Geology of the Magruder Mountain Area Nevada-California

GEOLOGICAL SURVEY BULLETIN 1251-H

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By EDWIN H. McKEE

CONTRIBUTIONS TO GENERAL GEOLOGY

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A description of the Precambrian, lower Paleozoic, and Tertiary stratigraphy and a discussion of the granitic rocks and structure

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1968

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

William T. Pecora, *Director*

For sale by the Superintendent of Documents, U.S. Government Printing Office
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GEOLOGY OF THE MAGRUDER MOUNTAIN AREA NEVADA-CALIFORNIA

By Edwin H. McKee

ABSTRACT

The oldest rocks exposed in the northern half of the Magruder Mountain and Soldier Pass quadrangles are Precambian, Cambrian, and Ordovician sedimentary strata. These are intruded by Middle Jurassic granitic rocks (dated by the potassium-argon method) and are unconformably overlain by late Tertiary and Quaternary sedimentary and volcanic rocks.

The more than 15,000 feet of Precambrian and Cambrian sedimentary rocks in the Magruder Mountain area constitute a conformable sequence of limestone, sandstone, and siltstone very similar to that in the classic Waucoban section of the Inyo Range to the west. The entire Lower Cambrian series is present in these sequences, and it contains olenellid zones representing the oldest Cambrian in North America.

Ordovician rocks in the area mapped include graptolitic shale and chert of Middle Ordovician age. The relationship of these rocks to the Cambrian strata is unclear, but possibly the Ordovician rocks are allochthonous and were thrust into the area during the Antler orogeny in Late Devonian or Early Mississippian time.

Tertiary rocks of the Miocene and Pliocene Esmeralda Formation fill the south end of the Fish Lake Valley graben. These rocks, which consist of tuff, tuffaceous siltstone, sandstone, and conglomerate, are more than 1,500 feet thick. Thickness and facies variations indicate that these sediments were deposited in a basin which coincided with the present Fish Lake Valley. Elsewhere in the area mapped, upper Tertiary rocks are in positions of erosion rather than of deposition and are thus truncated by an unconformity. West of the Palmetto and Magruder Mountains, a large pediment that is now being dissected is underlain by several hundred feet of poorly consolidated tuffaceous sandstone and conglomerate derived from the Palmetto Mountains.

Structure in the Palmetto Mountains and the adjoined region to the northwest is typified by thrust faults and recumbent folds, whereas that to the southwest at Magruder Mountain, in the Last Chance Range and the Inyo and White Mountains is characterized by open folds and numerous small steeply dipping faults. The difference in structural style from the south to north may be related to the Antler orogenic belt, which possibly extended into the northern part of the region. Rocks in the Last Chance Range lie south of this zone of late middle Paleozic thrust faulting.

Deformation in the Jurassic by plutonic intrusions is evident. The last phase of the structural history is Tertiary to Recent block faulting which cuts all rocks in the region. The largest single structural feature in the area is the Death Valley-Furnace Creek fault zone, along which there has apparently been at least 20 miles of right-lateral displacement.

INTRODUCTION

The Magruder Mountain area lies in eastern California and southwestern Nevada about 15 miles northwest of the northernmost part of Death Valley. The area includes approximately 200 square miles making up the northern half of the Magruder Mountain and Soldier Pass 15-minute quadrangles, California-Nevada. The Palmetto, Magruder, and Sylvania Mountains and a series of unnamed hills northwest of the Last Chance Range lie within the area (pl. 1; fig. 1). Fish Lake Valley, a large graben east of the White Mountains and northeast of the Death Valley graben, extends into the area from the northwest.

The mountains included in this study are composed of Precambrian, Cambrian, and Ordovician sedimentary rocks that are intruded by Jurassic plutonic rocks. Upper Tertiary lake beds and volcanic rocks mantle parts of the area and in places are overlain by Pleistocene sediments. These Cenozoic deposits are generally related to the present topography; they are thicker in valleys and are thin or absent in the mountains. In some places, however, these deposits have been uplifted along Cenozoic faults.

ACKNOWLEDGMENTS

Thanks are given to Prof. C. M. Gilbert, University of California, Berkeley, for numerous suggestions and help during the mapping of this area. The writer also expresses his thanks to Prof. J. Wyatt Durham, University of California, Berkeley, for helpful discussions on paleontology and to Prof. C. A. Nelson, University of California, Los Angeles, for stratigraphic orientation and identification of trilobites. Prof. W. B. N. Berry, University of California, Berkeley, identified the graptolites.

Field trips were made to adjacent areas, and discussions of regional features were held, with J. P. Albers and J. H. Stewart, U.S. Geological Survey, and with R. J. Moiola and J. W. Whetten, University of California, Berkeley.

SEDIMENTARY ROCKS

PRECAMBRIAN AND LOWER CAMBRIAN ROCKS

The Precambrian and Lower Cambrian strata in the Magruder Mountain area are similar to those in the Inyo Mountains. These strata in the Inyo Mountains were made classic by Walcott's study of the Waucoba Spring section (1908, p. 185). The same sequence of

