branch faults nearby, and the Trout Creek transcurrent



fault. *A–F* represent hypothetical stages in development of the structure.

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The vertical segment of the Roberts fault and its eastward projection to the sharp hairpin turn are, we believe, significant. So, too, are the branch faults, branches 1 and 2 in figure 34. Branch 1 leaves the Roberts thrust at the north side of the hairpin turn and extends east and southeast in the upper plate of the fault, branching about 4,000 feet from the main Roberts fault into branches 1A and 1B. Branch 2 leaves the Roberts fault nearly a mile west of the hairpin turn and itself shows a sharp drag fold (or perhaps only a stubby horse within it) in the wall of Trout Creek canyon almost north of the hairpin turn in the Roberts fault. Thence it, too, curves southeastward into the creek bottom and branches, with sharp curves in each branch.

A still higher branch fault, No. 3 of figure 34, which is called the Chert fault on plate 6, coincides with branch 2 for nearly 3,000 feet from the point where it leaves the Roberts thrust but continues straight on where branch 2 curves southward. (For clarity the relation of pl. 2 is somewhat exaggerated on figure 34*E* and *F*.) Undoubtedly the latest movement of the Roberts thrust in the segment north of the Goat window took place along this Chert fault surface.

All these branch faults, when considered together, suggest strongly that the hairpin fold in the Roberts thrust was actually formed during the main movement of the fault and concurrently with motion on the Trout Creek fault. The fold does not seem explicable by subsequent folding of an originally planar structure. This conclusion seems important because we have strong evidence only a few miles to the north of a second orogeny—the Lewis orogeny—also marked by thrusting. (See p. 123.) Because this thrusting might be thought to account for the overturn in the older and underlying Roberts thrust, we believe it worthwhile to develop our point more fully.

Figure 34*F* has been generalized from the geologic map, plate 1, by eliminating all the intrusive and Quaternary rocks, many minor faults, and the distinctions between the lithologic variants of the Valmy Formation (above the Roberts thrust) and between the formations of the eastern facies below the thrust. As a generalization from the map (whose area has a relief of about 4,000 ft) the diagram is a projection onto a horizontal surface of the traces of the faults on the present very irregular topography. In developing the geometry of this structure in the other diagrams of figure 34, it has seemed to us desirable to maintain the projection from this irregular surface (except for klippes and such minor fault blocks) even though, as is obvious, no such reference surface can be defined accurately except for the present stage of erosion and the surface itself has no significance to the analysis. It merely permits transfer of the geometry implied by the diagrams to the present map, so that the reader may place the growth of the structure in relation to its status as we see it now.

Our ideas of the sequence in development of the complex fault relations of figure 34*F* are shown in figures 34*A* thru *E*. These are, of course, purely diagrammatic, though they do possess enough control reasonably to produce the pattern of 34*F* by moderate deformation beyond that of stage 34*E*.

The heavy arrows show our conception of the relative movements of the major fault blocks: the overriding plate relatively eastward, the lower plate relatively westward, each with respect to the other. The faults most active at any particular stage are indicated with "half-barbed" arrows, though it is our conception that fault motion continued in a minor way on all faults almost throughout the deformation. So, too, did plastic deformation within the fault blocks, with growing drag and bending of faults under the moving weight of higher fault blocks, after the main period of shearing along any particular fault segment had ended. According to our reconstruction, after the Trout Creek transcurrent fault began to form, it continued active throughout the structural evolution. Because of the constant shortening of the structures north of the line of this fault, successively higher splits off the Roberts thrust became activated while successively lower faults became more and more locked and then bent.

Figure 34*A* illustrates the possible relations at the stage when the Trout Creek fault was just beginning to form—along the line between areas of relatively free differential motion of the thrust plates to the south and of relatively impeded motion to the north. It is doubtless significant that the northward plunge of the Roberts fault, which begins just north of the Trout Creek fault, appears to continue for several miles toward the north end of the map area. Relatively impeded motion north of the line of the Trout Creek fault as compared with the freer motion to the south is a rational inference from the existence of the fault itself, with left-lateral movement along it revealed by the relations at its western end. Figure 34*A* is drawn to suggest that the relative compression of the upper plate (east of the incipient fold in the Roberts fault), brought about by the left-hand torque expressed by the Trout Creek fault, had already led to a flexure in the Roberts thrust. The overriding block was unable to follow this flexure laterally. Accordingly a new shear, branch 1, has sprung from the main fault at the inflection point and extends 3 to 4 miles to the southeast.

It is assumed that between the stages represented by figures 34*A* and *B*, this differential compression and drag to the north of the Trout Creek fault had bent branch 1 into the trend of 1A enough to induce a higher split, 1B. Increasing drag at the fold inflection to the northwest had led to right-hand drag of the blocks on both sides of the Roberts thrust, bending the fault into an incipient hairpin fault. The stagnation of the movement along the Roberts fault to the south of this bend then diverted the main fault motion north of the Trout Creek fault to the path along the smooth curve marked by Roberts fault (closely spaced triangles), branch 1A (widely spaced triangles), and branch 1B (widely spaced triangles and short dashes). This is the stage illustrated in figure 34*B*.

Motion shown at stage 34*B* implies a right-hand drag couple at the bend of the Roberts fault. Continued compression of the material in the angle between the originally active (but now dormant) segment of the Roberts fault and the projection of the Trout Creek fault is assumed to increase the curvature of branch fault B so much that a new split, branch 2A, formed as shown in figure 34*C*. At the same time, the compression in the angle to the southwest of this branch steepened the dip of both the Roberts fault and the older, now passive, splits from it.

Figure 34*D* illustrates continued growth of the Trout Creek fault, continued compression of the material in the angle northwest of it, continued drag and sharpening curvature of the Roberts fault at its bend, and of branch 2A, so that a split, 2B, breaks off from it. At the same time the dip of the Roberts fault just north of the Trout Creek fault continues to steepen.

Figure 34*E* shows all these processes continued, with the drag kink in the Roberts thrust having grown to the point where a new split, branch 3 (the Chert fault of pl. 6), has emerged at a point still farther west, taking up the general eastward motion of the overriding plate to the north of the "pressure shadow" formed along the steep north-trending segment of the Roberts thrust. In this development, the Roberts thrust segment, just north of the Trout Creek fault, has become nearly vertical and the Devonian mass in the upper plate of the Roberts thrust has moved along the Chert fault for several miles past the lower plates of Ordovician rocks. The transition to figure 34*F*—essentially the present relations—involves a little further motion of the same type and the further steepening and overturning of the Roberts fault.

The above discussion and the sequences shown in figures 34*A* thru *F* are of course nothing more than schematic. They do, however, in our opinion, plausibly account for the geometric pattern now seen, and accord with the differential motion implied by the Trout Creek fault and by the regional plunges of the

Roberts thrust. No one could of course maintain that such conformity is evidence in support of the genetic pattern outlined. It is permissive only, and doubtless several other patterns and sequences could be devised to produce comparable conformity. It is submitted, however, that such a genesis of the overturn in the Roberts fault, of the hairpin bend in its near-vertical segment, and of the increasing simplicity in curvature of the successively higher off-splits, more satisfactorily accounts for their mutual relations than would the folding of a preexisting planar fault during a later independent orogeny.

The pattern expected on the assumption of later folding of a pile of thrust sheets is illustrated in figure 35*A*; that to be expected under the assumption of concurrent thrusting and folding is illustrated in figure 35*B*. It seems to us that this pattern is closely analogous with that of figures 34*A* thru *F* and plate 2. Only by such concurrent folding and thrusting can the convergence of the fault splits in the neighborhood of the overturn be accounted for as other than fortuitous.

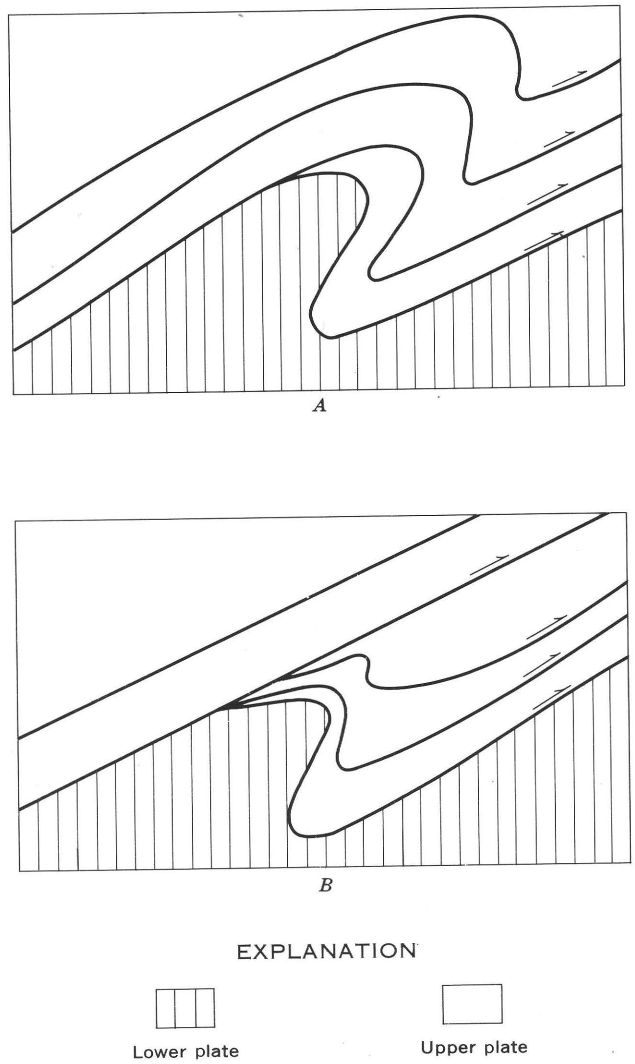

FIGURE 35.—Diagram to show the geometry to be expected of: *A*, a pile of thrust sheets later folded together; and *B*, a pile of thrust sheets formed and folded concurrently. Both diagrams may be thought of either as cross sections or as maps of plunging structures. All the curved lines are thrust faults.

STRUCTURE OF THE LOWER PLATE OF THE ROBERTS THRUST

The lower plate of the Roberts fault is either in place authochthonous, or at least parautochthonous—very much less transported than the upper plate. It consists wholly of rocks of the carbonate (eastern) facies: Prospect Mountain Quartzite, Eldorado Dolomite, Shwin Formation, Eureka Quartzite, Hanson Creek Formation, Roberts Mountains Limestone, limestone of Devonian age, and Pilot Shale. So little of these formations is exposed, except in parts of the Goat window, that it has not seemed worthwhile to try to extrapolate their structure to depth.

Accordingly, though they show the exposed formations by letter symbols, the sections of plate 2 continue all eastern facies rocks in depth under the single rubric "Rocks of the carbonate (eastern) facies." Here we merely discuss the several tectonic units recognized in the windows through the Roberts thrust.

It is impossible accurately to determine the relations of the rocks of the lower plate in one window with those in any other, except perhaps of the small ones immediately alongside the Goat window with those in it. Of necessity, each window must be considered of itself. The structure of the lower plate seems, however, generally to be less variable and the rocks less jumbled than in the upper plate of the Roberts thrust, as is perhaps to be expected because it is surely less transported than is the upper plate.

GOLD ACRES WINDOW

In the Gold Acres window the lower plate seems little sheared parallel to the thrust, as the upper plate so commonly is. Most of the faults in the carbonate facies, of which many have been recognized, are high angle. None has been found to cut the Roberts fault.

Possibly the fault that separates the Roberts Mountains Limestone from the limestone of Devonian age is a major one, inasmuch as no Lone Mountain Dolomite intervenes between them nor, indeed, crops out anywhere in the area. In Lone Mountain and in the Roberts Creek Mountains, the Roberts Mountains Formation is overlain by about 2,000 feet of Lone Mountain Dolomite, beneath the Devonian.

The local absence of beds referable to the Lone Mountain Dolomite may, however, be explained, in part or in whole, by a facies change between the Silurian carbonate formations rather than by faults. On the northwest side of the Roberts Creek Mountains the dolomite of the Lone Mountain is reported to change along strike to dark limestone indistinguishable from the Roberts Mountains Limestone (Winterer and Murphy, 1960).

At the north end of the window the Pilot Shale is in contact with the limestone of Devonian age, along a fault that seems to curve roughly parallel with, but to dip notably steeper than the northward-facing arc of the Roberts fault above it.

Doubtless there are many other faults in the Devonian limestone but no attempt was made to map them on the scale of plate 1.

GOAT WINDOW

The structure of the lower plate in the Goat window is highly complex; its salient features are shown in figure 33. Close to the Roberts thrust there is a very definite tendency not only for the internal shearing of the plate, but also for the stratification of the beds themselves to lie roughly parallel to the thrust surface. This is undoubtedly the result of localized drag, confined to a zone within about 300 feet of the fault; on a larger scale there seems to be no very close parallelism of the formations and fault.

In figure 33, the areas marked by roman numerals I to VI, inclusive, are tectonic blocks, each composed in general of more than one formation. Their numbering corresponds to our impression of their tectonic sequence or topology—not necessarily their age sequence; that is, block I is structurally beneath block II, and so on. Many other faults are mapped on plate 1 but they are believed to be minor, at least as compared to those that bound the numbered blocks.

Block I.—What seems to be the lowest exposed plate is at the west end of the window. Prospect Mountain Quartzite, at the southwest, stands vertically and faces north. It is separated from Roberts Mountains Limestone (in three fault blocks) by a high-angle fault. A similar fault separates the Roberts Mountains block from the Eldorado Dolomite. All these units, together with a small mass of Shwin, which is apparently conformable on the Eldorado Dolomite, are considered to constitute a tectonic unit, block I. This block is bounded above by a fault, strongly curved from a dip of about 40° SW. where it bounds the Prospect Mountain, through the horizontal to a dip of about 25° NE., on the spur north of Hancock Canyon. The fault may be curved on an axis plunging gently westward roughly parallel to higher faults and to the Roberts fault itself. It is marked by a conspicuous ledge of quartz that seems chiefly to represent limestone replaced by quartz, but may include some ground-up quartzite fragments derived from either the Prospect Mountain or the Eureka Quartzite.

Block II.—To the north, east, and southeast, block I is overlain by a large but much contorted mass of Roberts Mountains Limestone which seems everywhere highly discordant with rocks above and below. This mass, which extends as far southeast as the Shwin ranchhouse, is designated block II.

Block III.—Block III, which consists dominantly of Hanson Creek dolomite but also includes a little Eureka Quartzite and considerable Roberts Mountains Limestone, rests directly on the Prospect Mountain Quartzite of block I on the southwest spur of Goat Ridge. The bounding fault here dips steeply southwest. A mile or so to the southeast it even steepens to 70°, where block III is resting on block II. The fault separating these units is very irregular and curves roughly parallel to the ground surface so that small outliers of block III rest on block II on both sides of Hancock Canyon as well as far to the southeast, west of the canyon in which the Shwin ranchhouse lies.

West of the Shwin ranchhouse, the principal formation of the lower plate on the north side of Mill Canyon is the Roberts Mountains Limestone, with many infaulted blocks of Hanson Creek dolomite. Though exposures are inadequate to demonstrate this, these rocks seem to make up a thrust sheet that passes beneath block IV to the southeast. In the drainage of the west fork of the canyon of the Shwin Ranch block III forms a simple homocline, but farther west the plate has been considerably folded, as well as the fault on which it lies. The fault between blocks II and III, on which the Hanson Creek overrides the Roberts Mountains, is, in a very rough way, parallel to the Roberts thrust and, like it, is folded into a large antiform with minor smaller folds.

Block IV.—The largest tectonic unit of the lower plate is here referred to as block IV. At various places it rests on each of the two next lower blocks but comes into contact with block I only north of Hancock Canyon, at the west edge of the window, where it is composed of Shwin Formation.

On the northwest slope of Goat Ridge the dominant formation of block IV is the Shwin, resting in apparent fault contact on the Roberts Mountains Limestone of block II. The formation is much crumpled and folded and is notably more phyllitic than elsewhere. (See fig. 4 and p. 11.) The thrust sliver lying above the main body of the Shwin, and also considered a part of block IV (fig. 33) is made up chiefly of metadiabase and other igneous components of the formation.

This thrust sliver is considered part of the same thrust scale as that southeast of the Shwin Ranch which is also composed chiefly of rocks of the Shwin Formation. The formation appears generally to strike northwesterly except near North Fork, where the strike roughly parallels the fault trend. The dips range from rather gentle to essentially vertical, suggesting compression about northwest axes, parallel to the general trend of the window itself and at a very definite angle, not only with the Trout Creek fault but also with the structures in the upper plate to the north of that fault. This plate is interrupted just east of the canyon above the Shwin Ranch where it is overridden by the Roberts fault which there rests directly on block III. The small slivers of Hanson Creek dolomite and Roberts Mountains Limestone just west of the North Fork of Mill Creek appear to be only minor horses in the Roberts thrust.

Block VI.—In figure 33 the designation of block VI is a general one, inclusive of all rocks of western facies and hence of the upper plate of the Roberts thrust. It forms many klippes in the window as well as surrounding it.

SIGNIFICANCE OF THE STRUCTURE OF THE LOWER PLATE

Despite the uncertainties as to details, sufficiently emphasized above, it appears that the major structural elements exposed in the window are four. In geometric sequence they are: block I, a highly jumbled mass at the southwestern and western side of the Goat window, including bodies of the Prospect Mountain Quartzite, Roberts Mountains Limestone, Eldorado Dolomite, and a little Shwin Formation. This mass is overlain by block II, a composite, internally disturbed and much faulted plate of some Hanson Creek dolomite and dominant Roberts Mountains Limestone. This in turn is overlain by block III, a sheet composed chiefly of Hanson Creek dolomite but with subordinate Eureka Quartzite (west slope) and Roberts Mountains Limestone (south slope) of the ridge west of Goat Peak. Highest in the series is block IV, an interrupted sheet composed almost wholly of Shwin Formation, just beneath the Roberts thrust.

Compared to the Roberts thrust, the thrusts that separate these several tectonic blocks or scales are minor. Nevertheless, if the geometric arrangement suggested previously and illustrated in figure 33 is indeed the correct one, these minor thrusts demand movement of at least 5 miles and more likely twice this figure. It seems evident that the only block beneath the thrust that can possibly be rooted in place is block I, figure 33; all the rest are traveled blocks, of which block IV must have moved at least the full width of the window over block III, block III nearly as far over II, and II a minimum of 2,000 feet—probably far more—over block I. If, as seems probable from the fold in the Roberts thrust mentioned earlier, the fault movement has been more directly eastward, rather than northeastward, normal to the present trend of the window, these distances must be increased.

It seems necessary to assume an original extension of the carbonate facies beneath the Roberts thrust for at least half a dozen miles farther west than their present exposures. As this window is the most westerly known to expose this sequence, and as the most traveled formation within it, the Shwin, of block IV, is only in part transitional to a clastic facies, these facts are significant in appraising the original depositional relationships of the two facies brought together by the Roberts thrust.

MINOR WINDOWS NEAR THE GOAT WINDOW

Figure 33 shows three minor windows through the Roberts thrust. All are close to the Goat window and all expose rocks of the same formation, the Shwin, as those in that window, consequently all can readily be assigned to block IV of the tectonic units shown. The small mass of Roberts Mountains Limestone in Middle Fork is lithologically like that formation near the Shwin Ranch and distinctly different from that in the Mill Creek window not far to the west.

MILL CREEK WINDOW (BLOCK V)

The rocks of the lower plate exposed in the Mill Creek window all belong to the Roberts Mountains Limestone. They are nevertheless referred to a block V on figure 33, rather than to block III of that figure, which is largely made up of Roberts Mountains Limestone, because of the notable lithologic contrast between the rocks of the two units (p.18). This point is also discussed along with the Horse Mountain window, next described.

HORSE MOUNTAIN WINDOW

It has been pointed out in describing the stratigraphy that the Horse Mountain window furnishes the only large exposures of limestone in the Hanson Creek dolomite of the area. Furthermore all the contacts with the overlying dolomite blocks of the Hanson Creek or with the Roberts Mountains Limestone are obvious faults. It is likely, however, that these faults follow not far from the bedding, for slivers of Hanson Creek dolomite are widely present between the limestone beds of the Hanson Creek and the Roberts Mountains Limestone. For the most part, the internal structure of both the Roberts Mountains and Hanson Creek blocks seems relatively simple, with open folding, despite the intervening fault. It is clear that the Roberts Mountains Limestone is here rootless; the limestone of the Hanson Creek may perhaps be in essentially its depositional position, though reasons for suspecting that it, too, is a far-traveled block will be pointed out in the sequel.

The Roberts Mountains Limestone of the Horse Mountain window resembles that of the Mill Creek window. It contrasts with rocks of the same formation in block III to the north and in the Gold Acres window to the east. Their rocks are therefore referred to a higher structural plate, block V of figure 33.

These differences make it seem likely that several thrust masses composed of lower-plate rocks have been telescoped between the Mill Creek and Goat windows. The more clastic rocks presumably came from more westerly areas than the more calcareous, judging from the broad regional stratigraphic relations now known. They are in the southwestern windows and presumably represent a higher fault sliver, No. V, than that represented at the Goat window, No. IV in figure 33. In any event, even though the formation represents the eastern facies, it is almost surely not rooted in place in either window.

STRUCTURE OF THE UPPER PLATE

MAJOR FEATURES

The most striking feature of the upper plate of the Roberts thrust in this area is its internal shearing into discrete fault blocks. Although some of these blocks attain thicknesses as great as 8,000 or 9,000 feet, most are far thinner, a few hundred to a few thousand feet. The lateral dimensions of these blocks with respect to their thickness is so great that the term "fault scale" seems more appropriate than "fault block." Although many individual fault scales have been followed for as much as 6 or 8 miles and a few even farther, most are considerably less persistent. Most fault scales are divided internally by other faults, either at low or high angles with their trend. But this subdivision rarely extends down to the scale of a hand specimen, and then only in negligible volumes directly alongside major faults. Almost no slaty cleavage formed in the entire upper plate.

The names shown on plate 6 and on the sections of plate 2 require some explanation. Clearly, as long as a fault can be traced at the surface no serious question as to its identification arises, even though it joins another subparallel one at the end of one of the fault scales. But of course such junctions may be of several kinds and their significance is not everywhere obvious. For example, one fault may merely join the other to enclose a horse between them. Many such horses are miles long and thousands of feet thick. Also one fault may transect the other; the trace of their junction underground may not reappear for several miles so that the identity of the fault that has been cut off is uncertain where it does reappear. Or a fault may split into several strands, some following subparallel courses while others diverge widely. In such cases there is obviously room for much diversity of interpretation: we have cut the Gordian knot of fault nomenclature in this report by the arbitrary decision to apply a particular name to a fault wherever it is topologically possible for a continuous fault surface to exist between the separated segments and at the same time the several fault slivers can reasonably retain the same order of superposition.

Obviously this assignment need not be correct. But we hope that this usage is both consistent with geometry and conducive to a systematic arrangement of the discussion. It should not be misleading if so understood—we by no means consider it probable that the several fault scales continue as widely as this nomenclature suggests. Indeed this seems distinctly improbable. But the general tendency of the slivers to be subparallel makes this orderly treatment seem to us a useful expedient. Many faults named on the sections of plate 2 and on plate 6 are labeled simply to enable the reader to identify the fault on other sections. The discussion following relates only to some of these structures, selected as either especially significant or as representative of others—we by no means consider each named fault to be worth individual description.

It should be reemphasized that the local uncertainties as to stratigraphy have doubtless contributed to the difficulties of interpretation in two ways: (1) by obscuring faults perhaps more significant than some mapped and, (2) by our attributing more significance to some faults than would be demanded were no stratigraphic misidentifications present, further to confuse an already bewildering complex.

THE TROUT CREEK FAULT

EXTENT

The Trout Creek fault—so named because it crosses the range divide at the head of Trout Creek—has already been mentioned in connection with the overturn in the Roberts thrust, where it was noted that at its western end it cuts into and offsets the Roberts thrust. Its continuation to the east, and especially its termination, next to be described, show, however, that for much of its course it is confined to the upper plate of the Roberts fault and it is therefore described here.

The Trout Creek fault has been mapped from the north edge of the granodiorite stock on the crest of the ridge west of Goat Peak for 9½ miles to the east-northeast, as far as The Park, west of Granite Mountain, where it disappears beneath the alluvium. Throughout this distance, its course is a curve very gently concave to the north. Excellent exposures in and near Hancock Canyon permit the categorical statement that the fault does not extend into the lower plate of the Roberts thrust beyond the point where the thrust is exposed on the north side of the Trout Creek fault. Equally good exposures on the fault trend in the Crescent Valley quadrangle show that it does not exist east of the Granite Mountain stock.

Through nearly half its exposed length, the Trout Creek fault is occupied by dikes of quartz latite and quartz latite porphyry of Tertiary age.

RELATION TO ASSOCIATED FAULTS

Exposures on trend of the Trout Creek fault to the east of The Park are adequate to show that no comparable fault appears east or northeast of the stock of Granite Mountain. This trend is of course interrupted by the Corral Canyon fault which bounds The Park on the east. The displacement on the Corral Canyon fault—a high-angle normal fault—while not accurately determinable, is probably 1,000 to 3,000 feet (pl. 2, section *H–H'*). If the Trout Creek fault extended in depth in the way that might be inferred from the marked structural contrast on its two sides, so slight a fault could not be expected to terminate it.

Its disappearance might be attributed to strike-slip movement on the Corral Canyon fault. This seems unlikely; a fault of such obviously large displacement and independence of transverse structures as the Trout Creek fault to the west must have extended deep into the crust and, if so, many thousand feet of lateral displacement would be needed to obviate its continuation in the map area beyond the Corral Creek fault. Furthermore, the Corral Creek fault, when traced southward, seems to die out with no recognizable offset along its trend. There is accordingly no reason to suspect any considerable lateral slip on this fault.

On the other hand, it seems quite unlikely that the Trout Creek fault simply died out at the west side of The Park. The discordance in exposed structures on either side is as great just west of Granite Mountain as anywhere between that point and the west end of the Trout Creek fault, whereas to the east of Granite Mountain a block of the Slaven Chert extends unbroken right across the projected trend of the Trout Creek fault.

The clue to the difficulty seems to be furnished by the relations of the Crum Canyon, Hilltop, and Bateman faults to the Trout Creek fault. None of these extends south of the Trout Creek fault, though each is a major fault whose simple pattern of curvature implies continuity in both strike and dip. The Trout Creek fault thus terminates on the south all the fault scales associated with these faults.

Just west of the north end of The Park, the Bateman fault, a major structural feature which has been traced from the mouth of Crum Canyon across the divide into and up Bateman Canyon to its head, is cut off by the intrusive mass of Granite Mountain. As just mentioned, this stock also cuts the Trout Creek fault just west of The Park and its projected trend just to the east. The Bateman thrust carries rocks of Middle Ordovician age on the hanging wall over the Slaven Chert of Devonian age. Its minimum displacement thus involves faulting out of all the Upper Ordovician, all the Silurian, and the part of the Slaven Chert exposed in the footwall—a thickness of at least 5,000 feet and perhaps several times that amount. The near parallelism of the thrust to the bedding throughout its exposed length suggests a displacement measured in miles rather than in thousands of feet, as such a stratigraphic gap certainly did not develop by high-angle transection of the beds.

The failure of so large a thrust to emerge to the south of the projected course of the Trout Creek fault suggests that, like the Crum Canyon and Hilltop faults, it ends against the Trout Creek fault. The failure of the Trout Creek fault to continue beyond the projection of the Bateman fault implies that the two faults were formed at the same time; the Trout Creek marks a vertical south boundary for the upper plate of the Bateman fault. We have already seen the strong evidence that the Trout Creek fault was concurrently active with the Roberts thrust; we now see that the Bateman fault was active at the same time.

The diagram of figure 36 illustrates the geometry of the fault surfaces involved. The line of intersection of the Trout Creek and Bateman faults must rise eastward into the air from the place where the original junction was destroyed by the Granite Mountain intrusive. (See pl. 2, sections *10–10'* and *G–G'*.)

The vertically lined surface labeled *TC–TC* in figure 36 is the Trout Creek fault. Its west end is drawn as the junction with an overturned segment of the Roberts thrust just south of the hairpin bend toward the top of the figure. There can be little doubt that the Bateman fault is a branch of the Roberts; their junction at a fold in the surface of the major fault is drawn by analogy with the branch faults seen at the surface in Trout Creek Canyon and discussed in detail on pages 105–109 in connection with figure 34, *A* to *E*.

The Trout Creek fault only offsets the Roberts fault as far east as the line *l–l'* in figure 36. From here east it limits on the south the thrust surfaces of the Hilltop (*H*) and Bateman (*B*) faults. (For simplicity, the Crum Canyon fault is omitted from the diagram.) These faults join along the line *l–l'*. With the outcrop of the Bateman fault the thrust plate (Bateman) of which the Trout Creek fault forms the south limit, passed "into the air," high above the Roberts fault, from which it branched at depth, far to the west. The left side of the block shows the Goat window, bounded on the north by a steep segment of the Roberts fault and on the southwest and southeast by more gently dipping segments. Very similar relations have been inferred by Nolan (1935, p. 56–61, fig. 8) for the Gar-

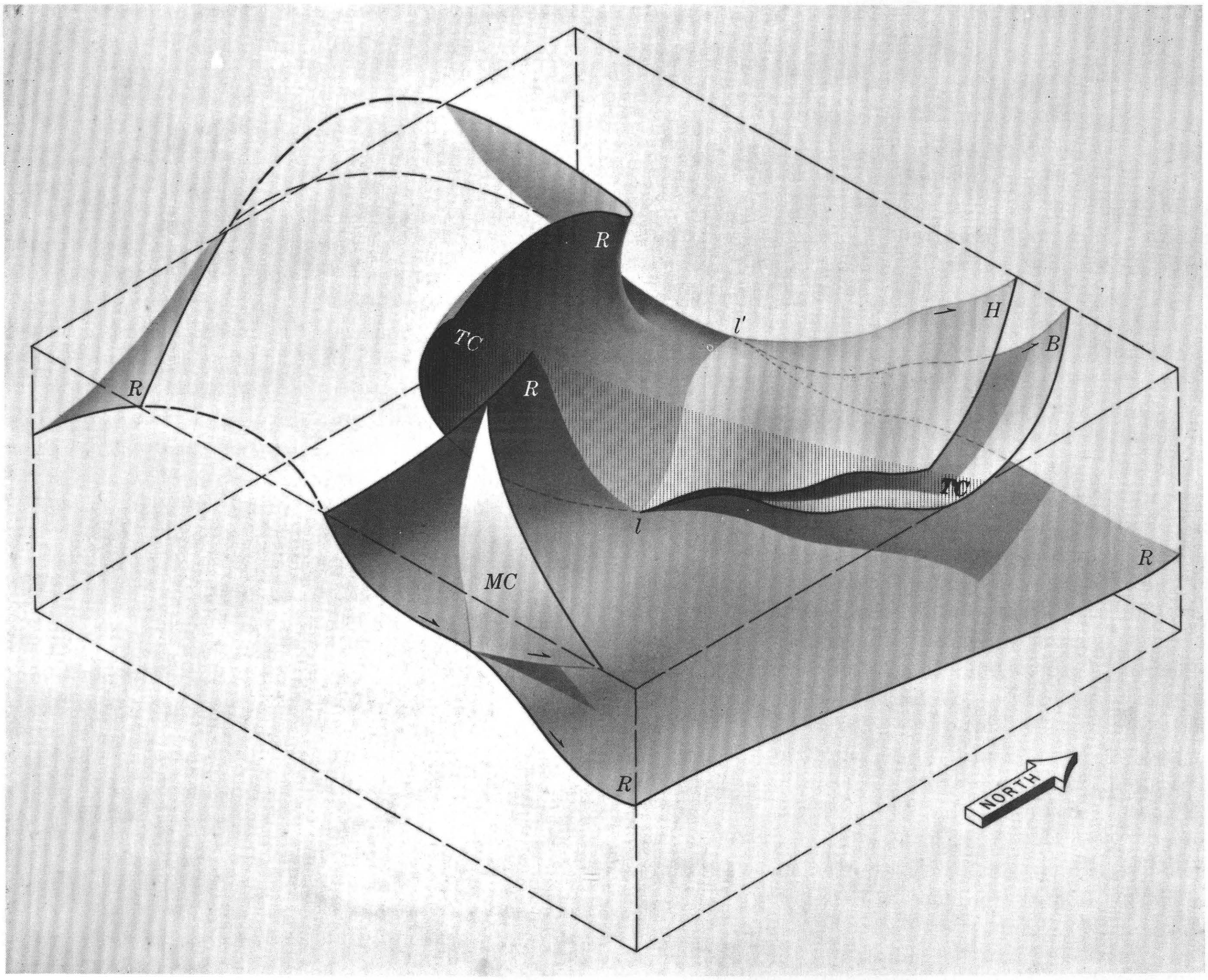

FIGURE 36.—Diagram (not to scale) showing mutual relations of the Trout Creek fault (*TC*, shown by vertical pattern on fault plane) and the Roberts (*R*), Bateman (*B*), Mill Creek (*MC*) and Hilltop (*H*) faults. *l–l′* is the common junction of Hilltop, Bateman, and Roberts faults.

rison Monster transverse fault in the Gold Hill area, Utah.

The relations expressed on plate 2, sections *9–9′* and *G–G′* and in figure 36, show that although the Trout Creek fault served as a channel for many dikes toward its west end (and possibly even for the small granodiorite plugs on Goat Ridge) it could have exerted no control over the emplacement of the Granite Mountain intrusion. The bottom of the Bateman thrust sheet, and hence of the Trout Creek fault, must have been at crustal levels far too shallow to have affected the emplacement of the stock.

Furthermore, of course, the transection by the Trout Creek fault of the fault scales to north and south extended to great depths—as far as the Roberts thrust—toward the west but to shallower and shallower depths toward the east. As seen in sections *8–8′* and *9–9′* of plate 2, the Shoshone Summit fault is probably not affected by the Trout Creek fault. Neither is the Slaven Canyon fault where it is present east of the present outcrop of the Bateman fault.

While the overturn was forming, the translation of the upper plate with respect to the autochthone in the area north of the Trout Creek fault was shifted from the Roberts thrust to successively higher branches of that fault: the Bateman, Hilltop, and Crum Canyon. These probably formed in the succession suggested as the squeezing of the upper plate became more and more intense in the synform east of the overturn. When the overturn reached essentially its present form the translation was shifted to still higher faults, the Chert Ridge and possibly the Crippen Canyon. From this time on, the main Roberts thrust in the northern Shoshone Range must have been stagnant for several miles to the east of the junctions of these branch faults with the main surface. Although bodily translation of the entire upper plate may have continued farther west (or even farther east also), in this area only that part of the upper plate above the highest of these branches could have continued to move significantly with respect to the autochthone.

Our analysis of the structural evolution had reached this stage when it was pointed out by our colleague, Robert E. Wallace of the U.S. Geological Survey, that a necessary corollary of the diversion of translation from the facies-bounding Roberts fault to the north of the Trout Creek fault was a similar diversion to the south. Otherwise a major torque in the upper plate along the line of the Trout Creek fault is demanded. As it is quite clear that the Trout Creek fault cannot extend down to the Roberts thrust for the eastern several miles of its exposed length, it is necessary to postulate that the part of the Roberts thrust to the south of the Trout Creek fault must have also have become inactive soon after the Goat Ridge upwarp began to develop. There must be one or more branches of the Roberts thrust to the south of the Trout Creek fault trend to fulfill a function similar to that of the Chert fault farther north.

Although we are somewhat uncertain as to the junction of the Mill Creek fault with the main Roberts thrust, this fault appears to be the only one in the southern complex whose topologic relations allow it to have served this function. We have therefore assumed that the Mill Creek fault began to develop late in the upwarping of the Goat window antiform and evolved in such a way as to cut off the group of Goat Ridge and related faults such as the North Fork, Feris Creek, and Slaven Canyon. These subsidiary faults of the upper plate thus became inactive just as the Bateman, Hilltop and Crum Canyon did to the north of the Trout Creek fault.

These are the relations diagrammatically shown in figure 36.

DISPLACEMENT

No single structure can be identified both north and south of the Trout Creek fault, so that its displacement cannot be measured. It clearly formed concurrently with the development of the antiform of the Goat window and the synform that trends north-south through Mount Lewis, and it permitted the differential adjustment of the rocks involved in these isoclines with those in the less complexly deformed thrust sheets to the south. The apparent displacement along the fault must therefore differ greatly from place to place, both in the amount and sense of the differential slip.

The displacement at the west end of the Trout Creek fault must involve a left-lateral component to be measured in several thousands of feet. There must be a comparable vertical component down on the north near the head of Trout Creek on the trend of the synform through Mount Lewis. As mentioned, the depth of this downwarp shown on the sections is very conservative; it probably is considerably deeper than shown. On the other hand the displacements on the Bateman, Hilltop, and Crum Canyon faults imply—if the hypothesis outlined above is correct—a right-lateral slip on the Trout Creek fault to the east of the synform. It seems necessary, therefore, to assume that the upwarp of the Goat window antiform migrated eastward as it was developing and compacted the thrust sheets to the north of the Trout Creek fault into isoclinal folds. At the same time the originally identical thrust sheets to the south at first ran more freely but eventually also became inactive as higher branches such as the Mill Creek fault carried the translation to higher structural levels. The differential distortion of the thrust sheets on the two sides was permitted by the varied displacements along the Trout Creek fault.

FAULTS NORTH OF THE TROUT CREEK FAULT

BATEMAN, HILLTOP, AND CRUM CANYON FAULTS

The Bateman, Hilltop, and Crum Canyon faults are each named from areas where they are conspicuously developed: Bateman Canyon, the Hilltop mining area, and Crum Canyon, respectively. As seen on plates 2 and 6, all are smoothly arcuate in plan, terminating at the Trout Creek fault on the south and the frontal fault (Basin Range fault) on the north. It is practically certain in the case of the Bateman fault and highly probable for the others, that they are not cut off by the Trout Creek fault, but originally formed with this structure as their southern limit; they were always confined to the area north of the Trout Creek fault, though their offset continuations undoubtedly extend far to the north of their intersections with the much younger Basin Range fault.

Their arcuate outcrop patterns, with essentially uniform centripetal dips, imply a troughlike downwarp of the several thrust sheets. This impression is fortified, and a westward closure of the trough is suggested, by the pattern of the Kattenhorn and Dean Mine faults at higher tectonic levels. (See p. 116–117.)

If this geometry is correct, it implies that the faults are successively higher branches, each of the next lower fault, and that the Bateman fault is a branch of the Roberts. This is the interpretation shown both on plate 2, sections *C–C′* through *G–G′* inclusive, and in figure 36. By analogy with the surface pattern at the overturn of the Roberts thrust in Trout Creek Canyon, as illustrated on plate 2 and in figure 34, we consider it probable that these near-parallel branches spring from fold axes in the parental Roberts fault. For this reason we show the fault inflexions in the lettered structure sections and the acute nodes in the contours of figure 29.

The displacement along these faults can only be surmised. The rocks above and below the Bateman fault are stratigraphically widely separated throughout its length, but those in the hanging wall, at least, lie nearly parallel to the fault. The cutting out of the entire Elder, at least 2,000 feet of Slaven Chert and of several thousand feet of the middle and upper parts of the Valmy, suggests a stratigraphic throw of at least a mile, perhaps 3 miles. As the fault is nearly a bedding fault, at least with relation to the hanging wall, a displacement several times the amount of the stratigraphic throw is almost certain.

The displacement on the Hilltop fault need not have been so great. The rocks on the two sides are similar enough to have been closely associated prior to the faulting. Nevertheless, the transection of individual belts of strata and the near-parallelism of bedding of both footwall and hanging wall to the fault itself imply a movement of many thousand feet.

The Crum Canyon fault is again like the Bateman in that the rocks on the two sides are entirely unlike throughout its length. The rocks in the hanging wall west of Crum Canyon are of Silurian age; the displacement may have been moderate, though still measured in many thousands of feet.

KATTENHORN FAULT

The Kattenhorn fault is named from the now abandoned Kattenhorn mine on the east side of Mount Lewis, whose portal lies just beneath the fault. The fault separates rocks of the Valmy below from an overriding sheet of Elder Sandstone 1,000 to 1,500 feet thick.

At the Kattenhorn mine the fault is marked by a zone of quartzite and sandstone of the Valmy, about 200 feet wide and standing nearly vertical. This zone continues northward for about 2 miles to the break in the ridge west of Crum Canyon. Here the fault becomes a single clean-cut break that curves more toward the northwest and flattens somewhat in dip. To the west (hanging wall) of the fault in this segment, the Elder Sandstone stands nearly vertical. The fault can be followed about halfway down the steep north front of the ridge where it is cut off at a moderate angle by another steeply dipping fault that carries Valmy sandstone on its hanging wall. This fault is discussed in connection with the Dean Mine fault.

The vertical breccia zone marking the fault continues south of the Kattenhorn mine for about a quarter of a mile to a point where it is cut off by a high-angle fault that trends about N. 40° W. This high-angle fault is downthrown to the northeast, judging from the offsets of higher thrust scales, but because it is restricted to three thrust slices and does not cut higher ones, it is considered a feature formed by strike-slip rather than dip-slip movement, contemporaneous with the thrusting.

The Kattenhorn fault is concealed for more than a mile to the south of this transcurrent fault by volcanics, dropped down on a normal fault of middle or late Tertiary age. Where it emerges from cover it still stands nearly vertical as a clean-cut fault, but farther south gradually flattens its dip to about 40° (see pl. 2, section 7–7′). Just north of the head of Trout Creek it swings in a gentle arc through a full half circle with a radius of about half a mile. The fault steepens on the southwest segment of this arc and where its strike has become north again, it is overturned, dipping steeply west, with Valmy on the hanging wall and Elder on the footwall. (See pl. 2, section *F–F′*.) This demonstrates that the fault with its overriding plate of Elder Sandstone has been folded into an isoclinal downwarp.

The fault can be followed northward on the west side of this downwarp for about a mile and a half to a point on the north wall of Crippen Canyon southwest of Mount Lewis, where it abruptly flattens, through a radius of a few score feet, in a sharp upfold and then trends southwest, west, and northwest, in a shallow downwarp just beneath the Whisky Canyon fault. In this downwarped segment it carries Valmy Formation on its upper plate. Half a mile west of the abrupt upfold, the fault is overridden by the Whisky Canyon fault, which we consider much younger and as belonging to the post-China Mountain(?) orogeny. It would of course be possible to interpret this westerly downwarp in the fault as belonging to a lower strand of the Whisky Canyon fault rather than to the Kattenhorn, but the comparable patterns of lower splits of the Kattenhorn fault, next to be described, incline us to believe that the Kattenhorn fault itself, as well as these lower strands, has been sharply folded.

The northern segment of the Kattenhorn fault, north of the mine from which we have named it, stands at a considerably steeper angle than the bedding in the Valmy of its lower plate, though the strikes are not far from parallel. South of the volcanic cover, however, the strikes diverge and higher beds appear beneath the fault as it is followed southward. Although there is obvious brecciation along both bottom and top of the quartzite member of the Valmy shown on plate 1, we do not consider these as representing significant fault surfaces.