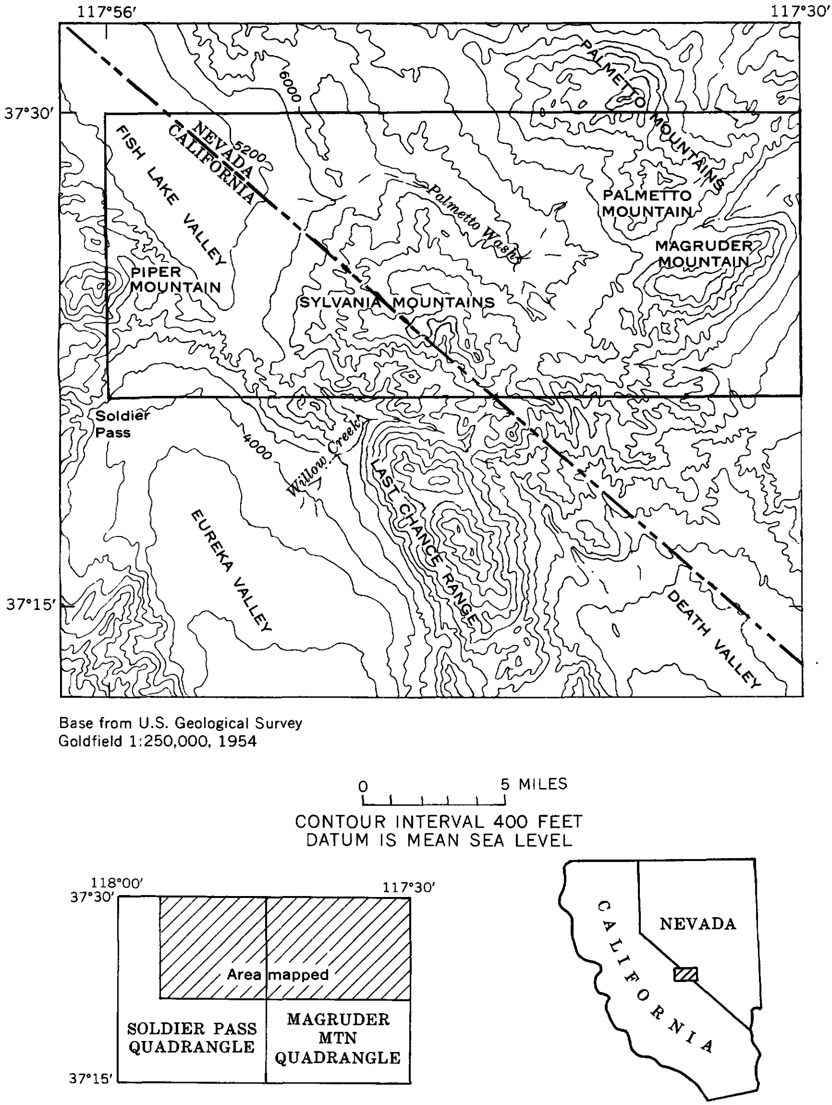

FIGURE 1.—Index maps of the Magruder Mountain area.

strata occurs to the north and east of the Magruder Mountain area for at least 30 miles and is clearly recognizable in the Silver Peak Mountains, Nev., and in areas as much as 15 miles east of that range. Thus, an area of at least 4,000 square miles contains strata of Precambrian and Early Cambrian age that are similar in thickness and lithologic character; this area extends west to the west edge of the White and Inyo Mountains, east to the Goldfield-Tonopah, Nev., region, north to the north end of the Silver Peak Range, and south to the north end of Death Valley (figs. 2, 3).

Southward from the region defined above, the upper Precambrian and Lower Cambrian rocks retain the gross character of the Inyo sequence, but formational thicknesses differ, and lithology, in many places, is not the same. Strata in this area are referred to as the Death Valley sequence and were described by Nolan (1929) from the Spring Mountains, Nev., by Hazzard (1937) from the Nopah and Resting Springs Mountains, Calif., and by others. Precambrian and Lower Cambrian rocks from the Inyo and Death Valley region have been correlated by J. H. Stewart (written commun., 1965).

The eastern limit of the Inyo sequence is poorly defined. The nearest Precambrian-Cambrian sections to the east are in the Quartzite Mountains and the Belted Range, approximately 70 miles east of Magruder Mountain. These rocks more nearly resemble the Death Valley than the Inyo sequence, although correlation between the units is possible.

Precambrian rocks of the Magruder Mountain area consist of alternating carbonate rock and fine-grained quartzose sandstone and siltstone. Four formations are recognized; in ascending order, they are the Wyman Formation, the Reed Dolomite, the Deep Spring Formation, and the Andrews Mountain Member of the Campito Formation. These strata appear to be conformable, but some evidence has been reported of unconformities at the top of the Wyman Formation (Maxson, 1934; Nelson, 1962) and at the top of the Deep Spring Formation (Kirk, in Knopf, 1918, p. 25).

WYMAN FORMATION

The oldest rocks in the area are referred to as Wyman Formation. This formation crops out only along the northeast side of the Sylvania Mountains. It consists of phyllitic to hornfelsic sandstone and shale with interbeds of dolomite.

Inasmuch as only the top of the Wyman Formation in the Soldier Pass and Magruder Mountain quadrangles is exposed, the total thickness of the formation is not known. At the type section, approximately 20 miles to the southwest, Maxson (1934) gave a thickness of about 3,700 feet and Nelson (1962) at least 9,000 feet.

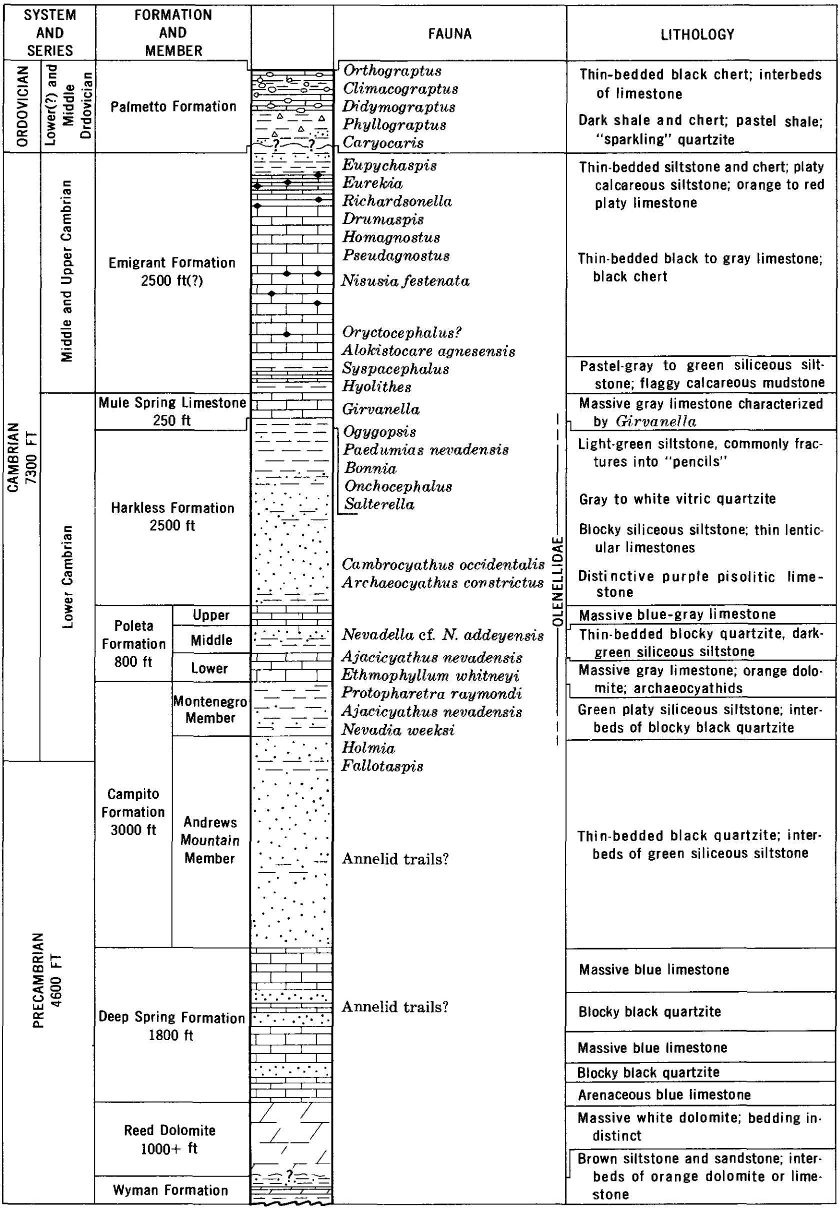

FIGURE 2.—Composite stratigraphic column of Precambrian and Paleozoic rocks.

The upper contact with the Reed Dolomite is sharp, and no angular discordance occurs between the Wyman and Reed Formations. The existence of an unconformity between these formations, as reported in the White Mountains (Maxson, 1934; Nelson, 1962), is not obvious.

The Wyman Formation in the Sylvania Mountains is probably correlative with the type Wyman of the White Mountains. Although no fossils are known in the formation, similarity in lithology and position in the stratigraphic sequence lead to this conclusion.

REED DOLOMITE

The Reed Dolomite, which overlies the Wyman Formation, is a massive orange-brown to white finely to coarsely crystalline dolomite. The dolomite commonly has a sugary texture and locally contains pisolitic and oolitic structures. Upon weathering, it commonly develops a coarsely pitted surface. Bedding is very weakly developed throughout most of the formation.

The thickness of the Reed Dolomite in the area mapped is not known. The writer estimates the thickest section to be more than 1,000 feet. In the White Mountains, 2,000 feet was reported by Kirk (in Knopf, 1918, p. 24) and Nelson (1962, p. 141). At Gold Point, Nev., to the east, Albers and Stewart (1962) estimated a thickness of approximately 2,000 feet.

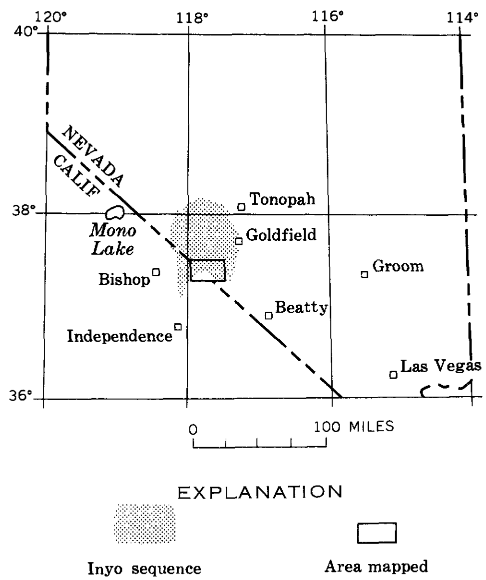

FIGURE 3.—Distribution of Precambrian and Cambrian rocks similar to the Waucoban section of the White and Inyo Mountains.

The upper contact of the Reed Dolomite seems to be conformable. Although the Reed and the Deep Spring Formations in the White Mountains were believed by Kirk (in Knopf, 1918) and Maxson (1934) to be separated by an unconformity, more recent work by Nelson (1962) in the same region indicated that no unconformity is present.

The Reed Dolomite of the Magruder Mountain area is similar lithologically, and probably correlates with, the type Reed of the White Mountains and with thick dolomites in other parts of Esmeralda County, Nev., mapped as Reed Dolomite by Albers and Stewart (1965).

DEEP SPRING FORMATION

Conformably above the Reed Dolomite is 1,800 feet of limestone and quartzite of the Deep Spring Formation. At Magruder Mountain, massive to thin-bedded blue to gray limestone (arenaceous in places and locally oolitic) makes up about four-fifths of the formation. The rest of the formation is dark quartzite interbedded with siliceous siltstone.

Lateral variations in thickness and lithology of individual units are characteristic of the Deep Spring Formation, and most thin units cannot be traced for more than a few miles. The overall thickness and lithology of the formation, however, remain essentially the same from the west flank of the Inyo Range to the central part of Esmeralda County, Nev., a distance of approximately 70 miles.

The Deep Spring Formation of the Magruder Mountain area correlates with the type Deep Spring Formation of the White Mountains and may also be correlative with the lower part of the Wood Canyon Formation of the Death Valley region (J. H. Stewart, written commun., 1965).

CAMPITO FORMATION

The Campito Formation conformably overlies the Deep Spring Formation and conformably underlies the Poleta Formation. It is divided into two members (Nelson, 1962) : a thick lower member—the Andrews Mountain Member—and a thinner upper member—the Montenegro Member. The contact between the two members is gradational.