But on the southwestern arc of the plunging downwarp at the head of Trout Creek, an obvious fault diverges at a high angle from the Kattenhorn fault.

This fault (and some minor splits from it) trends west and northwest to cross the Trout Creek-Crippen Canyon divide, then drops into Crippen Canyon in a deep, east-pointing "V" whose northern limb is cut off near the Crippen Canyon-Horse Canyon divide by the Whisky Canyon fault.

This lower fault, here considered a branch of the Kattenhorn fault, cuts Valmy rocks on both walls. The rocks of the footwall are not cut at a very large angle, and for most of the traceable length of the fault are probably within a few hundred feet, stratigraphically, of the horizon at which the branch fault joins the main Kattenhorn strand. But the strata of the hanging wall are cut so abruptly that a wedge about 4,000 feet thick is present between the two strands in the valley of Crippen Canyon. It seems very likely that simultaneous movement on both branches of the Kattenhorn fault might have dragged this thick wedge off the lower strand and induced the sharp fold in the upper strand which is seen on the north wall of Crippen Canyon. As is shown on sections (pl. 2, *E–E′*, *F–F′* and *G–G′*), the fault plates intervening between the Kattenhorn fault and the Roberts thrust west of the longitude of Mount Lewis must thicken comparably at lower tectonic horizons as well as between the faults here considered strands of the Kattenhorn fault itself.

Throughout the exposed length of the main Kattenhorn fault the bedding in the Elder Sandstone of the upper plate is roughly parallel to the fault. The upper plate of the fault is thus in a synform whose axis about coincides with that of the major downwarp in the fault, north of the Trout Creek divide.

DEAN MINE FAULT AND ASSOCIATED STRUCTURES

The portal of the Dean mine, on the east side of the ridge between Lewis and Crum Canyons, lies a few score feet in the hanging wall of a fault we here name for the mine.

Below the mine this fault carries quartzite and chert of the Valmy over the Elder Sandstone of the upper plate of the Kattenhorn fault. The strike is essentially parallel to that of the Kattenhorn fault below, but the dip is much gentler, perhaps 30° to 50°, rather than nearly vertical. The widening of the belt of Elder Sandstone in the footwall in each valley and its narrowing on each ridge constitute the most conspicuous evidences both of the fault and of the discordant dips. To the north of the eastern adit of the Dean mine the fault is readily traced by the quartzite in its hanging wall, which stands as a ledge above the gentler slopes carved on the Elder Sandstone.

About a mile north of the mine, the gentler dip of the Dean Mine fault carries it west of the persistently near-vertical north-trending slice of the Elder Sandstone; a fault with Valmy to the west (hanging wall) emerges from beneath the Dean Mine fault and bounds the Elder on the west. This fault, to which we have attached no name, is the one referred to in the description of the Kattenhorn fault as overriding it and cutting off the plate of Elder Sandstone about a mile and a half farther north on this ridge (p. 116). Our interpretation, as shown in sections *C–C′*, *D–D′*, *E–E′*, and *F–F′* of plate 2, is that it is virtually parallel to the Kattenhorn fault and overrides the Elder Sandstone in the hanging wall of that fault. The Dean Mine fault in turn overrides both faults, the intervening plate of Elder Sandstone and a higher plate of Valmy rocks. Although warped along the same axes as the Kattenhorn and the unnamed fault, the warping is far less intense.

From the point where it leaves the Elder Sandstone in its footwall, the Dean Mine fault can be followed westward and northwestward down the northwest spur of Bens Peak almost to the mountain front. Part of its course is obscure because of landslides not mapped on plate 1. Throughout this segment the fault brings Valmy over Valmy but is nevertheless easy to follow because of the discordant structures above and below: the ledges above slope so as to indicate a moderate dip to the west; the beds below are very steep or vertical.

The Dean Mine fault can be followed with its westerly dip down the northwest spur of Bens Peak to the 6,000-foot contour where it abruptly flattens into a sharp synform and then dips steeply east at angles near or even exceeding 60°. Here it carries massive quartzite of the Valmy in its hanging wall, resting on Valmy sandstone with much gentler easterly dips. The fault can be traced, with these same characteristics southward along the right wall of Lewis Canyon for nearly a mile and a half. Here it dips less steeply and curves abruptly across the canyon with a dip of about 40° S. The axis of the antiform lies a few hundred feet west of the canyon; from it the fault, dipping 40° W., can be readily traced northward. It is conspicuous, with massive quartzite in the hanging wall standing above smoothly worn slopes of sandstone and siltstone below.

At the northeast spur of Lewis Peak the Dean Mine fault flattens and can be walked out along a contour to the northwest spur where it crosses another synformal axis. From here it can be followed southwestward with a steep easterly dip for several hundred yards, to the point where it is cut off by a quartz porphyry dike and then presumably buried by Tertiary sedimentary rocks. Undoubtedly the fault was originally cut off and overridden by the Whisky Canyon fault in this direction.

To summarize, between the east portal of the Dean mine and the west side of Lewis Peak, the Dean Mine fault shows two synforms and an intervening antiform, all striking roughly north-south. The total relief of the fault outcrop (because of plunges not a true measure of the relief of the fault surface) is about 1,600 feet. The easterly one of these synforms has its axis passing beneath Bens Peak and Mount Lewis; the westerly one has an axis through Lewis Peak; the intervening antiform axis passes down the west fork of Lewis Canyon.

To the south of the eastern adit of the Dean mine the fault is offset by the same transcurrent fault described in connection with the Kattenhorn fault. The offset is about half a mile to the right. Thence the Dean Mine fault can be followed south, with persistent quartzite of the Valmy in its hanging wall and Elder Sandstone in the footwall to the end of the south spur of Mount Lewis. Here, like the Kattenhorn fault and the plate of Elder Sandstone below, it swings around a synformal axis. The dip on the east flank of the downwarp averages perhaps 30°; that on the west seems much steeper but is difficult to measure. From this axis the fault is readily traced northward to the head of Crippen Canyon where it is cut off and overridden by the Whisky Canyon fault. Thus, only a single synform is preserved in the Dean Mine fault to the south of Mount Lewis.

Everywhere south of Bens Peak the exposed footwall of the Dean Mine fault is made up of Elder Sandstone. The dips of the rocks above it are, however, so much gentler than the dips of the Elder Sandstone and Kattenhorn fault that we infer that throughout this area the Dean Mine fault overlies and has cut off another thrust sheet of Valmy sandstone, about conformable to the isoclinally folded Elder, such as is seen on the north spur of Bens Peak. Such a folded and transected sheet is shown in sections *B–B′*, *C–C′*, *D–D′*, *E–E′*, and *F–F′* of plate 2.

BENS PEAK AND ASSOCIATED FAULTS

The thrust sheet in the hanging wall of the Dean Mine fault is in turn cut off by a roughly parallel higher thrust, here called the Bens Peak fault, from its exposures below Bens Peak. Above the Dean mine adit on the east side of the ridge connecting Bens Peak and Mount Lewis, this fault brings a conspicuous ledge of highly contorted chert over quartzite and sandstone. All these rocks are of the Valmy Formation.

The transcurrent northwest-trending fault that offsets the Kattenhorn and Dean Mine faults likewise offsets the Bens Peak fault toward the southeast end of its outcrop. But this fault should again intersect the Bens Peak half a mile farther northwest and here there is no offset. This is part of the evidence from which we conclude that the northwest-trending fault formed as a transcurrent fault during the thrusting episode.

South of the transcurrent fault, the Bens Peak fault is not continuously traceable. The fault so labeled in sections *E–E′* and *F–F′* of plate 2 is at about the same tectonic level as that of the Bens Peak fault to the north but carries quartzite rather than chert in its hanging wall.

On the north spur of Bens Peak the trace of the Bens Peak fault, with chert on the upper plate, roughly parallels that of the Dean Mine fault. Like that structure, from which it is separated by only a few score feet of rock, it curves in a synform and can be traced from the axial bend southward high on the wall of Lewis Canyon. West of Bens Peak, however, the two faults diverge and a fault wedge composed of quartzite and sandstone comes in below the Bens Peak fault.

On the northeast side of the first spur north of Dean Canyon the Bens Peak fault, with chert in the hanging wall, is offset for a few feet by a high-angle, northwest-trending fault like that which cuts it to the south of the Dean mine eastern adit. The two high-angle faults are nearly parallel but en echelon. This one dies out to the southeast, the other to the northwest. Even though the Bens Peak fault has been displaced only slightly, the underlying fault wedge of quartzite shows a vertical offset of about 400 feet. (See fig. 37.) This feature also suggests that this high-angle fault, like the one earlier mentioned, was formed during the thrusting, probably by differential horizontal rather than vertical movement.

Owing to landslides, the Bens Peak fault cannot be walked out around the head of the alcove below the

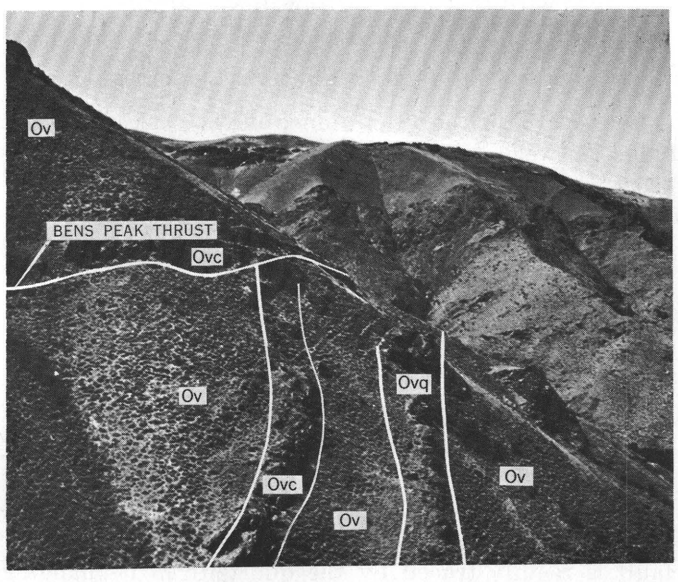

FIGURE 37.—View looking west from the northeast wall of Dean Canyon along the shoulder of the southwest wall. Horizontal ledge at left is chert forming the hanging wall of the Bens Peak thrust. Ledges below it, rising from left to right, are respectively chert and quartzite which are cut off at the top by the horizontal chert ledge. Hillside in distance is on west side of Lewis Canyon and exposes unnamed thrust sheets of chert (rough) and sandstone (smooth) of the Valmy, all dipping westward beyond the antiform of Lewis Canyon. Ov, sandstone; Ovc, chert; Ovq, quartzite, all of the Valmy Formation.

west adit of the Dean mine, in Dean Canyon. There is however, little doubt of its continuity here and thence it can be traced to the northwest on the left wall of Dean Canyon. A higher plate of quartzite of the Valmy here rides down and cuts off the chert which nearly everywhere to the northeast has made the hanging wall of the Bens Peak fault.

The south wall of Dean Canyon just below the west adits ot the Dean mine gives one of the clearest examples of thrusting that can be found in the area of this report. As shown in figure 38, the Bens Peak fault here clearly cuts off the east-dipping beds of the underlying fault wedge along a horizontal plane.

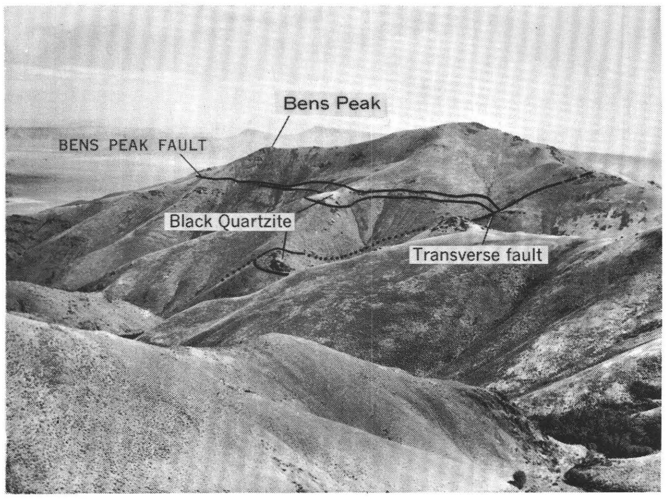

FIGURE 38.—Bens Peak from the southwest, looking across Lewis Canyon. The apparently horizontal ledge a third of the way below the summit to the right center of the view is chert that marks the hanging wall of the Bens Peak fault. The light outcrop on the prominent central spur is of quartzite below the Bens Peak fault. It is downdropped by the transverse northwest-trending fault to form the lower ledge on the same nose.

From the spur just described west of the Dean mine west adit, the Bens Peak fault can be traced down to Lewis Canyon where it crosses an antiform, dips west at fairly steep angles, and can be followed northward on the west wall of Lewis Canyon to Lewis Peak. Here it again closely approaches the Dean mine fault and, like it, can be traced around a synform and thence southwest for a few hundred yards until it is buried by the Tertiary sedimentary rocks just east of the Whisky Canyon fault. Like the Dean mine fault, the Bens Peak was undoubtedly cut off by the Whisky Canyon fault prior to burial by the Tertiary rocks.

Two higher thrust plates are exposed on the ridge crest southeast of Bens Peak, above the fault just described. One, here unnamed, carries a body of west-dipping quartzite and chert over undifferentiated Valmy rocks that here form the upper plate of the Bens Peak fault. This, in turn, is overridden by a plate of Battle Conglomerate that we consider to belong to the Havingdon Peak thrust, of the Lewis orogeny. It is discussed on page 124.

In summary, the Dean Mine and Bens Peak faults are moderately discordant with each other and more so with the Kattenhorn. All, though, are folded along axes that trend north and all plunge more or less parallel. They therefore are intermediate between the two idealized patterns illustrated in figure 35. We consider it highly likely that some of their folding was concurrent with the thrusting but it is also likely that part was induced by the overriding Havingdon Peak and Whisky Canyon thrust plates of the second, Lewis orogeny.

FAULTS SOUTH OF THE TROUT CREEK FAULT

GOAT FAULT GROUP

South of the Trout Creek fault the Roberts thrust crosses the ridge west of Goat Peak northeast of the Shwin Ranch and can be readily traced eastward as it gradually descends to the level of the North Fork of Mill Creek. Above it are thin sheet after thin sheet of rocks of the Valmy, forming a pile of fault slivers, many so thin as hardly to be mappable, even on the relatively large scale of plate 1. To these we have given the name Goat Ridge fault group. Although many individual slivers are traceable for more than a mile, some even for 2 to 3 miles, most can be followed only for shorter distances. The assemblage of thin slivers resembles that near the overturn in the Roberts thrust in Trout Creek and it seems likely to have been formed in the same way: that is, by the successive piling up of thin slices in the lee of a growing fold in the major thrust. With the growth of the fold, successively younger movements tend to follow straight courses and thereby to transect older splits from the main fault as they pass over the axis of the upwarp.

As the Roberts fault is traced around the east and southeast sides of the Goat window, lower and lower thrust slices diverge from it. Furthermore, additional wedges come in at horizons within the Goat Ridge fault group itself. (See map, pl. 1.) As a result, on the Middle Fork of Mill Creek a tectonic horizon, such as the upper strand of the Goat Ridge fault system, is separated from the Roberts thrust by more than a thousand feet of thrust slices not represented on the Trout Creek-Mill Creek divide. (See pl. 2, section *6–6′*.) Just north of Mill Creek the divergence of this horizon from the Roberts fault is several thousand feet more. As was pointed out in the discussion of the shape of the Roberts fault and the contour map of figure 29, such a wedging in of additional thrust sheets in the downwarps of the Roberts fault cannot everywhere be as well evaluated as here; the sections of plate 2 and the contour map of figure 29 all probably represent the maximum possible altitudes of the Roberts fault in the downwarps. The actual relief of this thrust surface is therefore probably considerably greater than is shown on these sections and on the contour map.

SHOSHONE SUMMIT FAULT

An area most critical to such a structural synthesis as is attempted in this report is that on both sides of the Shoshone Range divide at the head of the South Fork of Mill Creek. Many cuts near here along the road from Mill Creek to the Greystone mine show colluvium as much as 10 feet thick. Consequently, although outcrops suffice to show in a general way which formations are present, they do not permit accurate tracing of many contacts nor the determination of the mutual relations of many contiguous fault blocks. The interpretations shown on the map and sections (pls. 1 and 2) seem to us consistent with the available data, but we concede that they are not compelled by them.

It is clear that the lower members of the east-dipping Goat Ridge group of thrust slices are cut off along Mill Creek just above the junction with North Fork, by the Mill Creek fault (p. 121). Some superposed sheets are cut off along Middle Fork by a north-dipping fault that emerges from the Roberts thrust where it bounds the small window in the valley of that stream.

Higher members of the Goat Ridge group curve southwestward in rough conformity with the curve of the Roberts thrust around the southeast end of the Goat window. North of the divide at Mill Creek Summit they are overridden by the Mill Creek fault. They appear to be, and are here interpreted as being, cut off just south of the divide by the fault here called the Shoshone Summit fault.

The Shoshone Summit fault is cut off at the west by a hanging wall split from the Mill Creek fault or, perhaps more likely, by a higher fault. The southeast-dipping thrust slivers of the higher members of the Goat Ridge group end abruptly and a wholly different structural pattern prevails to the south and east, in the drainage areas of Cooks, Elder, and Feris Creeks. The boundary between these two structural subprovinces is made by the Shoshone Summit fault.

From the head of Cooks Creek, where it is cut off on the west, the Shoshone Summit fault extends east-southeast for nearly a mile. Its exposures here are so poor that the dip cannot be determined but is probably steep to the north. Its trace then curves northeastward and climbs the east side of the summit ridge of the range. It almost reaches the summit in the saddle at the head of the South Fork of Mill Creek, but remains on the east side of the divide as far north as the ridge spur that trends northeast to form the divide between the headwaters of Feris Creek and the North Fork of Mill Creek. Through most of the distance from the northeast bend to this point the fault is commonly exposed and dips very steeply to the west-northwest, but from a point near the saddle at the head of Middle Fork of Mill Creek the dip flattens. At the east end of the high spur between the headwaters of North Fork and Middle Fork of Mill Creek the fault actually dips east—its trace descends more than 400 feet as it is followed eastward on this spur. The axis of the fold in the thrust revealed by this dip reversal trends east of north, just west of the saddle at the head of the North Fork of Mill Creek.

From the ridge spur southwest of this divide the trace of the Shoshone Summit fault descends westward and northward into the valley of North Fork where it passes beneath the Feris Creek fault. It is considered nearly certain that the fault near the head of the most northerly fork of Feris Creek is the Shoshone Summit fault. This fault emerges from beneath the same block of Slaven Chert that forms the upper plate of the Feris Creek fault farther west. It dips northeast, so must here be on the east side of the fold axis just described. It is marked by a thick ledge of quartzite, as is the Shoshone Summit fault on the south wall of North Fork, and overlies the same block of undifferentiated Valmy.

From the point of its emergence from beneath the North Fork fault, the Shoshone Summit fault curves southeastward into the valley of Feris Creek. It dips about 30° to 40° NE. in this segment. (See section *I–I'*, pl. 2.) It crosses the creek and cuts across the Valmy, the upper split of the Elder fault, and the higher plate of Slaven Chert.

The fault trace passes southward beneath the alluvium of Feris Creek and thence southeastward cannot be certainly identified. It is considered, however, to join the Feris Creek and Slaven Canyon faults, which also can be traced into the same belt of alluvium, and to pass south of the isolated hill of Valmy that forms an island in the alluvium in the triangle made by the road up Feris Creek and its branches to the Utah Mine Camp.

The fault that follows the east side of the north-south ridge east of the Utah mine is considered an extension of the Shoshone Summit-Feris Creek fault. Other outliers in the country between Bald Mountain and Gold Acres are at about the same tectonic level.

The most instructive sections for illustrating our interpretation of the relations of the Shoshone Summit fault are those lettered from *F* to *L* on plate 2. From these it can be seen that we regard the Shoshone Summit fault as springing from the Roberts thrust along a line east of the junction of the Bateman fault with the

Roberts. The tight antiform shown in sections $I–I'$ and $J–J'$ is the one described at the head of Feris Creek. No particular reason for its localization is apparent. But the strongly overturned folds in the thrust sheets immediately beneath the projected level of the fault (see sections $K–K'$, $L–L'$, and $M–M'$) suggest that they may have been formed by the drag of the upper plate of the Shoshone Summit fault as it overrode them.

MILL CREEK FAULT

The Mill Creek fault can be followed with confidence only from a point west of the Mill Creek window to Mill Creek Summit. At the west it is cut off against a steep fault whose relation to the thrusting is obscure. We consider it likely that it is a branch of the Roberts thrust, as suggested on plate 6 and in figure 36. It practically parallels the upper course of the South Fork of Mill Creek from the junction with North Fork to the Summit. Throughout this distance it carries Slaven Chert in the upper plate, which overrides several thrust sheets belonging to the Goat Ridge group. (See section $J–J'$ near coordinate lines 4 and 5.)

In some respects the relations of the Mill Creek fault are similar to those of the Feris Creek fault, next to be described, and we once tried to identify the two as a single fault. They cannot, however, be the same. At the creek level just north of Mill Creek Summit, the block of Slaven Chert in the hanging wall of the Mill Creek fault is concealed by a small thrust slab of quartzite of the Valmy. This is on a probable branch of the Mill Creek fault which trends southward up to Mill Creek Summit with a nearly vertical dip. The lower branch of the Mill Creek fault continues eastward with a south dip at its intersection with the Shoshone Summit fault. The Shoshone Summit fault here appears to cut it off, but there is little doubt that the upper branch of the Mill Creek fault cuts off the Shoshone Summit fault a few hundred yards south of the divide, in Cooks Creek.