Typical quartzite of the Andrews Mountain member is dark gray to black, evenly bedded, and dense, with a few interbeds of dark siliceous siltstone. The quartzite is fine to medium grained, with less than 10 percent feldspar or lithic fragments set in a dark argillaceous matrix now converted to chlorite, biotite, sericite, and magnetite. Approximately 25 percent of the rock is matrix. The beds range in thickness from several inches to several feet and are almost everywhere broken by a system of joints normal to bedding. Individual beds have very uniform thickness, but bedding surfaces show irregularities including ripple marks, fucoid structures, and worm (?) trails. Locally, beds are cross-laminated, and small channel scours are present.

The thickness of the Andrews Mountain Member cannot be determined in the Magruder Mountain area, but in the White and Inyo Mountains it is 2,500–2,800 feet thick (Nelson, 1962, p. 141), and it is probably of comparable thickness in the Magruder Mountain area.

Siliceous siltstone is progressively more abundant toward the top of the Campito Formation. Where this lithology constitutes the dominant rock type, it is mapped separately and is designated the Montenegro Member. It is from 500 to 700 feet thick.

Lithologically, the Montenegro Member is similar to the underlying Andrews Mountain Member but is finer grained (silt to fine-sand size). It contains approximately equal amounts of fine-grained quartz and recrystallized argillaceous material.

At Magruder Mountain, the contact of the Campito Formation with the overlying limestone of the Poleta Formation is gradational through a stratigraphic interval of 10–20 feet. Siltstone of the Montenegro Member becomes increasingly limy upward and grades into massive limestone of the Poleta Formation. In other regions this contact is sharp; thin-bedded siltstone underlies massive limestone.

Trilobites occur throughout most of the Montenegro Member but are less abundant where the rock is relatively coarse grained. The fauna includes:

Trilobites (identified by C. A. Nelson, 1965; locs. 6, 8, 14, 15. 18, 19, 21, and 23 on pl. 1):
Fallotaspis sp.
Nevadella cf. N. addeyensis Okulitch
Holmia sp.
Nevadia weeksi Walcott
Archaeocyathids (identified by E. H. McKee, 1962; locs. 18 and 19 on pl. 1):
Ethmophyllum whitneyi Meek
Annelid trails (?)

The Campito Formation in the Magruder Mountain area is similar in thickness, lithology, and stratigraphic position to the type Campito of the Inyo Range with which it is correlated. Thick quartzite units, occupying a stratigraphic position comparable to that of the Campito Formation, are also recognized at many localities in the southern California-Nevada region. Probably the Montenegro Member of the Campito Formation correlates with units 4D and 4E of the Wood Canyon Formation, described by Hazzard (1937) in the Nopah and Resting Springs Mountains. Regional studies of the Precambrian and Lower Cambrian rocks in the southern Great Basin by J. H. Stewart (written commun., 1965) seem to indicate that the unfossiliferous quartzites and siltstones of the lower part of the Wood Canyon Formation probably correlate with the Andrews Mountain Member.

POLETA FORMATION

Strata of the Poleta Formation conformably overlie the siliceous siltstone of the Montenegro Member of the Campito Formation.

The formation can be divided into three members in the Magruder Mountain area: a lower carbonate member, a middle siltstone and quartzite member, and an upper limestone member.

The lower member of the formation is a massive blue and orange mottled limestone characterized by abundant archaeocyathids. Locally, it is oolitic. At Magruder Mountain this limestone is 300 feet thick, but its thickness varies markedly and is only 150 feet in the hills northwest of the Last Chance Range.

The middle member of the Poleta Formation consists largely of thin-bedded dark-green siliceous siltstone and sandstone. Four hundred feet of siliceous strata are exposed at Magruder Mountain, but the unit thins to less than 200 feet in the Last Chance Range.

Interbedded with siltstone beds are thin sandstone units, most of which are less than 1 foot thick. The upper surfaces of many siltstone and sandstone beds show ripple marks or worm(?) trails and trilobite tracks(?) (fig. 4); a few show some mud cracks.

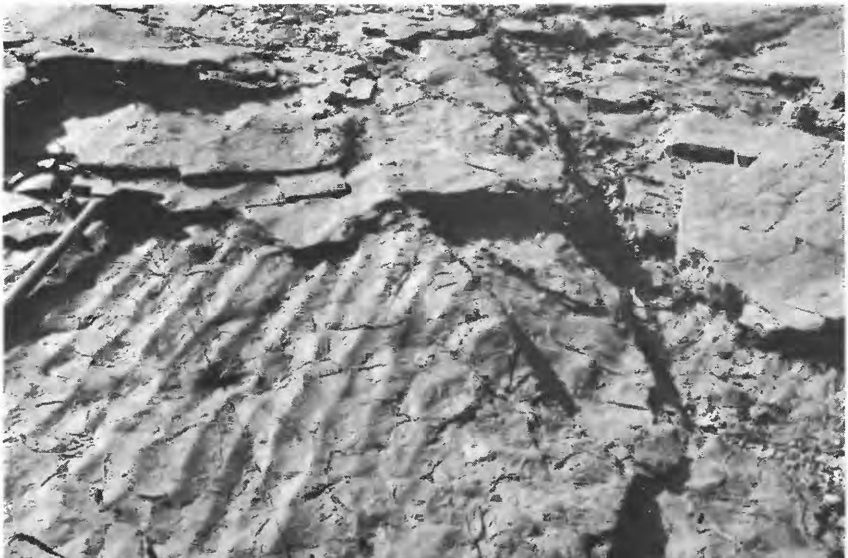

FIGURE 4.—Ripple marks and worm trails on a typical bedding surface of siltstone from the Poleta Formation.

Limestone and quartzite are the dominant rock types near the top of the member and grade upward into limestone of the upper member. Both the lower carbonate unit and the upper unit are medium-grained limestone containing a small amount of quartz and albite. The upper limestone can normally be distinguished from the lower limestone, however, by the absence or scarcity of archaeocyathids.

The Poleta Formation contains more fossils than any other Lower Cambrian formation. Fossils from this formation include the following:

- Trilobites (identified by C. A. Nelson, 1965):
  - *Nevadella* cf. *N. addeyensis* Okulitch (locs. 1, 3, 9, 12 and 33 on pl. 1)
  - *Holmia* sp. (locs. 13 and 34 on pl. 1)
- Archaeocyathids (identified by E. H. McKee):
  - *Ethmophyllum whitneyi* Meek (locs. 7, 16 and 30 on pl. 1)
  - *Protopharetra raymondi* Okulitch (loc. 30 on pl. 1)
  - *Ajacicyathus nevadensis* (Okulitch) (loc. 30 on pl. 1)
- *Hyolithes* sp.
- *Salterella* sp.
- "*Scolithus*"

The Poleta Formation can be recognized from its type area on the west side of the White Mountains to localities near Tonopah, Nev., a distance of approximately 65 miles. Sections in the Last Chance Range and at Magruder Mountain are approximately midway in this region.

As a distinct mappable unit, this formation is difficult to trace into the Death Valley region, but the position in the stratigraphic column and the occurrence of *Wanneria*(?) *gracile* Walcott (*Nevadella* cf. *N. addeyensis*) in unit 4E of Hazzard's (1937) Wood Canyon Formation indicates that the upper part of unit 4E, unit 4F, and the lower part of unit 4G are probably correlative with the Poleta Formation.

HARKLESS FORMATION

Dominantly detrital deposits conformably overlie the Poleta Formation and make up the Harkless Formation.

The Harkless Formation is divided into three parts, which are, in ascending order (1) a sequence of sandstone and siltstone about 300 feet thick; (2) a sequence of quartzite, siliceous siltstone, and thin lenticular limestones about 2,000 feet thick; and (3) a sequence of siltstone and shale about 600–700 feet thick.

The lower 300 feet of the Harkless Formation is mostly fine-grained sandstone with interbeds of siliceous siltstone. The strata are similar in appearance to the middle member of the underlying Poleta Formation but can be distinguished from it in the field by being lighter green and gray, by having a higher ratio of sandstone to siltstone, and having several persistent marker beds, the most distinctive of which are a micaceous purple siltstone and one or more thin limestones that contain large (½-in. diameter) purple pisolites (fig. 5). In addition, thin lenticular limestones composed almost entirely of archaeocyathids are numerous. These seem to be small reefs or bioherms.

In the Harkless Formation above the archaeocyathid beds, quartzite is the most abundant rock type. It is thinly and uniformly bedded and, in certain areas, is similar to thin-bedded Campito quartzite, from which it can be distinguished only by its lighter color and more vitric nature. Beds of this quartzite are from 1 to 3 feet thick and are extremely resistant to erosion. Dark siliceous siltstone is interbedded with Harkless quartzite but is obscured by blocks of quartzite. Locally interbedded in this sequence are thin limestones containing abundant well-rounded quartz grains, which etch out on weathered surfaces.

Light-green siltstone and shale make up the uppermost 600–700 feet of the Harkless Formation. The siltstone is less siliceous than any other in the Precambrian and Cambrian sequence. In many places it has the fissility of "paper" shale, whereas in other places it fractures in the form of "pencils" (fig. 6). The small conical fossil *Salterella* is abundant in the shales (fig. 7).

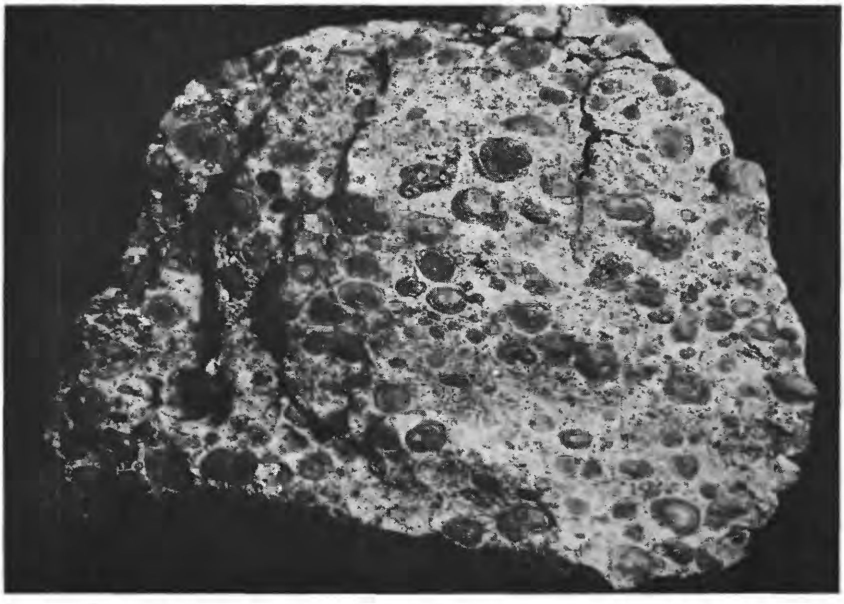

FIGURE 5.—Pisolites from the "purple pisolite" marker bed near the base of the Harkless Formation. × 0.5.

FIGURE 6.—Siltstone "pencils" in the Harkless Formation.

FIGURE 7.—*Salterella* in siltstone near the top of the Harkless Formation. This fossil is abundant in this stratigraphic interval. × 2.

The Harkless Formation is unfossiliferous except near the base and near the top. Fossils from the basal part include:

Tribolites (identified by C. A. Nelson (1965); locs. 2 and 26 on pl. 1):
- *Holmia* sp.
- *Paedumias clarki* Resser

Archaeocyathids (identified by E. H. McKee; locs. 11 and 22 on pl. 1):
- *Archaeocyathus constrictus* (Raymond)
- *Cambrocyathus occidentalis* (Okulitch)

Girvanella (?)