This relation may indicate that the Mill Creek and Shoshone Summit faults were in part simultaneously active, but it seems clear that movement on the Mill Creek upper branch continued after both Shoshone Summit and Feris Creek faults became dormant.

FERIS CREEK AND ASSOCIATED FAULTS

As can be seen from the map and sections (pls. 1 and 2), the Feris Creek fault abuts the Trout Creek fault. But it also appears to the north of the fault, east of the Bateman fault. These relations suggest that it is an older fault than the Trout Creek and Bateman and that it became inactive after the Bateman block began to move. It curves in a somewhat more open arc than the Shoshone Summit fault and, as already mentioned, cuts that fault off on the crest of the plunging antiform near the North Fork-Feris Creek divide. It then diverges eastward from the Shoshone Summit fault for about 2 miles but seems to join it under the alluvial cover along Feris Creek. To the east and southeast we have considered the two faults as coincident and where two faults are identified at about this tectonic level we have arbitrarily called the lower, "Shoshone Summit" and the upper, "Feris Creek."

The faults labeled Slaven and Indian Creek on plate 6 seem to require little discussion. Except between the Utah mine and Gold Acres, they everywhere seem to be only moderately folded or warped and to lie roughly parallel with the Feris Creek fault. Like that fault they are probably chiefly older than the Trout Creek-Bateman fault movement. None of the three seems to join the Roberts thrust; all seem wholly within the upper plate.

FAULTS BENEATH THE SHOSHONE SUMMIT FAULT

The northward-plunging upwarp outlined by the Shoshone Summit fault contains some of the most complexly folded thrust sheets in the area. The lowest tectonic levels exposed are the massive quartzites on the west fork of Elder Creek. Two faults, with a thin sliver of Slaven Chert between them, dip steeply away from the core mass of quartzite of the Valmy. The tectonically higher fault diverges at the axis of the fold to produce a "tail" of chert projecting southeastward into the tectonically higher mass of the Elder Sandstone. It is this tail that makes the peculiar offshoot to the surface shown in section $8–8'$ of plate 2, which cuts this overturned antiform. On the map of plate 6 and sections of plate 2, the fault separating the Slaven from the Elder is called the Utah Mine fault. The chert belt structurally beneath (west of) the Utah Mine fault can be followed north for about 2 miles to a point west of the Utah Mine Camp. Here, as the projection of the Shoshone Summit fault is approached, the whole structure is overturned, the chert sheared out, and the Utah Mine fault overturned until it lies flat or even dips slightly eastward, beneath the quartzite of the same block that farther south lies beneath the Slaven. The overturn continues to the alluvium in the open park near the forks of Feris Creek. Beneath the alluvial cover, however, the fault must again be twisted through the vertical to a steep east dip, for an arrangement of Elder, Utah Mine fault, Slaven Chert, the unnamed fault and finally quartzite of the Valmy east of the Utah Mine, symmetrical with that near the west fork of Elder Creek, strongly implies a synform between them. A synformal curve of the Utah Mine fault and its associates almost surely must be concealed by the alluvial fans of Feris Creek.

West of Elder Creek the faults above and below the vertical belt of Slaven Chert converge and cut it out. Half a mile to the north, this fault and the rocks of both adjacent blocks are overridden by a steeply northdipping fault, not named on plate 6.

This north-tilted thrust plate is in turn overridden by the lowest of three splits of the Elder fault. The Elder fault shows only a slight antiformal warp south of, and on the trend of the overturned isoclinal fold in the Utah Mine fault exposed on the west branch of Elder Creek. But when followed to the east, it dips about parallel with the eastern flank of the synform south of the Utah mine. This is along a high-angle thrust which is inferred to be a segment of the Utah Mine fault that was reactivated after folding of the thrust piles and, late in the orogeny, cut and offset all the higher thrust plates from a point about a mile south of the Utah mine to and beyond the border of the map area. (See pl. 2, sections *L–L'*, *M–M'*, and *N–N'*.)

Still higher fault strands lie in the hanging wall of the Elder fault, notably the Greystone fault zone and other unnamed ones above that. All these are considered to lie structurally beneath the Shoshone Summit fault.

The north plunge of the synform of the Shoshone Summit fault implies that the fault was at higher levels over the south edge of the map area than it was farther north. Perhaps this is responsible for the less complex warping of the higher thrust sheets toward the south. The complexities increase the closer the antiform in the Shoshone Summit fault is approached. Possibly this implies that much of the complex structure in the heart of this upwarp was formed at the time the fault was warped.

SUMMARY OF THE ANTLER OROGENY AND ITS RESULTS

The regional pattern of facies of the rocks of early and middle Paleozoic age has been described both in the stratigraphic and structural sections of this report, and is summarized in figure 28. The clear implication that the upper plate of the Roberts thrust has traveled more than 55 miles seems inescapable.

The direction of travel is less certain. Some elements, such as the generally northwestern trend of the windows over much of the extent of the fault plate, perhaps suggest a northeasterly component of relative motion in the general area. A similar trend here might be given the same interpretation. On the other hand, the course of the Trout Creek fault, certainly active during the Roberts faulting, suggests that the travel was almost due east. So, too, does the generally north-south trend of many of the folded thrust sheets such as the isoclinal sheet of Elder Sandstone riding on the Kattenhorn fault, and the tight folds in the Elder Creek area. It seems reasonable to conclude that the upper plate moved relatively nearly due eastward with respect to the lower.

The complex faulting described above shows that during the travel of the Roberts thrust the upper plate was sheared on every scale between one of miles and one of scores of feet. Not only were these slivers of the plate intimately interleaved in an almost random pattern by faulting, but they were, along with the Roberts thrust itself, highly folded, locally to the stage of overturned isoclines. Folds like those of the Kattenhorn and associated plates of the Mount Lewis area may involve several thrust sheets, but as the same area shows, the folds in the upper plate need not be directly related to or coincident with the folds of the major fault surface.

All these facts seem to demand that the upper plate of the Roberts thrust has moved completely independently of the basement rocks. No rocks older than Cambrian are known for many hundred miles to the west. The evidence of "structural ungluing" (décollement) seems inescapable.

The problem remains as to the cause of the movement. Certainly it seems inconceivable that a "push from behind" could have carried the plate more than 50 miles wide—the distance we know it to have moved. A gliding under gravitational forces—that is, a movement of the plate by body forces—seems demanded by the thorough shearing and folding of the upper plate.

But it seems fully as difficult to find an adequate slope for gliding so great a distance as it is to provide for distant transmission of horizontal stress, unless some way of lowering the coefficient of sliding friction can be found.

The suggestion by Rubey and Hubbert (Hubbert and Rubey, 1959; Rubey and Hubbert, 1959) that the buoyant effect of confined pore water may reduce the load of the upper plate so that a very small shear stress is adequate for movement is worth much study. In favor of its application here may be cited the common tendency for the upper plates of many of the thrusts to have their bases nearly parallel to bedding. Any aquatards in the sheet would thus act uniformly over a considerable area of the plate. On the other hand, the highly sheared and broken character of the whole upper plate makes it seem a poor medium for producing a high fluid pressure (easy leakage).

The interesting suggestion has been applied to the northern Appennines by Migliorini (1952), and further elaborated by Merla (1952), that the slopes necessary for gravitational gliding were developed by the migration of successive axes of uplift from west to east, each such moving axis giving rise to a gravitational slide.

Such a process may indeed have been responsible for such composite sediments as the *argile scagliosa* of Italy, but it does not seem to us to offer much promise as an explanation of the structures here, except perhaps for some of the isoclinal folding of the thrust sheets in and east of the Goat Ridge antiform. The main objection to the general application of the theory is that we are not here dealing with surficial movement, at least as recorded by contemporary conglomerates involved in the advancing thrusts. Such contemporary conglomerates are common in the Gabbs Valley Range, Nev. (Ferguson and Muller, 1949, p. 25–29) and in the Italian localities studied by Migliorini and Merla, but so far as known are lacking here. All the deformation in the northern Shoshone Range seems to have taken place under thick overburden. Further, the pattern of windows, almost randomly distributed over the known area of the Roberts thrust, does not seem expectable on the theory of migrating uplifts; the Italian examples are more systematic and persistent in strike.

However formed, the structural relief of the area, and almost surely the topographic relief at the end of the Antler orogeny, must have been very great. We have no reason to believe that the relief was entirely reduced by Pennsylvanian time when the Battle Conglomerate was deposited in areas to the west, for the Battle is a coarse sediment. No upper Paleozoic or Triassic rocks are known in the Mount Lewis area that could have been deposited here, though it seems likely that the Brock Canyon Formation is parautochthonous in the Cortez Mountains. The general similarity but local differences in the sections of upper Paleozoic rocks in the general region and the sporadic conglomerates they contain suggest that the relief may not have been wholly reduced until much later. The Lewis orogeny, next to be discussed, may have affected a region of considerable topographic diversity, perhaps comparable to that of present-day coastal California where orogenic movements are now going on.

THE LEWIS OROGENY

All the structural features hitherto discussed are attributed to the Antler orogeny of probable Early Mississippian age. Several widely separated structural elements seem to us demonstrably concurrent in origin with the Roberts thrust; we believe that few of those so far mentioned have been modified by later deformation.

Nevertheless there is unequivocal evidence of a second, distinctly later, episode of orogeny. The proof, already outlined briefly, lies in the presence of huge thrust sheets that include rocks of Pennsylvanian, Permian, and Triassic ages—all younger than the Antler orogeny. In the Mount Lewis quadrangle these rocks all are associated with two major thrusts, here called the Whisky Canyon and Havingdon Peak faults. These fault blocks all lie north of the latitude of Crippen Canyon and the most easterly outlier is on Havingdon Peak. That a far wider country was involved in this deformation is, however, shown by the presence of far-traveled facies in the Mount Lewis area and is suggested by the décollement shearing of the Brock Canyon Formation far to the east in the Cortez Mountains, as seen in the southeast corner of the Crescent Valley quadrangle. This tectonic episode we here name the "Lewis orogeny" from Mount Lewis, where structural features belonging to it are well exposed.

WHISKY CANYON FAULT

The Whisky Canyon fault is poorly exposed but obviously present in or near the bottom of Whisky Canyon, all the way from the Basin Range fault at the foot of the range to the point where it is cut by a Tertiary intrusive southwest of Lewis Peak. For about half a mile the fault has been destroyed by the Pipe Canyon caldera, from which it reappears high on the west wall of Lewis Canyon about a mile southwest of the junction of Dean Canyon. Half a mile farther south it is again concealed for a few hundred yards by an overlapping thrust sheet of the Pipe Canyon group. Beyond this it is well exposed on the north side of the southernmost western tributary of Lewis Canyon, but its course farther south is concealed by thick colluvium and can be traced only approximately. Though concealed, the fault must cross the head of Lewis Canyon and be overlapped by the intrusive breccia of Mount Lewis.

Through all this distance the Whisky Canyon fault is identifiable by the presence of Antler Peak, Battle, Havallah, or China Mountain(?) Formations on the hanging wall (west) and of Valmy rocks of various kinds on the footwall (east). Though exposures only locally permit measurement, the trace of the fault over topographic irregularities testifies to a steep west dip, probably at least 50°, for most of this distance. A sharp truncation of steeply dipping beds below by nearly flat lying ledges of greenstone above (see fig. 39) is taken to mark the Whisky Canyon fault on the east side of the head of Lewis Canyon. For some distance to the east the fault is uncertainly discriminated from comparable underlying faults that bring Valmy over Valmy in the general area. Farther east the Whisky Canyon fault carries greenstone in the hanging wall with quartzite and chert of the Valmy intervening between it and the lower, nearly parallel fault.

Owing to much talus, it is impossible to trace the fault continuously across the north face of Mount Lewis. But at about the same altitude on the northeast spur of the mountain there is a flat fault, probably the Whisky Canyon. It can be traced from here

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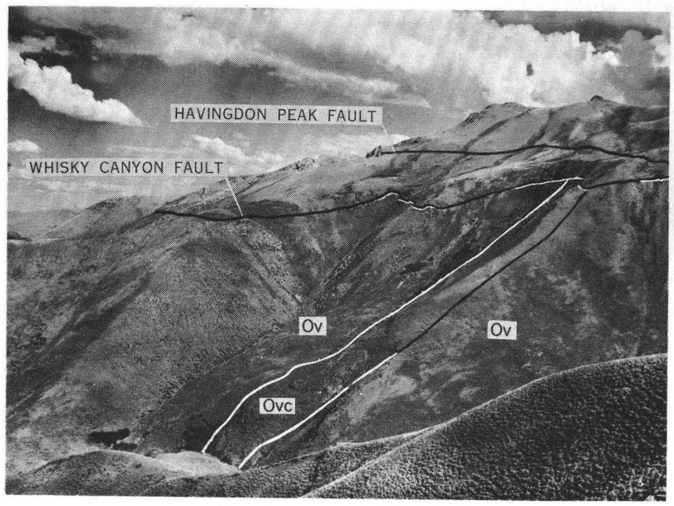

FIGURE 39.—Structural bench on the north flank of Mount Lewis from a point northwest of the cataract near the head of Lewis Canyon. The east summit of Mount Lewis is to the right. The Whisky Canyon and Havingdon Peak faults are shown. The dark crags in the middle foreground are on the lower edge of a nearly vertical chert ledge that can be followed straight up the ridge to a point where it is sharply transected by the Whisky Canyon fault, which brings a flat-lying ledge of greenstone over the projected trace of the chert ledge. Ov, Valmy Formation, undifferentiated; Ovc, chert of Valmy Formation.

south along the east side of Mount Lewis, with chert in the footwall and undifferentiated Valmy in the hanging wall. It is covered by breccia for half a mile, but beyond that a similar fault can be traced across the southwest spur of the mountain. Where it leaves the breccia, this fault places gently dipping sandstone of the Valmy across steeply dipping chert; farther southwest it descends the south side of the spur from Mount Lewis to the Crippen Canyon-Horse Canyon divide, cutting off the Elder Sandstone in the hanging wall of the Kattenhorn fault. The dip steepens notably in this segment.

The trace of the Whisky Canyon fault crosses westward into Horse Canyon drainage but remains high on the slope just below the divide as far west as the west side of the Horse Canyon volcanic center. In this interval the dip has steepened and become very high to the north. The fault cuts off the east-dipping block of Battle Conglomerate and Antler Peak Limestone of its hanging wall on the south side of Horse Canyon. This is one of the localities where a depositional contact of the Battle Conglomerate with the underlying Valmy can be seen. (See p. 43.)

To the west of the Battle Conglomerate, the Whisky Canyon fault again has Valmy Formation on both walls. The trace crosses to the Crippen Canyon side of the divide and, when followed westward, remains high on that slope for about a mile. It then curves sharply northward to the divide again. At the saddle, it is cut off by the Basin Range fault that here swings abruptly westward to the mountain front.

The Whisky Canyon fault thus forms in plan two sides of a triangle, of which the third side is the Basin Range fault system. Both legs of the fault are steeply dipping inward. Where they would be expected to meet, west of Mount Lewis, however, both limbs abruptly flatten to almost horizontal dips and the fault sharply truncates the synform in the underlying fault slivers without deflection. Within the irregular area thus outlined are contained all except three of the known outcrops of upper Paleozoic rocks in this part of the Shoshone Range.

HAVINGDON PEAK FAULT

Before discussing the internal structures of the Whisky Canyon fault block, it seems best to describe the very closely associated feature here called the Havingdon Peak fault.

In many places no distinction may be made between them. But elsewhere two faults may be distinguished; the lower is the Whisky Canyon fault, just described, the upper one is shown on plate 6 and plate 2, sections *5–5′*, *6–6′*, *7–7′*, and *E–E′* as the Havingdon Peak fault.

On both of the converging limbs of the Whisky Canyon fault there is no nearly parallel higher fault. However, in the head of Horse Canyon a fault crosses from one limb of the Whisky Canyon to the other and cuts off the body of upper Paleozoic rocks on the east. This fault seems reasonably to be projected to the level of the fault on Mount Lewis, above the Whisky Canyon fault, which carries the thrust plates of Battle Conglomerate and Antler Peak Limestone. (See pls. 1 and 6, and the sections cited in the preceding paragraph.)

Because this higher fault plate contains little but upper Paleozoic rock, it has seemed best to identify it with the small outliers—klippes—of Battle Conglomerate that are present: (1) on the crest of the ridge north of the Dean mine and southeast of Bens Peak, (2) on the southern ridge from Mount Lewis, about a mile south of the edge of the mass of intrusive breccia, and (3) on the very crest of Havingdon Peak.

In each of these localities the local block of Battle Conglomerate has an attitude wholly inconsistent with its being in depositional contact with the underlying rocks; each is obviously a klippe. If these blocks are, in imagination, united as parts of a single thrust sheet, it appears to have rested on a nearly flat fault all the way from the longitude of the head of Lewis Canyon to Havingdon Peak. In this extent of nearly 4½ miles, the fault cuts the projections of the Bens Peak, the Dean Mine, the Kattenhorn, and Crum Canyon faults in a nearly plane course. This transection of so many underlying thrust sheets, with an aggregate thickness of at least 2 miles, seems another argument to support the stratigraphic evidence that the Havingdon Peak fault belongs to an orogeny wholly independent of that that emplaced these lower sheets.

INTERNAL STRUCTURE OF THE WHISKY CANYON-HAVINGDON PEAK BLOCK

A glance at the map of plate 1 shows how much broken by faulting is the rock mass above the Whisky Canyon fault. Along its eastern side the mass is a jumble of much faulted blocks of Battle Conglomerate, Antler Peak Limestone, Havallah and China Mountain(?) Formations. These are still further confused by a whole series of fault slices, grouped on plate 6 and and on sections *D–D′*, *E–E′* and *6–6′* of plate 2 as the Pipe Canyon fault group. The rocks involved in the Pipe Canyon fault group are all mapped as belonging to the Valmy Formation. There may be some question about this assignment, however, as there seem to be thicker greenstone and shale units here than are common in the Valmy.

The faults of the Pipe Canyon group dip gently eastward into, or perhaps converge and join to make a somewhat irregular west-dipping fault that separates the probable Valmy slices from the jumbled blocks of late Paleozoic and Triassic(?) ages beneath. In places, near the head of Lewis Canyon, this irregular fault allows the higher plates of Valmy to rest directly on the lower plate of the Whisky Canyon fault. The masses of Antler Peak Limestone and associated rocks nearby appear only in windows through these higher plates. (See pl. 1, about half a mile southeast of the Pipe Canyon volcanic center.)

PROVENANCE OF THE TRANSPORTED PLATES

All the upper Paleozoic and Triassic(?) rocks of the Mount Lewis quadrangle are allochthonous. Some blocks carried in on the Whisky Canyon and Havingdon Peak faults, such as those composed of Valmy, Antler Peak, and Battle Formations, could have been derived from the Antler Peak area or from some equally near site now buried by the alluvium of the Reese River Valley. All these formations are autochthonous in the Antler Peak quadrangle. (This is perhaps doubtful as respects the Valmy.) But the Havallah Formation of the Antler Peak quadrangle is itself allochthonous and must have been brought in from the west on the Golconda thrust (Roberts, 1951). The only known autochthonous bodies of Havallah Formation in the region are in the Mount Tobin quadrangle, but so much of the intervening country is covered by younger rocks that it is not necessary to assume such distant provenance for the masses carried here by the Whisky Canyon fault. The travel of the post-China Mountain(?) thrust masses must have been a few miles, but need not have been comparable to that of the Roberts thrust.

EFFECT ON OLDER STRUCTURES

The south limb of the Whisky Canyon fault seems to cut cleanly across the underlying structures, which show no deflection alongside it. On the segment in Whisky and Lewis Canyons, however, the folding of the Dean mine, Bens Peak, and associated fault plates near Lewis Peak seem reasonably related to the drag of the overriding Whisky Canyon fault. It is not likely, however, that the folds of the Kattenhorn fault plates, shown in the isoclinal synform trending through Mount Lewis a mile farther east, are due to drag of this sheet. First, the plunges of the synform of the Kattenhorn sheet seem independent of the plunge of the Whisky Canyon fault but to conform with the plunge indicated by the arcuate trends of the structurally much lower Crum Canyon, Hilltop, and Bateman fault blocks. Second, the flat plate of the Havingdon Peak fault, as expressed by its scattered remnants, crossed several thrust sheets of the Antler epoch without offset. Evidently this thrust sheet exerted but little drag on the underlying rocks.

FAULTING IN THE CORTEZ MOUNTAINS

The near-bedding shearing displayed by the Brock Canyon Formation was mentioned in describing the rocks. The disturbance at the base of the formation is such that any features that may once have shown it to be authochthonous have been destroyed. Nevertheless it seems likely that the formation is not far from its place of deposition, for the beds are generally roughly parallel and enough transitional members are preserved to show that the sequence is the original one, even though members are locally faulted out. The inference seems clear that a higher thrust sheet may have passed over the area at a level not far above the present surface. If this is so, no remnants of such a plate have been recognized during this survey. Work in the adjoining Frenchie Creek quadrangle may permit evaluation of the hypothesis here suggested.

There is no record in the nearby rocks by which the date of the shearing off of the Brock Canyon Formation may be determined. Economy of hypotheses, but no evidence, suggests that it occurred at the same time as the Lewis orogeny of the Mount Lewis area. Even so, the date may have been anywhere between Middle Triassic and Eocene.

FAULTING DURING THE VOLCANIC EPISODE

Data are not available as to the history of the area between the post-China Mountain(?) orogeny and the time of Tertiary volcanism. We do know there was deep erosion, enough to deroof the Granite Mountain pluton and leave a highly irregular topography at the time the volcanics began to accumulate.

Evidence of faulting unequivocally related to the volcanic episode is sparse. The arcuate faults around the several volcanic centers along which Tertiary rocks have been dropped into the pipes have been described and are conspicuous on the map of plate 1. The close association of the north-south faults between Indian Creek and Mount Lewis with the Mount Lewis plugs suggests that they may have formed at this time.

The suggestion has been made (Masursky, 1960) that the faults near Wilson Pass are part of a set bounding a "volcano-tectonic" depression to the south. Masursky suggested that such a depression formed during the volcanic episode and extended about 30 miles, from an area near Cortez to the Mount Moses quadrangle; in it accumulated extraordinary thicknesses of welded tuffs. Most of the area concerned has not been mapped.

If this is indeed a volcano-tectonic depression it seems unusual that the attitudes of the various blocks of volcanics seem quite indifferent to the bounding fault. The faults at Wilson Pass trend north of west but the volcanics to the south strike into, rather than parallel to, the faults. This relation seems to prevail in the Toiyabe Range also; to us it seems not to support the suggestion of a volcano-tectonic depression, especially in view of the very thick welded tuff far to the northeast of the mouth of Lewis Canyon which seems identical with that south of Wilson Pass.

BASIN RANGE FAULTS

SHOSHONE RANGE

Along much of the northwest side of the Shoshone Range in this area, the only evidence for Basin Range faults is physiographic. But in several places normal faults along the mountain front cut back into the range and from stratigraphic evidence can be shown to be normal faults.

At the mouth of Crum Canyon the fault bounding the range is visible and dips about 60° N.; sheared, semiconsolidated gravel clings to the hanging wall for a height of nealy 30 feet. Both east and west from here the scarp continues, but no rocks are preserved on the hanging wall above the general level of the frontal fans. West of Crum Canyon the fault scarp splits into several smaller scarps, some of which die out westward. Two miles west of Crum Canyon a fault diverges from the mountain front, runs south up the valley with volcanics to the west, crosses a low divide where it is exposed as dipping steeply west, and curves into conformity with the trend of the mountain front just east of Lewis Canyon. To the north of the curve in strike, this fault has dropped volcanics down to make the mountain front. As no volcanics are present east or southeast of the scarp for several miles, the throw of this fault is several thousand feet.

This fault splits near the bend. The northern split has Antler Peak Limestone in the hanging wall; the southern has quartzite of the Valmy. Again the absence of large blocks of these rocks in the immediately adjacent country of the footwall shows large displacement.

Near and west of the mouth of Lewis Canyon the faulting is highly complex. At least three strands are present, of which the southernmost cuts bedrock on both walls, offsets the Whisky Canyon fault and joins the frontal fault again only at the mouth of Rocky Canyon. The displacement of the next northerly fault is doubtless larger—it carries volcanics on the hanging wall. Several other strands, a mile or more out on the fans of the Reese River Valley, are indicated by scarps as much as 30 feet high.

According to Dr. V. P. Gianella (oral commun., 1950), who examined the workings of the now abandoned Betty O'Neal mine, the lowest adit, which started well out on the alluvial fans, penetrated at least three strands of the Basin Range fault before entering bedrock.

A split from the frontal fault passes south up the lower course of Pipe Canyon, crosses the divide into Horse Canyon, cuts off the volcanic vent at the lower bend of Horse Canyon, passes through a saddle on the Horse Canyon-Crippen Canyon divide (where it cuts off the Whisky Canyon fault) and thence turns abruptly west to the mountain front. This fault is readily traced over the topography. It dips about 50° or 60° W. As mentioned on page 69, the volcanic vent to the west is probably the displaced upper part of the Horse Canyon vent. The implied dip slip on the fault is nearly 10,000 feet.

South of the mouth of Horse Canyon, the mountain front is marked by a simple scarp to a point half a mile south of Crippen Canyon where a downdropped block of Valmy lies to the west of the big mass of Slaven Chert making up the ridge north of Trout Creek. The fault separating these blocks must have a throw of at least 2,000 feet. The western scarp, bounding the Valmy block, extends south of Trout Creek as the frontal fault of the mountain. Other scarps in the alluvium, as much as 2 miles to the west, testify to distributive faulting in this area.

South of Mill Creek several faults in the volcanics show relatively little offset, a few score feet at most. The main frontal fault must lie at the range front or under the fan gravel.

The only fault surely referable to the Basin Range system that lies wholly within the range is the Corral Canyon fault. This fault must die out under the alluvium of the east fork of Indian Creek or else be reduced to very minor displacement at the crossing of Indian Creek. From there northward its throw rapidly increases and becomes more than 3,000 feet on the two branches where they leave the north boundary of the Crescent Valley quadrangle. Both branches continue far to the northeast toward Beowawe.

The faults that bound the volcanic area of Cooks Creek may possibly belong also to the Basin Range system. Exposures are so poor that little can be learned of their details.

CORTEZ MOUNTAINS

Only a short segment of the magnificent frontal scarp of the Cortez Mountains lies within the Crescent Valley quadrangle. The frontal fault here is marked not only by its steep straight scarp but also by offsets of gravel at stream mouths, locally as much as 15 feet. The scarp of this range front is fully 2,500 feet high for several miles. As no bedrock whatever appears along the hanging wall, the downdrop along the fault is at least this great. As indicated by Mr. Plouff in his discussion of the gravity data collected in Crescent Valley, the actual movement has probably been several times as great.

If the cuesta slope of the Mal Pais continued beneath Crescent Valley at a constant dip of 5°, equal to that in the exposed slope, it would be at a few hundred feet below sea level at its junction with the frontal fault of the Cortez Mountains in the southeast corner of the Crescent Valley quadrangle. The basaltic andesite reappears on the crest of the Cortez Mountains just outside the limits of the Crescent Valley quadrangle, at an altitude of nearly 8,000 feet, where it has nearly the same attitude as on the Mal Pais, 15 miles away. This ridge is 3 miles from the frontal fault of the Cortez Mountains. The projected elevation of the Cortez Mountains cuesta is about 9,400 feet. On this simplified assumption the structural relief caused by the frontal fault of the Cortez Mountains would be about 10,000 feet.

Donald Plouff's analysis, p. 127–129, shows that this simple assumption should probably be modified, but the step faults mapped in the Cortez quadrangle to the south need only reduce the relief of the bedrock surface by less than a quarter in order to conform with his results.

AGE OF THE BASIN RANGE FAULTS

In some parts of the Basin Range province there is evidence of normal faulting of the Basin Range type having begun as early as Oligocene time (Stock and Bode, 1937; Nolan, 1943, p. 183). Here, however, we have no such record. Although comparable faults may of course be buried beneath the alluvial valleys, no Basin Range faults here can be identified as certainly older than the andesitic basalt of the Mal Pais. The age of this lava sequence is not known, it may be Miocene but more likely is Pliocene. Whether the faulting began immediately after the eruption or was delayed for some time, it has certainly been recurrent in Pleistocene time along both the Shoshone and Cortez range fronts and probably has occurred in Recent time at places along both.

GRAVITY SURVEY OF PART OF CRESCENT VALLEY

By Donald Plouff

A total of 47 gravity stations was established near Crescent Valley by Donald Plouff and S. W. Stewart on August 15 and 16, 1955. The observed gravity values for all stations were tied to a base station 1 mile west of Dean Ranch (fig. 40). The absolute value of observed gravity at this base station is 979,723.9 (mgal) (D. R. Mabey, written commun., 1957). Stations were established at 17 bench marks and 30 locations where altitudes were determined by surveying. Standard methods were used to calculate the Bouguer gravity anomaly for each station (Nettleton, 1940, p. 51–62).

A density of 2.67 g per cm^3 was assumed for calculation of the elevation and terrain corrections. Terrain corrections were carried to about 36 miles, using the U.S. Coast and Geodetic Survey system (Swick, 1942). The relative Bouguer anomaly (fig. 40) may be converted to the absolute Bouguer anomaly, corrected to sea level, by subtracting 217.2 (mgal) from the indicated values.

The negative gravity anomaly, centered along Crescent Valley, is associated with a thick section of Cenozoic volcanic and sedimentary rocks, whose average densities are less than those of the older rocks of the Shoshone and Cortez mountains. The axis of the gravity low probably corresponds to the thickest section of Cenozoic rocks. The closely spaced contour lines along the southeast side of the valley trend parallel to the Basin and Range fault along the northwest flank of the Cortez Mountains.

Interpretation of the thickness of the Cenozoic rock in Crescent Valley is limited by a lack of information on the densities of the Cenozoic rock within the valley, as contrasted to the densities of the older rock enclosing the valley. Interpretation also is limited by the influence of possible contrasts of density within the more dense rocks.

Gravity relief on a profile across Crescent Valley (fig. 41) is considered to be largely the result of changing thickness of the Cenozoic rock section. The maximum thickness probably occurs near the location of the most strongly negative gravity value. The range of rock densities described by Mabey (1960, p. 57–59) for rocks from the region near the Mojave Desert is presumed to be consistent with the densities of rocks

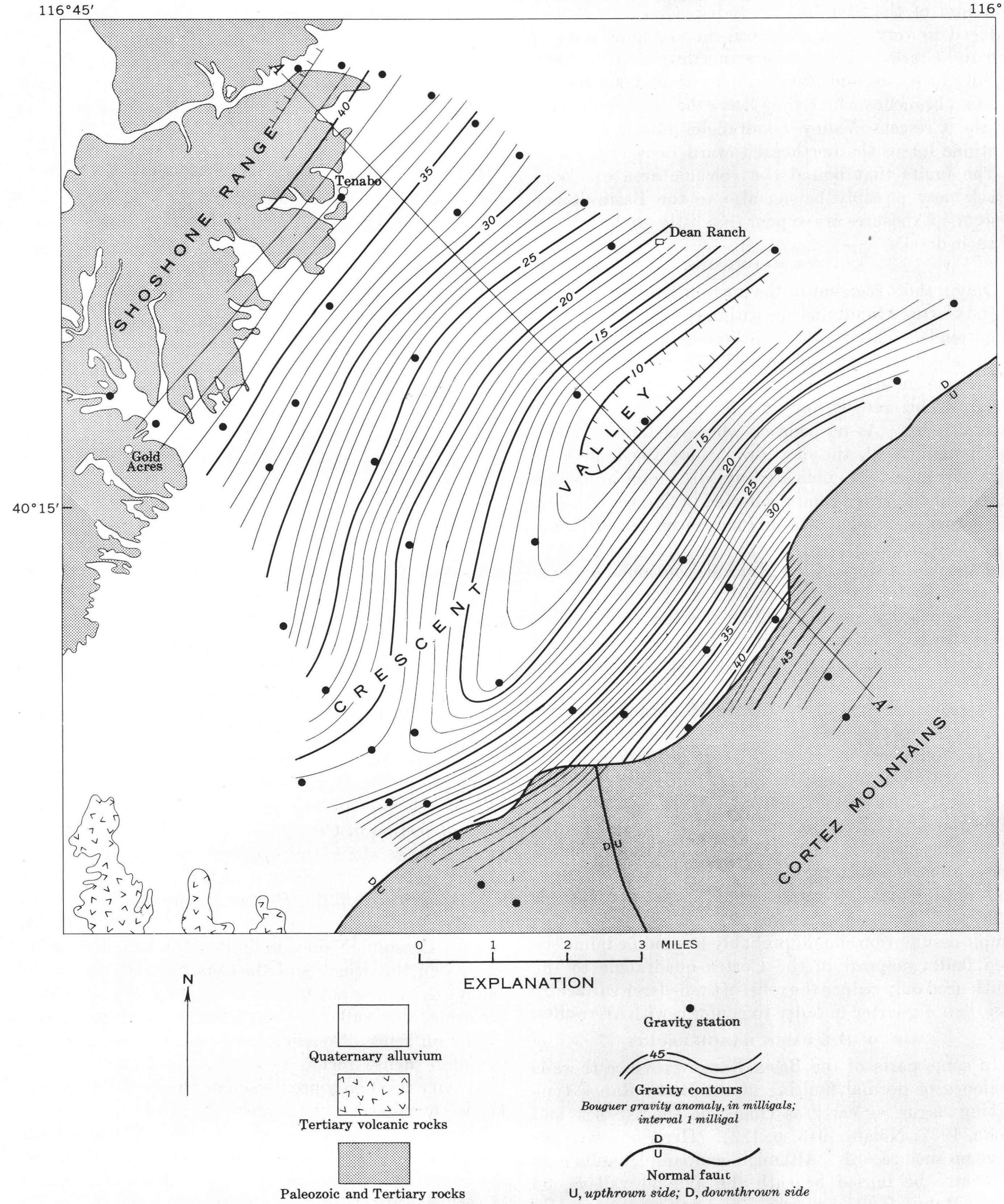

FIGURE 40.—Bouguer gravity anomaly map of part of Crescent Valley, Nev., by Donald Plouff.

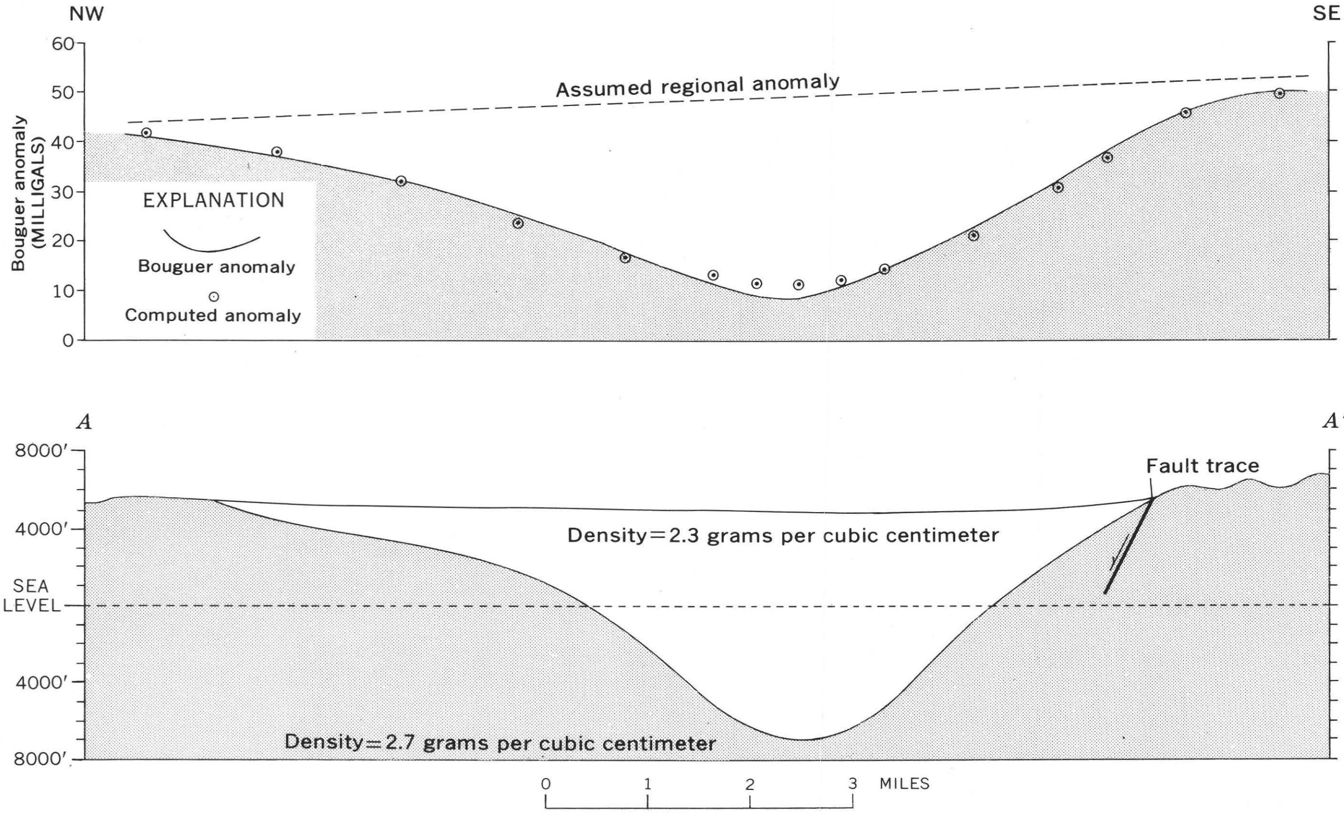

FIGURE 41.—Bouguer gravity anomaly and diagrammatic geologic profiles along line *A–A′* of figure 40.

in the vicinity of Crescent Valley. As in the western Mojave Desert, a constant density contrast of 0.4 g per cm^3 between the Cenozoic and older rocks is assumed.

A procedure of two-dimensional analysis (Dobrin, 1952, p. 96–99) was used to calculate a hypothetical geologic cross section for Crescent Valley, which fits the Bouguer gravity anomaly. The gravity contours are assumed to extend perpendicular to line *A–A′* (fig. 40) for an infinite distance. Calculation of the approximate thickness of the Cenozoic rocks is simplified by assuming that there are no changes of density within the Cenozoic rocks or within the bordering older rocks. One calculated anomaly, which nearly matches the observed Bouguer anomaly, requires a total thickness of 12,000 feet of Cenozoic rock (fig. 41). The dip of the interface between the Cenozoic and older rocks along the southeast side of Crescent Valley is about 40°.

STRUCTURE SUGGESTED BY GRAVITY DATA

The inflection of the interface between valley fill and bedrock in the diagrammatic geologic section (fig. 41) doubtless corresponds to distributive faulting along the Cortez Mountains front. Although only one fault is exposed at the surface here, distributive faults are very common, both along the Cortez front and along others in the region; it is very likely, from the gravity data, that an inactive fault lies buried beneath the valley fill northwest of the one exposed at the surface. The dip of the exposed fault is about 70°—much steeper than the dip of the interface suggested by the gravity data. To the southwest, in the Cortez quadrangle distributive faults are well exposed in the volcanic rocks on the trend of Crescent Valley (Gilluly and Masursky, 1965).

ECONOMIC GEOLOGY

By KEITH B. KETNER

A belt of intrusives and associated mesothermal silver, gold, copper, and lead deposits extends northwestward across the Shoshone Range (fig. 42). The mines of this belt comprise three districts: the Bullion district on the east slope, the Hilltop district near the crest of the range, and the Lewis district on the west slope. Bedded barite deposits, generally along a northeastward trend, appear to be unrelated to the metal deposits in distribution and origin.

HISTORY AND PRODUCTION

Following discovery of silver at Virginia City in 1859 and at Austin in 1862, a flurry of exploration activity resulted in discoveries of precious and base metals at Cortez in 1863, at Eureka in 1864, and at Battle Mountain in 1866. History does not record how, when, and

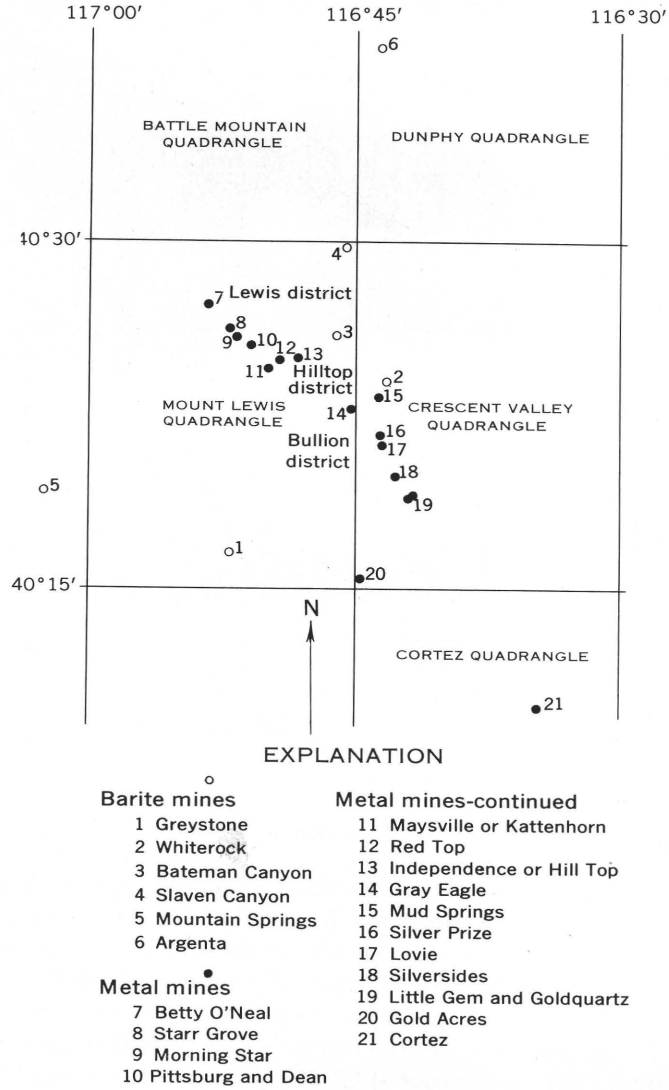

FIGURE 42.—Index map of Mount Lewis and Crescent Valley quadrangles and neighboring areas showing principal mines.

by whom the first discoveries were made in the Mount Lewis area. However, county records show production in the Bullion district beginning in 1869. In the summer of 1874 Jonathan Green and E. T. George found silver ore in the area which later became known as the Lewis district and production beginning in 1876 is recorded. The Hilltop district was organized about 1906.

The most valuable deposits in the area were discovered late. Although the Betty O'Neal mine was discovered in 1880, its richest ore bodies were not encountered until 1923, after which it far surpassed all other mines combined in silver output. Only after the discovery of gold at Tenabo in 1905 and at Hilltop in 1906 did gold and copper production achieve notable heights. The Gold Acres mine, whose production of gold has far exceeded that of the rest of the area, was opened in 1935.