"*Scolithus*"

The fauna from the upper siltstone is:

Trilobites (identified by C. A. Nelson (1965); locs. 17, 20, 24, 25, 28, 29, and 31 on pl. 1):
- *Ogygopsis* sp.
- *Onchocephalus* sp.
- *Paedumias nevadensis* Walcott
- "*Ptychoparia*" sp.

Salterella sp.

The Harkless Formation in the Magruder Mountain area is correlated with the Harkless and overlying Saline Valley Formations described by C. A. Nelson (1962) in the Waucoba Mountain quadrangle in the Inyo Mountains. In the Death Valley region, it probably corresponds to part of unit 4H and possibly units as high as 4K in the Wood Canyon Formation as described by Hazzard (1937).

MULE SPRING LIMESTONE

Siltstone in the upper part of the Harkless Formation grades upward into limestones of the Mule Spring Limestone.

The lowermost 50 feet of this formation is thin-bedded platy gray aphanitic limestone, which locally has concentrations of trilobite fragments on bedding surfaces (fig. 8). Higher in the formation the bedding is thicker and the texture is coarser; the upper 200 feet is a massive dark-gray limestone that typically contains concentric nodules of the algae *Girvanella*. These nodules, ½–1 inch in diameter, are the most distinctive feature of the Mule Spring Limestone (fig. 9).

Trilobites have been found by J. P. Albers and J. H. Stewart (oral commun., 1961) in platy limestone in the lower part of the formation at localities north of the Magruder Mountain area. These fossils, identified by A. R. Palmer, U.S. Geological Survey, are:

Trilobites:
- *Bonnia* sp.
- *Bristolia* sp.
- *Fremontia* sp.

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FIGURE 8.—Trilobite "hash" in thin-bedded limestone near the base of the Mule Spring Limestone. This "hash" is typical of the stratigraphic horizon. × 0.5.

The upper massive limestone contains no fossils except *Girvanella*, which is common at the top of the Lower Cambrian sequence.

Cambrian algal limestone is known from many localities in the desert region of California-Nevada. In regions west of Las Vegas, Nev., algal limestone was considered by Hazzard (1937) to be characteristic of the uppermost Lower Cambrian, whereas in central Nevada and in the Grand Canyon (McKee, 1945) it occurs well into the Middle Cambrian. Similar limestones has been noted in other Lower and Middle Cambrian formations of the Inyo Range, but algae are nowhere as abundant as in the Mule Spring Limestone.

The Mule Spring Limestone is considered Lower Cambrian because it contains olenellid trilobites. Middle and Upper Cambrian trilobites are present in the Emigrant Formation that directly overlies it. The Mule Spring Limestone is probably correlative with upper Lower Cambrian limestones of the Wood Canyon Formation, described by Hazzard (1937) from the Nopah and Resting Springs Mountains.

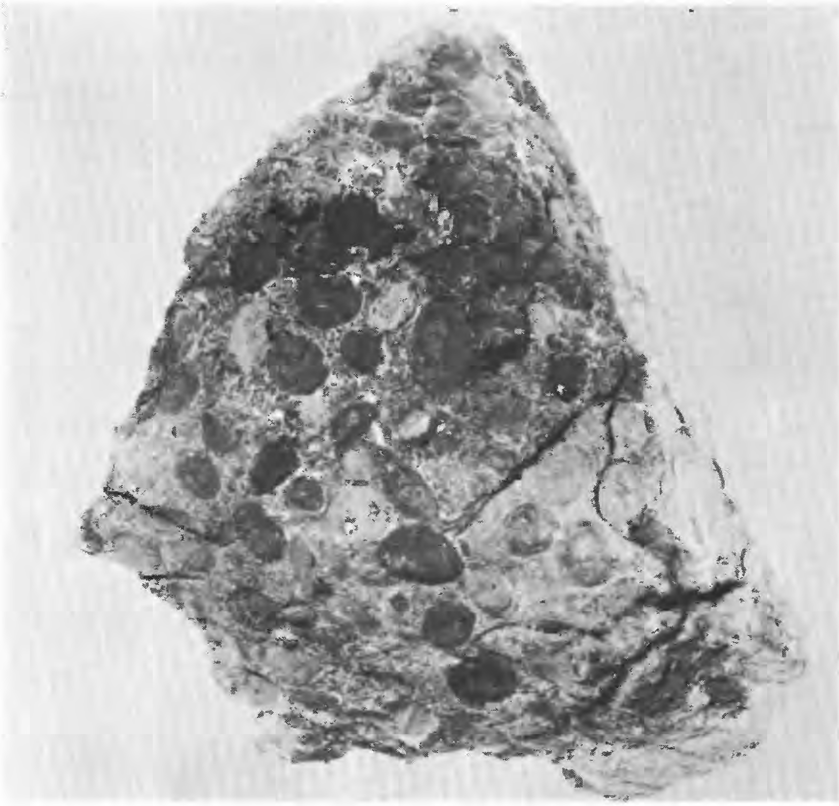

FIGURE 9.—*Girvanella* typical of the upper Lower Cambrian Mule Spring Limestone. × 0.5.

MIDDLE AND UPPER CAMBRIAN ROCKS

Conformably above the Precambrian and Lower Cambrian strata, which are composed mostly of sandstones and shale, is a thick sequence of limestone and minor amounts of clastic quartzose rock and chert. Carbonate rock also occurs throughout the Middle and Upper Cambrian in other parts of the California-Nevada region; it is present in the southern Inyo Mountains, in the Death Valley region, and in central and eastern Nevada.

In the Palmetto Mountains and the hills northwest of the Last Chance Range, there is at least 1,500 feet of thin-bedded platy limestone containing Middle Cambrian fossils. In adjacent areas to the east and north, similar thin-bedded limestone contains an Upper Cambrian fauna. A complete Middle and Upper Cambrian section cannot be described from any one locality, but at least 2,000 feet of thin-bedded limestone ranging from Middle to Late Cambrian is believed to occur in the area. At one locality in the Silver Peak Range and another area 10 miles east of the Silver Peak Range, shale, chert, and massive arenaceous limestone appear to overlie the thin-bedded limestone, and these beds are probably the uppermost Cambrian rocks of the region.

EMIGRANT FORMATION

The Emigrant Formation, which conformably overlies the Mule Spring Limestone, is composed mostly of limestone. Siliceous siltstone and platy limestone, which form the lower part of the formation, grade upward into limestone, which makes up the bulk of the formation.

The upper part of the Emigrant Formation is a thick and distinctive unit. It is composed of thin-bedded blue-gray to gray limestone and dolomite alternating with buff to black bands of chert or of limonite-stained carbonate rock ½–2 inches thick. The entire thickness of this limestone is not known, but several partial sections show more than 1,000 feet of carbonate rock with either the base or top missing.

Distinctive beds of intraformational limestone conglomerate or breccia occur at many horizons throughout the unit (fig. 10). These beds are 1–10 feet thick and consist of flat plates of blue-gray limestone 1–6 inches in diameter in a matrix of similar blue-gray limestone or dolomite.

Trilobites from the lower siltstones (identified by C. A. Nelson, 1959) are:

Syspacephalus sp.
Ehmaniella(?) sp.

Collections from the same stratigraphic position from localities approximately 15 miles northeast of the Magruder Mountain area made by J. P. Albers and J. H. Stewart and identified by A. R. Palmer include:

Oryctocephalus(?) sp.
Alokistocare cf. *A. agnesensis* Walcott
Syspacephalus sp.

These trilobites represent the oldest Middle Cambrian fauna in this area.

Fossils are rare throughout the upper limestone. Several brachiopods from approximately 500 feet above the base of the unit (locs. 4 and 5 on pl. 1) were identified by W. B. N. Berry as *Nisusia festinata*, a species typical of the Lower and Middle Cambrian (Schuchert and Cooper, 1932, p. 45).

FIGURE 10.—Intraformational conglomerate typical of some limestones in the Emigrant Formation. Flat plates of limestone in limestone matrix.

Fossils collected from higher stratigraphic positions in the Silver Peak Range by J. P. Albers and J. H. Stewart comprise characteristic middle Upper Cambrian faunas, as determined by A. R. Palmer:

Richardsonella sp.
Drumaspis sp.
Homagnostus sp.
Idahoia(?) sp.
Pseudoagnostus sp.

The trilobites *Eupychaspis* and *Eurekia*, from beds near the top of the Emigrant Formation, are characteristic of the uppermost Cambrian (A. R. Palmer, oral commun., 1961); hence, the boundary between the Middle and Upper Cambrian lies somewhere within the thick, continuous sequence of Emigrant limestone.

ORDOVICIAN ROCKS

Lower(?) and Middle Ordovician shale and chert more similar to the siliceous-volcanic western assemblage than to the carbonate eastern assemblage of Ordovician strata in the Great Basin are exposed in the area mapped. The relationship of these rocks to the Upper Cambrian limestone is uncertain, for in most places the Ordovician rocks are in fault contact with older strata. The total thickness of the Ordovician is also unknown, because faulting has made all sections of these rocks very difficult to measure. All the Ordovician rocks are assigned to the Palmetto Formation.

PALMETTO FORMATION

Ordovician rocks of the Silver Peak Range were named the Palmetto Formation by H. W. Turner in 1902 (p. 265). They consist of graptolitic shale, thin-bedded limestone, and black chert. Turner believed them to be Early Ordovician (Beekmantown and Normanskill) and to overlie conformably the Emigrant Formation, which he believed to be Late Cambrian.

In the Palmetto Mountains, rocks of the Palmetto Formation are abundantly exposed; however, no single stratigraphic section including them is complete. Structural relationships are so complex in this area that a reliable composite section has not been compiled, and the relationship to the underlying Cambrian has not been determined.

Four principal types of rock make up the Palmetto Formation: (1) thin-bedded black chert; (2) interbeds of blue limestone; (3) pale-gray to purple shale; and (4) brown quartzite that has the texture and color of maple sugar and shows no stratification. The chert, shale, and quartzite are all distinctive rock types and differ from all the rocks of the underlying Precambrian and Cambrian sequence. Locally, where limestone interbeds in the black chert are numerous, the Palmetto Formation resembles the upper part of the Emigrant Formation.

The suite of Ordovician rocks (black chert and dark graptolitic shale) indicates a change in depositional environment from that of the Cambrian, when limestone, quartzite, and siliceous siltstone were the chief deposits.

The occurrence of light-colored felsic rocks that seem to be altered volcanic lavas or tuffs was reported by Turner (1909, p. 243), who concluded that "it is thus certain that in Ordovician time there were volcanic eruptions in the region." These rocks, however, are felsite dikes and sills which intrude the Cambrian as well as the Ordovician south of the area that Turner mapped. Without further information, they cannot be definitely assigned an Ordovician age.

Fossils in the Palmetto Formation (locs. 27 and 32 on pl. 1, identified by W. B. N. Berry, 1961) are:

Graptolites:
- *Orthograptus truncatus* (Lapworth)
- *Orthograptus* cf. *O. pageanus* (Lapworth)
- *Climacograptus* cf. *C. typicalis* Hall
- *Phyllograptus anna* Hall
- *Didymograptus* sp.

Caryocaris

Linguloid Brachiopods

The graptolites are indicative of an age span within the late Early and the Middle Ordovician (W. B. N. Berry, oral comm., 1961).