A few mines account for most of the production in the area. The Betty O'Neal mine produced about 80 percent of the silver and nearly half the lead, and the Gold Acres mine produced nearly 90 percent of the gold. The Little Gem mine produced about 65 percent of the copper. Table 9 shows reported metal production for the years 1869 through 1957.

Although a large deposit of barite was discovered in the area in the early days of metal mining, barite did

TABLE 9.—*Recoverable metal in ore mined in the Bullion, Hilltop, and Lewis districts, 1869–1957, inclusive*

[Data from Couch and Carpenter, 1943, and U.S. Bur. Mines Minerals Yearbooks, annual volumes, 1943–57]

Year	Districts producing [1]	Gold (fine ounces)	Silver (fine ounces)	Copper (pounds)	Lead (pounds)	Total value
1869	B	----	----	----	----	$1,018
1871	B	----	----	----	----	4,486
1874	B	----	----	----	----	25,480
1876	L	----	----	----	----	47,709
1878	L	----	----	----	----	71,371
1879	L	----	----	----	----	140,320
1880	L	----	----	----	----	94,500
1881	B L	----	----	----	----	107,455
1882	L	----	----	----	----	79,430
1883	B	----	----	----	----	6,872
1887	L	----	----	----	----	9,444
1888	B L	----	----	----	----	48,955
1889	B	----	----	----	----	2,764
1890	L	----	----	----	----	16,632
1902	L	591	547	----	----	19,942
1903	L	855	420	----	----	17,895
1905	B	88	----	----	----	1,812
1906	B	484	31,930	295	76,506	35,806
1907	B	4,113	107,102	325,604	679,102	256,827
1908	B L	151	6,663	3,159	57,714	9,495
1909	B L H	646	3,081	2,023	18,930	16,032
1910	L H	214	231	28	188	4,545
1911	H	293	297	----	4,046	6,318
1912	B H	2,607	20,406	12,993	87,087	72,488
1913	B L H	2,897	35,242	1,010	972	81,395
1914	B L H	725	148,263	7,244	37,896	99,415
1915	B H	735	57,467	4,101	32,558	46,579
1916	B L H	1,800	123,326	7,112	74,353	125,269
1917	B L H	1,062	76,610	41,637	50,193	100,826
1918	B L H	1,420	85,853	257,681	1,234	180,600
1919	B L H	1,096	44,357	22,024	273,824	90,943
1920	B L H	132	27,991	1,466	16,165	34,804
1921	B L H	837	11,382	27,372	20,311	33,151
1922	B L	3	34,655	2,505	35,939	37,031
1923	B L H	340	666,702	54,270	291,914	582,124
1924	B L H	43	482,775	28,831	172,556	341,931
1925	B L H	192	607,137	77,167	287,950	461,335
1926	B L H	1,542	766,106	85,457	366,320	551,187
1927	B L	599	781,866	54,114	351,422	484,932
1928	B L H	402	571,722	38,580	237,568	362,111
1929	B L	63	202,902	25,142	87,391	119,372
1930	B L H	417	18,897	10,548	7,481	17,631
1931	B L H	383	1,261	3,725	1,927	8,688
1932	B H	300	425	1,900	715	6,463
1933	B L H	300	6,959	3,770	1,570	10,424
1934	B L H	1,193	62,980	52,148	15,337	87,194
1935	B L H	1,492	77,111	47,483	19,238	112,710
1936	B L H	3,940	61,085	81,863	55,075	195,376
1937	B L H	4,113	57,299	121,300	33,400	204,918
1938	B L H	7,032	84,838	29,400	64,100	306,044
1939	B L H	3,438	42,053	132,100	27,600	163,553
1940	B L H	4,023	66,946	305,300	51,900	227,505
1941	B L H	2,195	29,153	83,900	26,000	108,973
1942	B H	13,177	15,529	75,100	1,200	481,475
1943	B H	8,042	1,312	100	4,200	283,011
1944	B	9,358	1,058	6,000	5,500	328,281
1945	B H	9,490	2,745	----	----	335,417
1948	H	22	----	----	----	790
1950	H	22	----	----	----	794
1951	B	16,791	1,519	----	----	595,360
1953	B	18,400	1,408	----	100	645,287
1954	B	20,643	2,602	----	----	724,860
1955	B	18,983	1,970	----	----	666,188
1956	B	17,686	1,378	2,100	----	621,150
1957	B L H	15,162	1,287	1,900	100	533,330
Totals	----	200,532	5,434,848	2,038,452	3,577,582	11,496,023

[1] B, Bullion; H, Hilltop; L, Lewis.

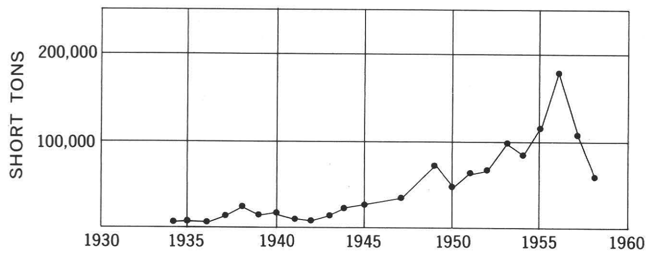

FIGURE 43.—Barite sold or used by Nevada producers, 1934–58.

not come into production until the early 1930's when increasing demands for barium products in the West made it profitable to mine. Probably owing to its proximity to a railroad, the Argenta mine just north of the Crescent Valley quadrangle began northeastern Nevada barite production in 1934. Bureau of Mines records show production of barite beginning in the Hilltop (Valley View & Slaven Canyon) mine in 1936 and in the Bateman Canyon and Greystone mines in 1953.

In 1956 Nevada ranked third among the states in barite sold or used by producers, with 178,440 short tons produced (fig. 43). Nearly all of this amount was produced from bedded barite deposits. Since production from bedded deposits began in the early 1930's Nevada's share of national production has attained about 14 percent (U.S. Bur. Mines, 1956, Minerals Yearbook).

METAL DEPOSITS

W. H. Emmons (1910, p. 113–126) visited the Mount Lewis area briefly in 1908. Even at the time of his visit the stopes of some old mines were inaccessible and we therefore know little about them, but Emmons had access to many mines then operating which are now inaccessible; for these his descriptions are the only ones available. The present discussion attempts to relate deposits to the regional geology and to record what is known of their mineralogy.

HOST ROCK

Metal deposits are confined largely to the Valmy and Slaven Formations and particularly to the chert facies of these formations. Shale, sandstone, and quartzite are less favored. Many deposits, including the Dean, Gray Eagle, Mud Spring, and Little Gem are partly in intrusive rocks. None are known in extrusive rocks.

Wallrock commonly is somewhat recrystallized, bleached, and sericitized to various distances from the lode. Emmons (1910, p. 124) reported sericitization in igneous rocks close to centers of metallization and chloritization farther away.

DISTRIBUTION

Metal deposits and intrusives mainly of quartz diorite and quartz monzonite are scattered in a northwest-southeast belt across the Shoshone Range. Aside from their apparent confinement to the upper plate of the Roberts thrust, the metal deposits of the Shoshone Range do not appear to be related to mappable structural features. In the Mount Lewis quadrangle the metal belt crosses the traces of faults and bedding in host rocks at an angle of about 30°. Possibly the distribution of intrusives and consequently of mineral deposits is controlled by major concealed structures in the lower plate of the Roberts thrust (Roberts, 1960).

STRUCTURE OF DEPOSITS

Judging by limited field observations and the brief descriptions of Emmons, most ore deposits in the metal belt are tabular lodes, sheeted zones, and fissure veins which do not seem to be at all consistent in attitude. Commonly the ore zones are 1 to 6 feet thick and have a short vertical range—probably no more than a thousand feet, commonly less than 500 feet. Apparently the ore was unevenly distributed in the lodes and veins; stopes in the Lovie, Silver Prize, and Silversides deposits are sporadically distributed along the lode, intervening parts of the lode are completely barren. Emmons (1910, p. 116) reported small postore faults. Textural relations of ore minerals indicate some fracturing between deposition of primary and secondary minerals.

WEATHERING

Weathering of the exposed parts of the deposits has converted most of the primary minerals to oxides, carbonates, sulfates, and chlorides to a depth of 50 to 100 feet.

LODE MINERALS

Available specimens of ore are now so few as to make a thorough study of the ore minerals impossible. The following minerals were mentioned by Emmons or were observed during the present study in thin and polished sections of specimens obtained from pillars, dumps, and ore hoppers. Mrs. Kattenhorn of Battle Mountain kindly contributed samples, otherwise unobtainable, of rich tetrahedrite ore from the Kattenhorn mine, and Prof. L. S. McGirk of the University of Nevada loaned ore specimens from the Mackay School of Mines Museum.

Gold, Au. Native gold was observed in quartz and iron oxide of the Red Top and Goldquartz mines (Emmons, 1910, p. 118).

Silver, Ag. Native wire silver in the Betty O'Neal mine was reported by Burchard (1882, p. 123).

Copper, Cu. A little native copper was reported by Emmons (1910, p. 123) in the Morning Star and Pittsburg mines.

Stibnite, Sb_2S_3. Antimony sulfide is common in the Gray Eagle, Betty O'Neal, and Kattenhorn mines. It is said to contain considerable silver but analysis of one sample (table 10) does not bear this out. Texture of ore in the Kattenhorn mine indicates stibnite was shattered after deposition and that the voids between grains were filled with marcasite.

Molybdenite, MoS_2. Molybdenite was reported by Emmons (1910, p. 116) from the Tenabo area.

Galena, PbS. Galena, one of the commonest ore minerals, contains a fraction of a percent of silver (table 10). It alters to cerussite and anglesite along cleavage fractures. Emmons reports rich silver-bearing galena at the Silver Prize mine.

Chalcocite, Cu_2S. Small amounts of the secondary mineral, chalcocite, were observed in ores of the Hilltop district. Emmons reported chalcocite in the Tenabo area.

Sphalerite, ZnS. Both brown and black zinc sulfide are common ore minerals. Sphalerite at the Dean mine contains many chalcopyrite inclusions so small as to be barely distinguishable under the microscope. Sphalerite at the Silver Prize mine is veined with marcasite. Grains of light and dark sphalerite from several mines were analyzed spectrographically with the results shown in table 10. Apparently sphalerite contains appreciable amounts of silver and cadmium.

Pyrite, FeS_2. Iron sulfide, one of the commonest minerals in the mines is generally in cubes and commonly contains minute inclusions of chalcopyrite, sphalerite, and galena. Usually it is closely associated with arsenopyrite and in places forms inclusions in it. It is one of the earliest minerals to form.

Marcasite, FeS_2. Marcasite is secondary and fills fractures in earlier formed minerals. It is a common associate of stibnite in the Kattenhorn mine.

Arsenopyrite, FeAsS. Arsenopyrite is commonly associated with pyrite. It appears from textural relations to be one of the earliest minerals deposited.

Boulangerite, $5PbS{\cdot}2Sb_2S_3$. Boulangerite, carrying a small amount of silver (table 10), is present in the Mud Spring area as small alined blades in large crystals of calcite. The carbonate-boulangerite aggregates appear to replace pyrite and galena. Identification of this unusual mineral was confirmed by X-ray (A. J. Gude 3d) and spectroscopic analyses (Nancy M. Conklin).

Bournonite, $2PbS{\cdot}Cu_2S{\cdot}Sb_2S_3$. This lead-copper sulfantimonide, sometimes called cog-wheel ore, is associated with tetrahedrite and galena in the Betty O'Neal mine.

Ruby silver. A small amount of an unidentified ruby silver mineral, possibly pyrargyrite, was seen in specimens from the Kattenhorn and Betty O'Neal mines. Ruby silver minerals are generally secondary. Emmons (1910, p. 123–126) reported small amounts in the Morning Star, Pittsburg, and Starr Grove mines. Burchard (1882, p. 123) reported large amounts with wire silver in the Betty O'Neal mine.

Tetrahedrite, $3Cu_2S{\cdot}Sb_2S_3$. Tetrahedrite is associated with galena. It is probably the main silver-bearing mineral (table 10) although standard microchemical tests fail to reveal silver in it. Tetrahedrite high in silver is commonly known as freibergite. Silver probably substitutes for copper in the silver-bearing variety. Tetrahedrite was seen in samples from the Betty O'Neal, Dean, Kattenhorn, Gray Eagle, Silversides, and Mud Spring mines and the Hilltop district.

Cerargyrite, AgCl. According to Emmons and old-timers of the area, cerargyrite or "horn silver" was a mainstay of early silver production. Microchemical tests of fine-grained aggregates of secondary minerals showed a little silver chloride, but none was seen microscopically.

Quartz, SiO_2. Quartz is the commonest gangue mineral.

Oxidation products of stibnite. Stibnite is commonly partly altered to antimony oxides which could not be positively identified.

Iron oxides. Iron oxides replace pyrite, commonly without altering the original crystal form.

Cassiterite, SnO_2. Cassiterite, including some of the banded variety called wood tin, was found in ore of the Independence mine in the Hilltop district.

Calcite, $CaCO_3$. Calcite is a common gangue mineral.

Copper carbonates were reported by Emmons (1910, p. 116–118) in the oxidized zone of Little Gem and Lovie mines.

Cerussite, $PbCO_3$. Lead carbonate replaces galena along cleavage fractures in ore near the surface. It is especially common at the Silver Prize mine.

Muscovite, $KAl_2(AlSi_3)O_{10}(OH)_2$. The fine-grained variety of this mineral, sericite, is present both in ore and altered wallrock.

Chlorite (hydrous magnesium-aluminum-iron silicate). Chlorite is a common alteration product of igneous rock distant from the centers of mineralization.

Barite, $BaSO_4$. Barite is an uncommon gangue mineral. Emmons (1910, p. 126) reported quartz and sulfides of the Starr Grove mine to be in part deposited later than barite.

Anglesite, $PbSO_4$. Anglesite replaces galena along fractures in near-surface ore.

The textural relations of ores are sufficiently clear in a few specimens to establish the following sequence of deposition:

1. Quartz, arsenopyrite, pyrite.

2. Galena, sphalerite, chalcopyrite, tetrahedrite(?).
3. Tetrahedrite, stibnite, boulangerite, calcite, covellite, cerussite, anglesite.

CLASSIFICATION AND AGE

The most common primary hydrothermal minerals of the Lewis, Hilltop, and Bullion districts are quartz, arsenopyrite, pyrite, galena, sphalerite, chalcopyrite, and tetrahedrite. Specimens from Cortez and descriptions by Roberts (1951) and Emmons (1910, p. 100) indicate that the mineral assemblages of Battle Mountain, Cortez, and Mill Canyon districts, on the extensions of the metal belt shown in figure 42, resemble those of the Lewis, Hilltop, and Bullion districts.

This assemblage of primary minerals places the metal deposits in the mesothermal category of Lindgren, (1933, p. 565), implying that they were deposited at moderate depths from ascending solutions of moderate temperature.

Mesothermal silver-lead veins were grouped into five classes by Lindgren on the basis of the chief ore and gangue minerals:

1. Galena, tetrahedrite, quartz.
2. Galena, tetrahedrite, siderite.
3. Galena, siderite.
4. Galena, sphalerite, siderite, calcite, barite.
5. Galena, pyrite, quartz.

Deposits of the Lewis, Hilltop, and Bullion districts resemble No. 1 in the presence of tetrahedrite and No. 5 in the prevalence of pyrite.

Some of the Tertiary rocks, but none of the uppermost Tertiary rocks contain metal deposits, so the metalization probably preceded late Pliocene time.

GUIDES FOR FURTHER EXPLORATION

Because the area has been thoroughly prospected it is unlikely that any more large metal deposits will be found at the surface, though it seems probable that some remain concealed below. Intrusives or solutions from which the known mineral deposits in the upper plate originated must have penetrated the Roberts thrust. The disturbed zone along the thrust is likely to be favorable for ore deposition because of its porosity and permeability. The Gold Acres mine where the ore is in brecciated rocks along the Roberts thrust illustrates the point. Intersections of lodes or intrusive contacts with the Roberts thrust would therefore seem to be logical places to drill for ore. The Roberts thrust is closest to the surface near the borders of windows.

In addition, drilling beneath the valley fill in the Reese River Valley and Crescent Valley along the northwest-southeast trend of known metal deposits in the Shoshone Range is likely to reveal additional deposits. Because the richest mines of the area, the

TABLE 10.—*Semiquantitative spectrographic analyses, in percent, of selected ore minerals*

[Analyst, Nancy M. Conklin]

Mineral and source	Si	Al	Fe	Mg	Ca	Na	Ti	Mn	Ag	As	Ba	Bi	Cd	Co
Tetrahedrite composite:														
Betty O'Neal, Dean, Kattenhorn mines	0.7	0.015	3.0	0.015	0.15	<1.0	<0.005	0.7	>10.0	3.0	0.0007	0.007	0.07	0
Boulangerite:														
Mud Spring area	3.0	.15	1.5	3.0	>10.0	0	.0015	1.5	.03	.7	.0007	.003	.015	0
Stibnite:														
Kattenhorn mine	.7	.07	1.5	.003	.03	0	.003	.0015	.015	.15	.0015	0	0	0
Galena:														
Betty O'Neal mine	.3	.003	.03	.03	.3	0	0	.03	.07	0	.0003	.03	.007	0
Hilltop district	.3	.07	.07	.0015	.03	0	0	.0015	.15	.15	.015	.03	.015	0
Gray Eagle mine	.3	.07	.3	.007	.03	0	0	.003	.7	.7	.0003	.07	.015	0
Silver Prize mine	.3	.015	.07	.007	.03	0	.0015	.003	.7	0	.0003	0	.015	0
Sphalerite (brown and black):														
Betty O'Neal mine	.3	.003	1.5	.007	.15	<1.0	0	1.5	.3	.3	.0003	0	1.5	0
Sphalerite (brown and black) composite:														
Gray Eagle and Mud Spring mines	1.5	.007	7.0	.007	.15	<1.	0	.7	.15	.7	.0003	.0015	1.5	0
Sphalerite (black):														
Hilltop district	.7	.07	7.0	<.0015	.03	<1.	.0015	.7	.007	1.5	.007	.0015	1.5	.0015
Silver Prize mine	.3	.0015	7.0	<.0015	.015	<1.	0	.7	.03	0	.0003	0	1.5	.003

Mineral and source	Cu	Ga	In	Mo	Ni	Pb	Sb	Sn	Sr	Te	V	Zn	Zr
Tetrahedrite composite:													
Betty O'Neal, Dean, Kattenhorn mines	>10.0	0	0	0	0.0007	0.3	>10.	0.07	0	1.5	0	7.0	<0.005
Boulangerite:													
Mud Spring area	.07	0	0	0	0	>10.	>10.	.0015	.007	0	Tr.	.7	<.005
Stibnite:													
Kattenhorn mine	.03	0	0	0	.003	0	>10.	.0015	0	0	0	.07	0
Galena:													
Betty O'Neal mine	.07	0	0	0	0	>10.	.7	.003	0	0	0	.15	0
Hilltop district	.3	0	0	0	0	>10.	.3	.03	Tr.	0	0	.07	0
Gray Eagle mine	.07	0	0	0	0	>10.	.7	.0007	0	0	0	.03	0
Silver Prize mine	.03	0	0	0	0	>10.	.7	.007	.0015	0	0	.03	.0015
Sphalerite (brown and black):													
Betty O'Neal mine	1.5	.0015	0	0	.0007	.3	.3	.03	0	0	0	>10.	<.005
Sphalerite (brown and black) composite:													
Gray Eagle and Mud Spring mines	1.5	.0003	<.007	0	0	1.5	.3	.03	0	0	0	>10.	0
Sphalerite (black):													
Hilltop district	.7	0	.007	<.001	0	.7	.07	.15	.0015	0	0	>10.	0
Silver Prize mine	.3	Tr.	<.007	<.001	.0007	.15	.07	.003	0	0	0	>10.	0

Looked for but not found: Au, B, Ce, Cr, Dv, Er, Eu, Gd, Ge, Hf, Hg, Ho, Ir, La, Li, Lu, Nb, Nd, Os, P, Pd, Pr, Pt, Re, Rh, Ru, Sc, Sm, Ta, Tb, Te, Th, Tl, Tm, U, W, Y, Yb.

Betty O'Neal, Little Gem, and Gold Acres, are near the edges of the range, this course is especially attractive. The valley fill is thinner on the southeast than on the northwest side of the Shoshone Range; indeed the fill may be prohibitively deep for some miles into the Reese River Valley.

GOLD ACRES MINE

The Gold Acres deposit had not been discovered at the time of Emmons' visit and is therefore described here. The deposit was first exploited in 1935 by underground workings but open-pit methods were soon introduced. Figure 44 is a map of the western wall of the open pit where mining was in progress at the time of the investigation, 1957. The mine is at the edge of a window in the upper plate of the Roberts thrust. The floor of the pit is composed of carbonate rocks of the eastern facies. The lower half of the wall is a breccia zone composed chiefly of siliceous sedimentary rock with less carbonate rock. The upper half of the wall is composed of a sill-like intrusive of highly altered rock, probably quartz latite, overlain by Silurian and Devonian siltstone and chert belonging to the western facies.

Although both top and bottom of the ore are assay walls, the base is virtually the upper limit of undisturbed eastern facies carbonate rocks. The top lies within the breccia zone. The breccia zone and overlying sill dip gently to the west. Consequently the overburden becomes thicker westward.

The breccia zone, particularly the siliceous rocks, and the intrusive are mineralized. Small amounts of iron oxides, some apparently pseudomorphous after pyrite, are disseminated and also concentrated in veinlets in siliceous rocks of the breccia zone. Pyrrhotite is present in the intrusive. The tenor of gold ranges from about 0.06 to 0.30 oz per ton. No gold was visible microscopically in polished sections made from heavy concentrates and doubtless it is submicroscopic.

The breccia zone in which the gold is contained was obviously made by the Roberts thrust and therefore dates from Early Mississippian time or later. The intrusive resembles Tertiary rocks of the area. Probably the breccia zone both localized the intrusive and provided channels for the passage of gold-bearing solutions emanating from it.

BARITE DEPOSITS

In the United States, barite is obtained mainly from three distinct types of deposits: (1) fracture fillings and replacements in carbonate rocks, (2) residual mantles on barite-rich carbonate rocks, and (3) beds in siliceous sedimentary rocks. Vein and mantle deposits are well known. The purpose of this study is to record some of the important features of the less well known bedded deposits.

In the Mount Lewis and neighboring quadrangles are several barite deposits of the bedded type which have been exploited at irregular intervals since about 1930. In the regular course of mapping these deposits were examined and three in the Mount Lewis quadrangle were mapped at a large scale (figs. 45–47). Numerous specimens of barite and host rocks were studied in the laboratory by means of thin sections, polished sections, and spectroscopic and chemical analyses. The evidence of mapping and laboratory studies indicates that the bedded deposits are replacements of bedded chert and limestone.

V. P. Gianella (1941, p. 294) of the Nevada Bureau of Mines reported briefly on the barite deposits of northern Nevada including some of those described here, and D. A. Brobst (1958, p. 96, 112) of the U.S. Geological Survey recently estimated reserves of the region.

HOST ROCK

The rock which encloses bedded barite deposits in the Mount Lewis quadrangle is the Slaven Chert of Devonian age, composed of thin-bedded, dark, radiolarian chert with less common interbedded argillaceous chert beds and limestone lenses. Although the following description is based on rocks surrounding the barite deposits of Mount Lewis quadrangle, reconnaissance indicates that the lithology is representative of the host rocks of bedded barite deposits of the region generally.

Composition.—Host-rock chert is composed largely of quartz and chalcedony with minor amounts of pyrite in various stages of oxidation, clay, carbonaceous matter, a mineral resembling sericite, and traces of other minerals too small to be identified.

Chemical and spectrographic analyses reveal that the chert usually contains more than 90 percent SiO_2 and small amounts of calcium, barium, aluminum, and iron. Additional elements present in very small amounts are shown in table 11.

Argillaceous chert such as No. 19 in table 11 differs chemically from ordinary chert principally in containing more aluminum and potassium and consequently less silicon.

Limestone lenses in the chert consist almost entirely of calcite, quartz, and chalcedony. Chemical and spectrographic analyses of limestone from the Slaven Canyon deposit indicate an assemblage of minor chemical constituents similar to that of the chert except that limestone contains consistently more manganese and strontium but less boron (table 11). The magnesium content of limestone is less than 0.1 percent.

Texture.—In chert, grains of quartz and chalcedony large enough to be seen clearly under the microscope have interpenetrating mutual contacts. The grain size

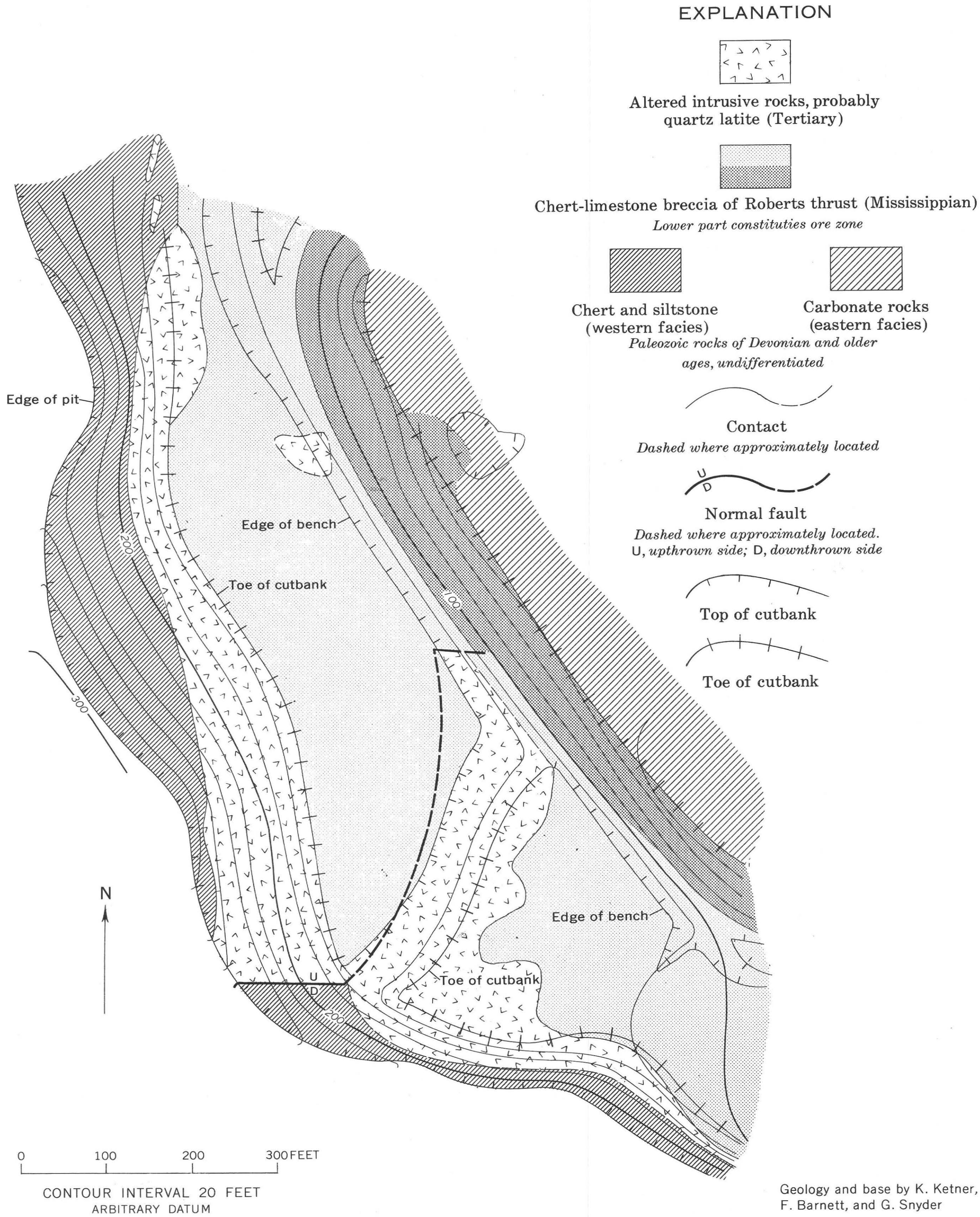

FIGURE 44.—Sketch map of part of the Gold Acres mine, unsurveyed sec. 36, T. 28 N., R. 46 E.

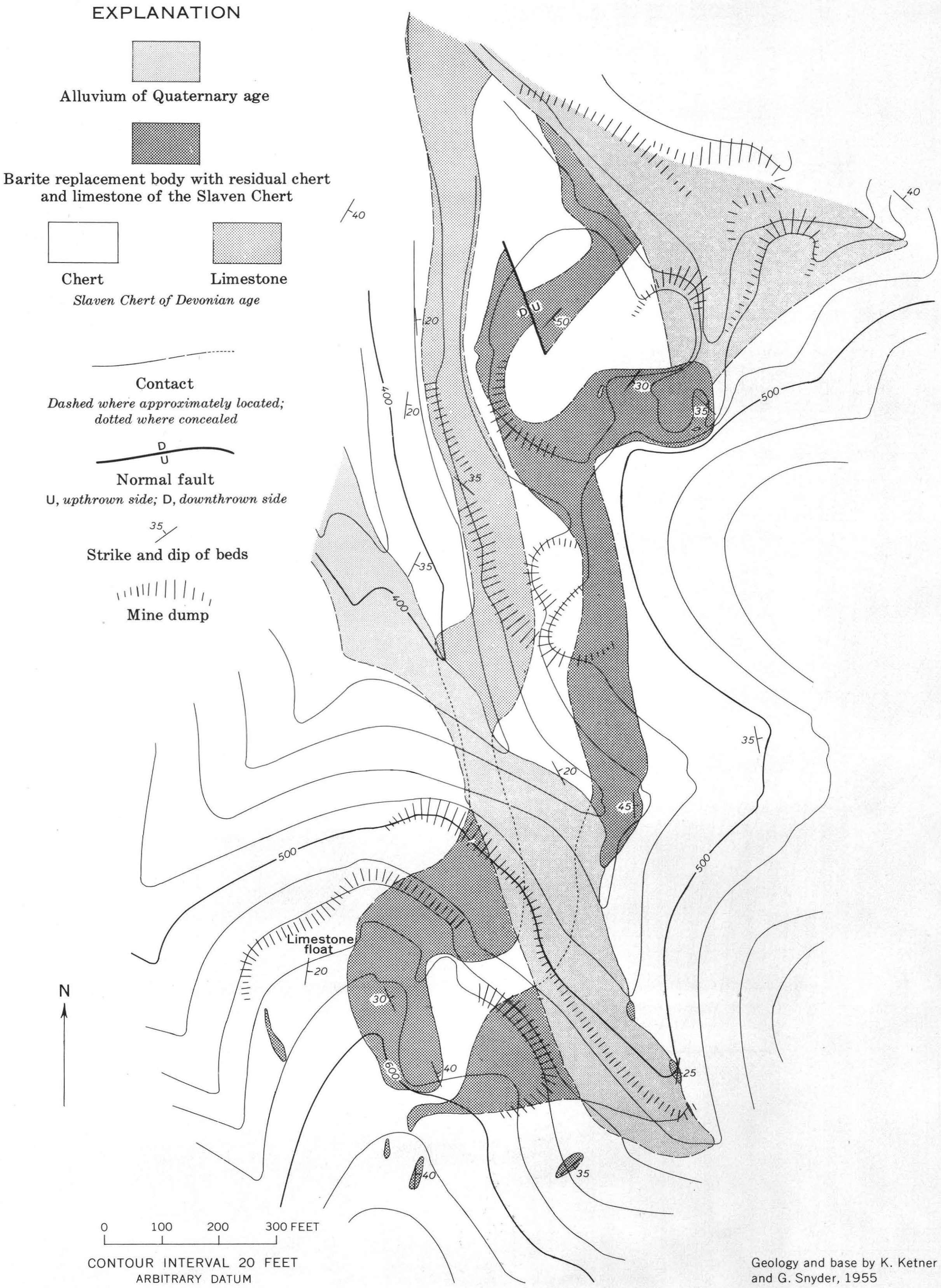

FIGURE 45.—Map of the Hilltop (Valley View) barite deposit, sec. 2, T. 30 N., R. 46 E.

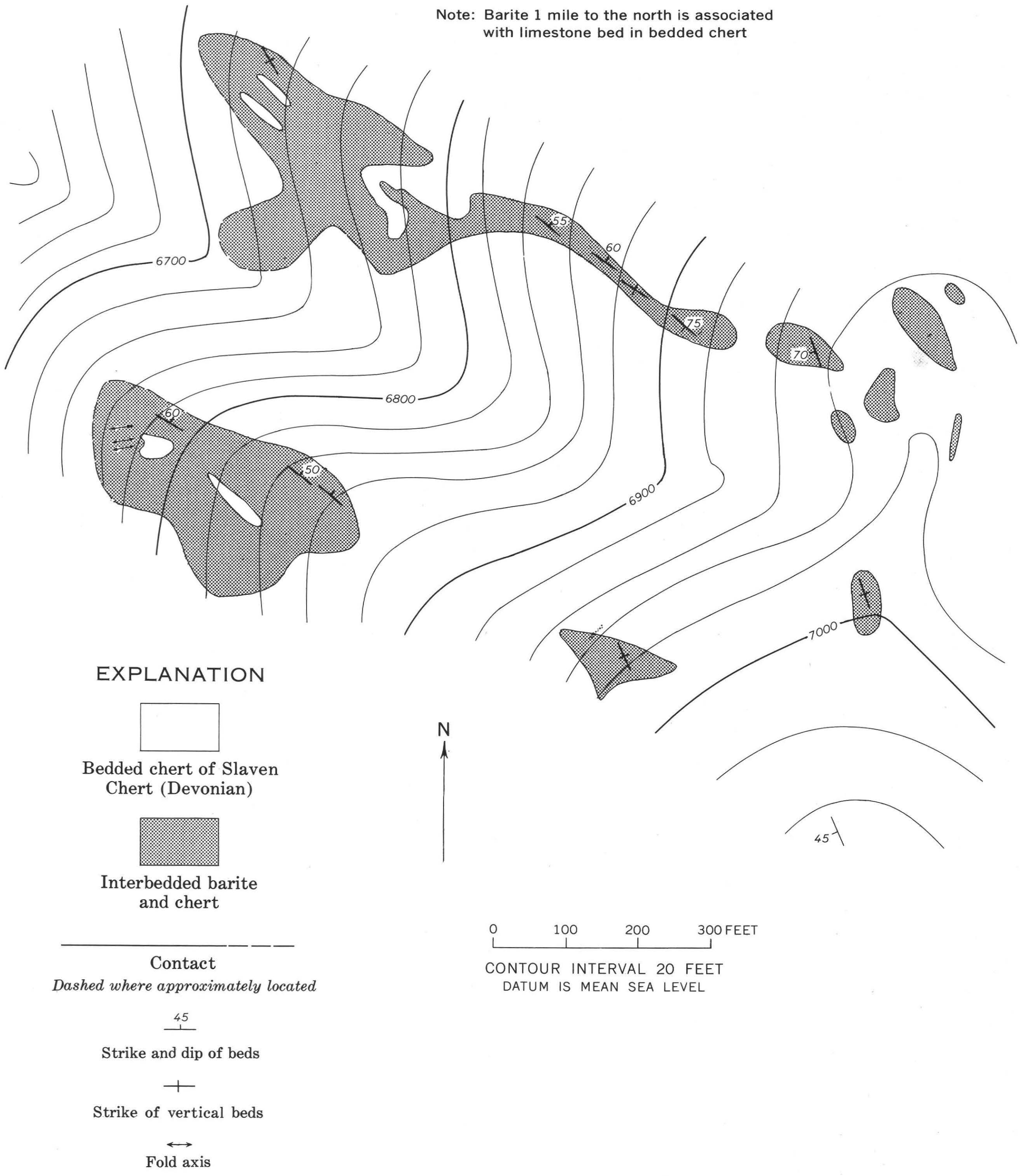

FIGURE 46.—Map of Bateman Canyon barite deposit, NW¼ sec. 35, T. 30 N., R. 46 E.

ranges from about 0.002 to about 0.5 mm in diameter. The coarser grains have commonly resulted from recrystallization as most of them are concentrated near fractures rather than along bedding surfaces. Some large grains, however, are clearly clastic particles, distinguishable from the rest by their smooth outlines and concentration in beds.

Shaly laminae at 1- to 3-inch intervals permit the chert to break up into platy fragments. In many beds variations in grain size, concentration of organic matter, clastic particles, or radiolaria produce nonfissile bedding on the order of a millimeter thick. Stylolites nearly parallel to bedding are common.

Limestone lenses in the chert are much more coarsely

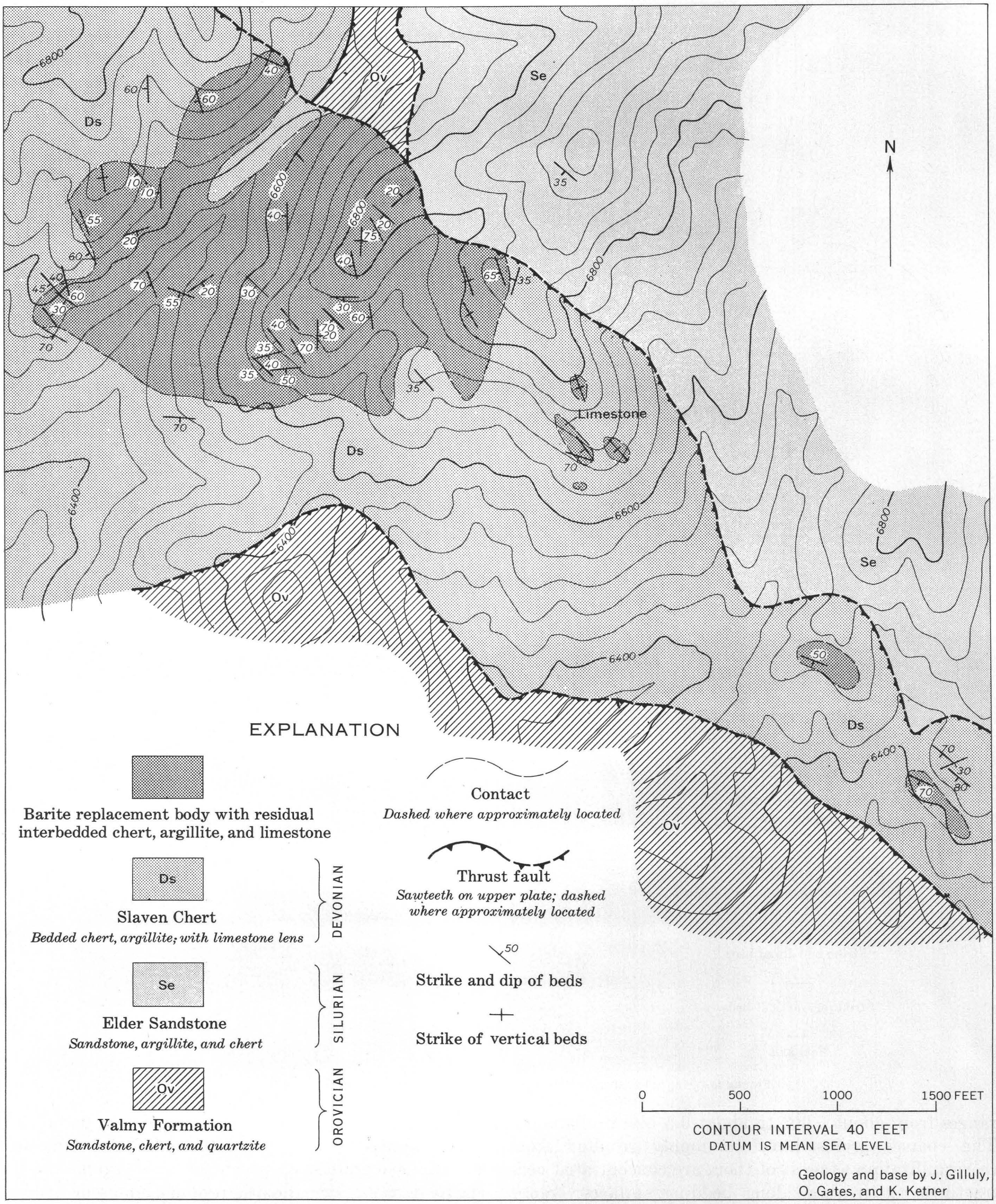

FIGURE 47.—Map of the Greystone barite deposit, unsurveyed secs. 23, 24, 25, 26, T. 28 N., R. 45 E.

TABLE 11.—*Analyses, chemical and semiquantitative spectrographic, of host rocks and ore at the Hilltop (Valley View), Bateman Canyon, and Greystone barite mines*

Sample	Si	Ca	Ba	Al	Ag	B	Co	Cr	Cu	Fe	Ga	K	Mg
Hilltop Canyon mine													
Bedded chert outside mineralized area:													0.03
307	[1]44.2	0.15	0.15	0.7	0	0.003	0	0.0003	0.003	0.7	0	0	.015
313	[1]45.1	.07	.03	.3	Tr.	.007	0	.0007	.007	.15	0	0	.03
Bedded chert within mineralized area:													
32	[1]44.3	.15	[1].4	.7	0	0	.003	.0003	.007	.3	0	0	.03
30	[1]44.2	.3	.3	.7	0	.003	0	.0003	.003	.3	0	0	.03
Limestone within mineralized area:													
44	[1]2.1	[1]36.8	.3	.0015	0	0	0	Tr.	.0003	.07	0	0	.03
52	[1]2.9	[1]36.4	.07	.15	0	0	0	Tr.	.0007	.15	0	0	.07
Barite:													
25	.7	.03	[1]57.3	.07	0	0	0	Tr.	.0007	.15	0	0	<.001
27	[1]1.8	.07	[1]55.0	.15	0	0	0	Tr.	.0015	.07	0	0	.003
Bateman Canyon mine													
Bedded chert outside mineralized area:													
305	[1]42.4	0.3	0.07	1.5	0	0.007	0	0.0007	0.03	0.7	Tr.	0.7	0.07
12	[1]43.0	.7	1.5	.7	Tr.	.003	0	.0015	.003	.3	0	.7	.03
Bedded chert within mineralized area:													
19	[1]41.3	.03	.7	3.0	Tr.	.015	0	.003	.0015	.15	.00015	1.5	.15
15	[1]42.3	.15	1.5	1.5	0	.003	.0003	.0007	.003	.3	0	.7	.03
Barite:													
8	0	.015	[1]58.6	.015	0	0	0	0	.00015	.003	0	0	.001
16	[1]1.8	.015	[1]54.2	.15	0	0	0	Tr.	.0007	.15	0	0	<.001
Greystone mine													
Bedded chert within mineralized area:													
3	[1]43.0	0.15	1.5	0.7	0	0.007	0	0.0007	0.0015	0.15	0	0	0.03
5	[1]39.4	.07	3.8	1.5	0	.015	0	.0015	.007	.3	Tr.	.7	.07
Barite:													
1	[1]2.0	.15	[1]55.0	.15	0	0	0	0	.0007	.07	0	0	.003
4	[1]5.6	.07	[1]48.9	.15	0	0	0	.0003	.003	.15	0	0	.007

Sample	Mn	Mo	Na	Ni	Pb	Sc	Sr	Ti	V	Y	Yb	Zr
Hilltop Canyon mine												
Bedded chert outside mineralized area:												
307	0.03	0	<0.05	0.0015	0	0	0.003	0.07	0.0007	----------	----------	Tr.
313	.007	0	.07	.0015	0	0	0	.007	.0015	----------	----------	0.0015
Bedded chert within mineralized area:												
32	.007	0	<.05	.0007	0	0	.0015	.015	.0015	----------	----------	.0015
30	.015	0	<.05	.0015	0	0	.0015	.03	.0015	----------	----------	.007
Limestone within mineralized area:												
44	.15	0	<.05	.0003	0	0	.03	.03	.0015	----------	----------	Tr.
52	.3	0	<.05	0	0	0	.03	.03	.0015	----------	----------	Tr.
Barite:												
25	.0015	0	.07	0	0	0	.015	.07	0	----------	----------	0
27	.0015	0	.07	0	0	0	.03	.07	Tr.	----------	----------	0
Bateman Canyon mine												
Bedded chert outside mineralized area:												
305	0.007	0	<0.05	0.0007	0	0.0007	0.0015	0.07	0.003	0.0015	0.0003	0.003
12	.007	Tr.	.07	.0003	0	0	.003	.03	.003	.0015	.0003	.0015
Bedded chert within mineralized area:												
19	.007	0	<.05	0	.007	.0007	.003	.07	.015	0	0	.007
15	.03	0	<.05	.003	.0015	0	.003	.015	.003	0	0	.0015
Barite:												
8	Tr.	0	.07	0	0	0	.15	.03	0	0	0	0
16	.007	0	<.05	Tr.	0	0	.07	0	.0007	0	0	0
Greystone mine												
Bedded chert within mineralized area:												
3	0.003	0	<0.05	Tr.	----------	0	0.007	0.007	0.003	----------	----------	0.0015
5	.007	0	.07	0.0003	----------	.0007	.015	.007	.003	----------	----------	.0015
Barite:												
1	Tr.	0	.07	0	----------	0	.15	.007	Tr.	----------	----------	0
4	.0015	0	.07	0	----------	.0007	.15	.0007	.0015	----------	----------	Tr.

[1] Wet method. Chemical analysis by Vertie C. Smith. All other figures were determined spectrographically by semiquantitative methods by Paul R. Barnett.

Looked for but not found: As, An, Be, Bi, Cd, Ce, Dy, Er, Em, Gd, Ge, Hf, Hg, Ho, In, Ir, La, Li, Ln, Nb, Nd, Os, P, Pd, Pr, Pt, Re, Rh, Rn, Sb, Sn, Sm, Ta, Tb, Te, Th, Tl, Tm, V, W, Zn.

736-006 O—65——10

crystalline than neighboring chert. Grains are commonly 1 to 4 mm in diameter. The coarse grain size appears to be due to recrystallization because many microfossils replaced by calcite are enclosed entirely within crystallographically continuous calcite crystals. Silica forms corroded remnants of lenses or beds of sedimentary chert and clusters of quartz and chalcedony grains which replace limestone.

The bedding is generally poorer than in chert or barite, being discontinuous and rather coarse (fig. 48). Radiolaria are less common in limestone than in chert. Possibly most of those originally present did not survive recrystallization.

Color.—Characteristically the enclosing chert is dark gray owing to the disseminated intergranular carbonaceous matter. Lighter shades of gray appear to result from recrystallization which, by increasing the grain size, diminishes the number of discrete particles of carbonaceous matter. Less commonly the chert is greenish gray. Chert beds having a greenish cast contain unusually high concentrations of a micaceous mineral, probably sericite.

Limestone lenses are dark gray on freshly broken surfaces but brown on weathered surfaces.

Alteration.—No noteworthy mineralogic, textural, or chemical differences were detected between chert interbedded with barite and other cherts at various distances up to 2 miles from known barite deposits.

Pyrite is commonly partly or wholly oxidized. Some fractures contain calcite or barite; most are filled with quartz. In veinlets containing both quartz and barite, the barite is in the center of the vein and presumably was introduced last. Both chert and limestone have been partly recrystallized but apparently not during the barite mineralization because recrystallized material is not concentrated near barite deposits.

Age.—The Greystone, Bateman Canyon, and Hilltop or Valley View deposits are in the Slaven Chert of Devonian age. The brachiopod *Halorella* is present in the limestone and barite of the Greystone mine and in barite of the Mound Springs area in the Mount Moses quadrangle. The attitude of beds between the Valley View and Bateman Canyon mines suggests that these two deposits are at about the same horizon within the Slaven Chert. The Mound Springs deposit is in a formation mapped originally as the Pumpernickel Formation of Carboniferous age (Ferguson, Muller, and Roberts, 1951b). Recent discoveries of *Halorella* (see p. 40–41) show the host rocks there to be of Devonian age also. The Argenta deposit and others farther northeast are in chert formations similar to the Slaven in lithology but of uncertain age. Small deposits are present also in bedded chert of Ordovician age, but as yet none of these have been commercially exploited in this area.

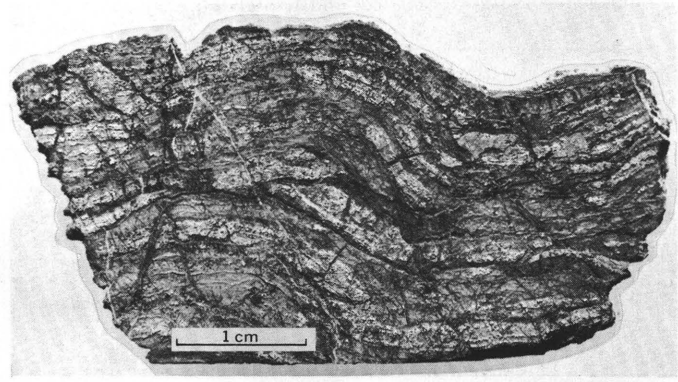

FIGURE 48.—Photograph of polished and etched slab of limestone of the Slaven Chert, showing chert beds (light gray) in relief. Greystone deposit.

DISTRIBUTION

Known bedded barite deposits are scattered from near the southwest corner of the Mount Lewis quadrangle and the south edge of the Cortez quadrangle northeastward for at least 80 miles through parts of Lander, Eureka, and Elko Counties. Like the metal deposits, barite deposits do not appear to be directly related to mappable faults and folds. Unlike the metal deposits, the barite deposits do not seem to be related to the distribution of exposed intrusive rocks in the Mount Lewis quadrangle. They seem more closely related to bedded chert than to intrusive rocks.

SIZE AND SHAPE

Individual deposits range from a few feet to about 2,000 feet in longest exposed dimension.

The outcrops are generally somewhat elongate parallel to host-rock bedding. Commonly, deposits consist of a large irregular body associated with smaller pods at distances of a few hundred feet or less.

Premineral and postmineral faulting and folding and irregular replacement of beds make the boundaries irregular.

INTERNAL FEATURES

The barite deposits are largely interbedded barite, chert (figs. 49, 50), and minor limestone. Notable are the sharp bedding contacts between barite and chert and the absence of such features as altered wallrocks, abundant gangue minerals, and complex ore mineral suites which are commonly associated with other mineral deposits in the same host rock.

Composition.—Chert and limestone interbedded with barite in the deposits do not differ mineralogically or chemically from the host rock and need no further description.

Barite beds contain microscopically visible chert and limestone (apparently unreplaced remnants), carbonaceous matter, and small amounts of pyrite and iron oxide.

Veinlets of barite, quartz, and calcite transect many

FIGURE 49.—Photograph of polished and etched slab of barite with interbedded chert, Hilltop barite mine deposit. Light-colored beds are chert, whitened by hydrofluoric acid.

barite beds. In veinlets that contain both barite and quartz, quartz coats the walls and barite occupies the centers.

A salient feature of the barite deposits is the simplicity of the mineral assemblage. At most, only small amounts of quartz, calcite, and pyrite, in addition to the barite, have been introduced. Quite possibly, quartz and calcite are premineral and the pyrite a relic of the host rock. If so, only barite has been introduced.

The chemical composition of numerous specimens of barite are given in table 11. Large amounts of $CaCO_3$ and SiO_2 reflect inclusions of sedimentary limestone and chert in the sample, rather than introduction of calcite and quartz with the barite.

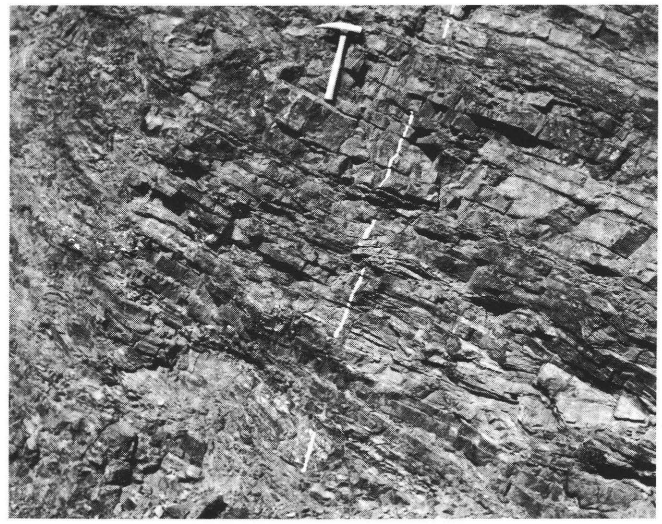

FIGURE 50.—Interbedded chert and barite, Argenta barite deposit. Barite beds are marked with a white stripe.

The forms in which many of the elements listed in table 11 occur are unknown but it is certain that some aluminum, magnesium, potassium, and sodium are present in the clay minerals of shaly partings. Some magnesium, strontium, and manganese are doubtless contained in limestone inclusions. Iron is present as iron oxide. Some sodium may be present as NaCl the chloride in fluid inclusions. Chromium, copper, nickel, titanium, and vanadium cannot be so easily accounted for.

The trace-element assemblage is virtually identical to that of neighboring limestone in the host rocks and differs from that of host-rock chert mainly in absence of boron (table 11).

Texture.—Most barite is nearly as fine grained and as thinly laminated as the host-rock chert (fig. 51). Laminae are commonly less than 1 mm thick. Separations between laminae resemble stylolites in their jagged cross sections and dark color. Where barite has been recrystallized it is bleached and the original textures are obliterated. Such barite is bladed and massive as shown in figure 52. Like the host-rock chert, barite contains thin shaly beds at intervals of 1 to a few inches which give it a slabby fissility.

Microfossils resembling those present in the host rock are rather common in some barite beds (fig. 53). Those in barite have been baritized and only the gross features are preserved. Where they are numerous they give the barite an augenlike texture in thin section.

Whereas chert is everywhere broken by cross fractures at about right angles to the bedding, barite is broken at less frequent intervals by random fractures. Consequently, barite float commonly consists of larger more irregular pieces than chert float. The miners

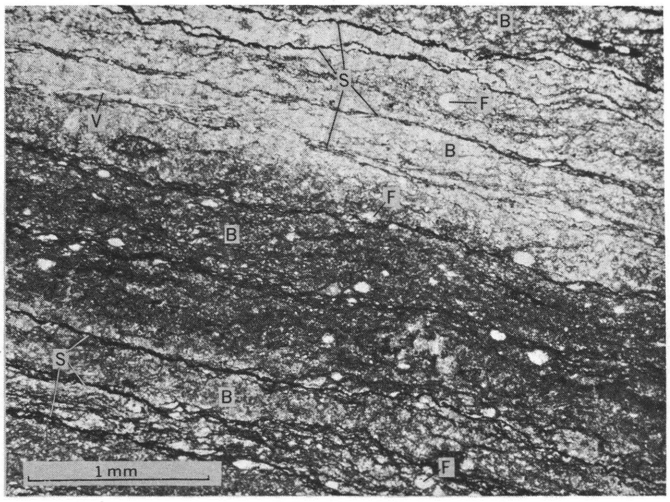

FIGURE 51.—Photomicrograph of thin-bedded barite, Hilltop barite (Valley View) mine. B, barite (equigranular texture, very fine grained); S, bedding surfaces or stylolites parallel to original bedding; F, baritized radiolaria; V, barite veinlet.

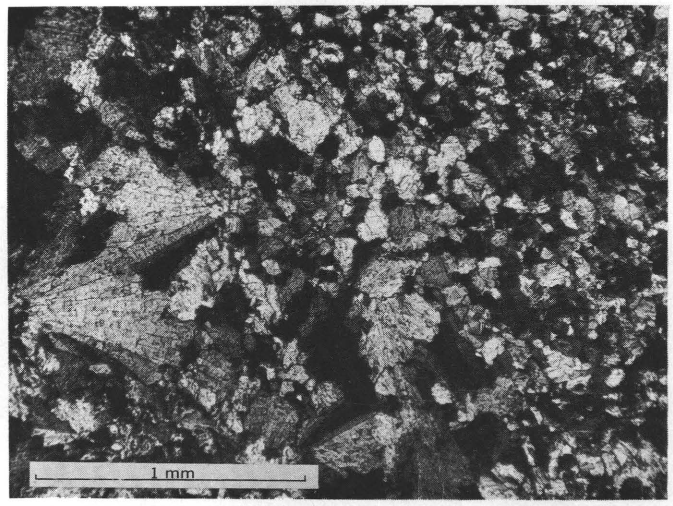

FIGURE 52.—Photomicrograph of massive barite showing mosaic texture, and some radiating blades, Bateman Canyon barite deposit.

exploit this tendency by screening out the smaller pieces of chert and retaining the coarser barite.

Color.—Barite is commonly dark gray, much like the associated chert, but in some places, owing to lack of organic matter, it is almost white.

Alteration.—Minute cubes of iron oxide scattered irregularly throughout the barite are apparently derived from pyrite by oxidation. However, these are no more numerous in barite than in the host rock distant from the barite deposits and nothing suggests the oxidation is connected with barite deposition.

ORIGIN AND AGE

Apparently barite replaced both chert and limestone. The contact between barite and host rock on the west side of the Greystone deposit (fig. 47) goes directly across the bedding without the aid of any known fault. Blades of barite cutting chert beds and the similarity between bedding of chert and barite indicate replacement of chert.

The abrupt change in composition from limestone to barite along the strike of what appears to be a single bed in the Hilltop deposit (fig. 45), the cutting of calcite grains by barite blades, the presence of brachiopods in both limestone and barite but not in chert, and the similarity in trace-element content between barite and limestone (table 11) indicate limestone replacement.

The discontinuity and coarseness of beds in all limestone observed in the Slaven Chert are inconsistent with other evidence pointing to limestone replacement, because barite is commonly finely and continuously bedded. But the stylolites that border and embay the calcite and included chert beds indicate solution may have destroyed some of the original texture. Possibly the limestone was baritized before the original bedding in the limestone was modified by recrystallization.

The field and microscopic evidence in favor of replacement seems quite convincing and is accepted as proving the barite deposits are of replacement origin. However, discordant contacts such as the abrupt termination of the Greystone ore bodies (fig. 47), the apparent transformation along the strike of limestone to barite in the Hilltop deposit (fig. 45), as well as other field evidence, while not incompatible with replacement, are subject to other interpretations. Exposures are not good and undiscovered faults may account for at least some of the discordant contacts. Although a search was made in every deposit, no bed was observed to pass along the strike directly from chert or limestone to barite. Microscopic evidence of replacement such

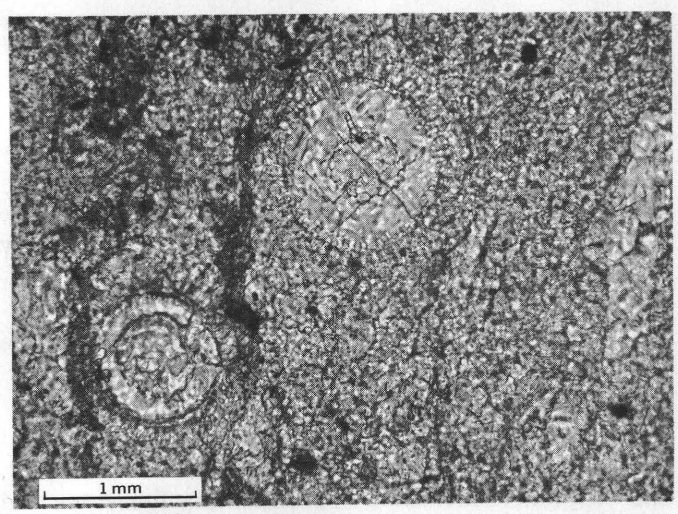

FIGURE 53.—Photomicrograph of barite, showing radiolarian preserved in baritic replacement body.

as blades of barite cutting chert bedding and baritized fossils might be the result of minor secondary redistribution such as commonly occurs in all rocks.

The barite deposits differ from neighboring metal deposits in several important respects. Whereas metal deposits are obviously associated with the northwest-trending band of intrusive rocks, the barite deposits apparently are unrelated to concentrations of intrusives but rather are associated with bedded chert, especially with Devonian chert. The form of metal deposits is controlled by shear zones whereas that of barite deposits is controlled to a large extent by bedding. Whereas metal deposits contain the common mesothermal assemblage of pyrite, arsenopyrite, chalcopyrite, and tetrahedrite, little, if anything, but barite has been introduced into barite deposits. Metal deposits are commonly associated with recrystallized, bleached, and sericitized rock but barite deposits are remarkable for the absence of host-rock alteration. It is therefore concluded that the barite deposits are not directly related in origin to metal deposits of the region.

The lack of correlation with intrusive rocks, lack of accessory hydrothermal minerals and wall alteration, the positive correlation with sedimentary chert, the tendency to conform with bedding, the generally high barium content of bedded chert, all suggest the possibility of diagenetic replacement of sea-bottom sediments by barium present in sea water. However, the discontinuity of deposits, great stratigraphic thickness in relation to lateral extent, the abrupt discordance revealed in places by mapping and unexplained by visible faults, indicate epigenetic replacement.

If the barite deposits are distinctly younger than the host rock, as seems most likely, barium might have come from the host rock or from deeply buried or distant intrusives. The latter possibility is in keeping with the unusually high barium content of most of the Tertiary igneous rocks of the eastern Great Basin. (See p. 88.)

The association of large barite deposits with bedded chert formations and the concentration of known deposits in a restricted area are obvious guides to exploration.

HILLTOP BARITE DEPOSIT

The Hilltop or Valley View mine is low on the north slopes of the Shoshone Range, about a mile west of Slaven Canyon. It is readily reached from Battle Mountain, the nearest shipping point, by a good gravel road about 15 miles long. In the area of Slaven Canyon (fig. 45), beds of the Slaven Chert strike north and dip east rather uniformly. Both the barite and limestone have the same general attitude.

The barite crops out discontinuously along a north-south belt extending from about the center of sec. 2 to the center of sec. 11, T. 30 N., R. 46 E. Most of the exploratory work and all the production of the Hilltop barium mine have come from an area just north of the Mount Lewis quadrangle. Nearly coincident with the barite are scattered outcrops of limestone. The barite bed is about 40 feet thick and the limestone bed about 70 feet. To the north both the limestone and barite beds are cut by the Basin Range fault. In the southern part of the area shown in figure 45, the limestone appears to merge into barite, and both this bed and the main barite bed pinch out and reappear south of the map area. Barite is everywhere interbedded with chert. In the main pit at the northeast corner of the area of figure 45 it is in places interbedded with limestone also.

BATEMAN CANYON DEPOSIT

The Bateman Canyon deposit (fig. 46) is high on the east wall of the canyon in the NW ¼ sec. 35, T. 30 N., R. 46 E. It is reached by a good gravel road from Battle Mountain up Crum and Bateman Canyons, a distance of about 18 miles. It is a cluster of pods of both well-bedded and massive barite in the Slaven Chert. The country rock is not well enough exposed to determine the attitude of the chert bedding but the strike of the bedding within the barite pods is fairly consistently northwest. Dips are very steep. The deposit is not well exposed, but the relation between contacts and bedding suggests that barite cuts across chert bedding (that is, occurs in disconformable lenses). The only limestone seen near the Bateman Canyon deposit is 1 mile north where limestone and barite with obscure mutual relations crop out together.

GREYSTONE DEPOSIT

The Greystone deposit (fig. 47) is in secs. 23, 24, 25, and 26, T. 28 N., R. 45 E. (unsurveyed), about a mile northeast of Cooks Creek. It is accessible from Battle Mountain by a paved road as far as Mill Creek (about 20 miles) and a fair, but very tortuous gravel road across Mill Creek Summit, an additional 12 or 13 miles. It is also reached from Beowawe by a good paved road to a point near Gold Acres (about 25 miles), then by gravel road southwest and west into Elder Creek and then northwest to the mine, an additional 14 or 15 miles. Because this road has easier grades and curves than the Mill Creek road, it has been used more for hauling barite, despite its greater length.

The mine is in a slice of Slaven Chert between two thrust faults. It consists of a large body of interbedded chert, argillite, limestone, and barite nearly as thick stratigraphically as it is long. On strike at distances of several hundred feet are similar smaller bodies and at least one limestone lens. The strike of bedding in the deposit and in the host chert is prevailingly northwest and the dip is southwest. Although the deposits are somewhat elongate parallel to the bedding, the discordance of the east and west boundaries of the main body with the bedding indicates that replacement transgressed bedding surfaces extensively.

Molds of the Devonian brachiopod *Halorella* are present both in the limestone lens shown in figure 47, and locally in barite of the main deposit.

In view of the discordant contacts and presence of thrust faults, the extent of the deposit in depth is unpredictable.

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U.S. GOVERNMENT PRINTING OFFICE: 1965 O—736-006