Ordovician rocks closest to those in the Silver Peak Range are exposed to the north and northwest at distances of approximately 20 miles. A section in the Candelaria Hills that is approximately 4,000 feet thick was described by Ferguson and Muller (1949, p. 45). It consists of dark silicified limestone overlain by black chert and slate. East of this area, in the Monte Cristo Range, chert and shale similar to that in the upper part of the Ordovician in the Candelaria Hills crop out (Ferguson and Muller, 1949, p. 49). Exposed approximately midway between the Monte Cristo Range and the Silver Peak Range are Ordovician shale and cherty limestone that are referred to the Palmetto Formation by H. W. Turner (unpub. data) and by Albers and Stewart (1965). Graptolites from this area correspond in age to those from the type area of the Palmetto Formation.

This facies of dark graptolitic shale, black chert, and cherty limestone extends to the northeast and is seen in the San Antonio Mountains north of Tonopah (Ferguson and Muller, 1949, p. 49) and at the Manhattan district (Ferguson, 1924) in the south end of the Toquina Range about 70 miles northeast of the Silver Peak Range. The eastern-most extent of the facies may be in the Monitor Range (F. J. Kleinhampl, oral commun., 1965), because to the east Ordovician rocks are represented by limestone of a different facies (fig. 11).

All the dark Ordovician argillites from this region are probably roughly correlative. Precise correlation on the formational level, however, will not be possible until details of the different rock sequences have been determined.

Ordovician rocks to the south, in the Inyo Range and Grapevine Mountains, are different in lithologic character from those of the Silver Peak Range, and no transitional rock types are known from intermediate areas (fig. 12). The Ordovician rocks of these two regions cannot now be correlated satisfactorily.

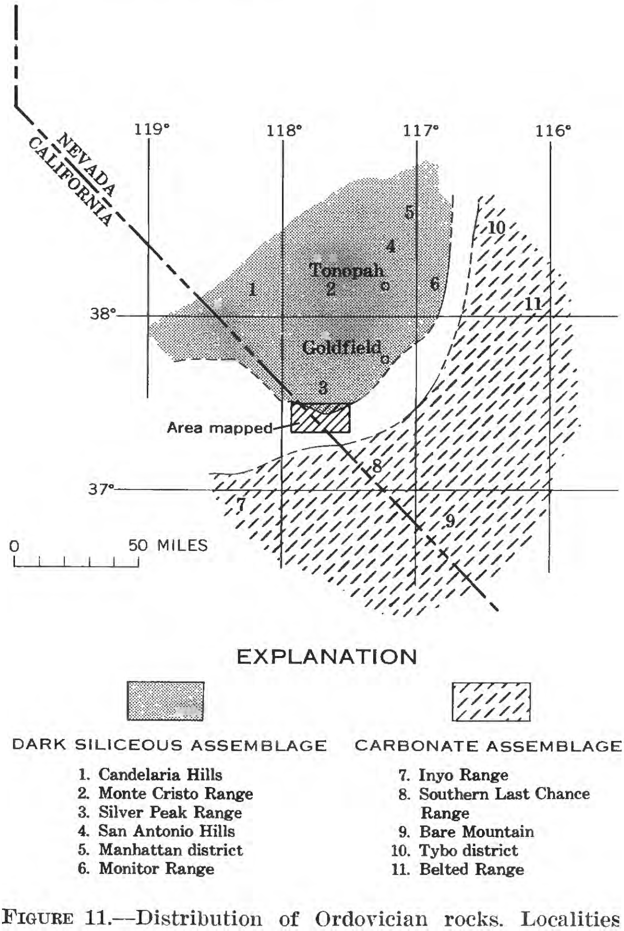

FIGURE 11.—Distribution of Ordovician rocks. Localities from Ross (1961).

The oldest Ordovician rocks in the area mapped are Early (?) and early Middle Ordovician, and the youngest Cambrian rocks are early (?) Late Cambrian. In adjacent areas of the Silver Peak Range, rocks of late Late Cambrian (Trempealeauan) age have been reported (J. P. Albers and J. H. Stewart, oral commun., 1965). Rocks mapped as the Palmetto Formation (J. P. Albers and J. H. Stewart, oral commun., 1965) overlie the uppermost Cambrian strata in one locality, but no Ordovician fossils have been found in these rocks. To date (1968), no unquestionably Lower Ordovician strata have been found anywhere in the region. In most places Middle Ordovician rocks are in fault contact with older rocks, and possibly all the Ordovician rocks are allochthonous.

CENOZOIC ROCKS

Tertiary rocks ranging in age from late Miocene through Pliocene and Quaternary rocks occur in the area mapped. They are predominantly volcanic flows, tuffs, tuffaceous silt and sandstone, and coarse pebble to boulder conglomerate (fig. 13). Similar rocks have been mapped by reconnaissance methods in numerous localities of southwestern Nevada, but detailed petrographic or stratigraphic studies of them have rarely been made.

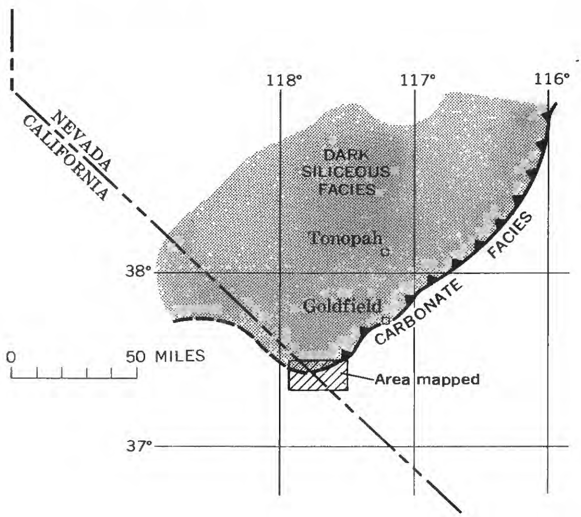

FIGURE 12.—Distribution of Paleozoic rocks after thrusting in Early Mississippian time. Modified from Roberts (1964).

ESMERALDA FORMATION

The name Esmeralda Formation was proposed by H. W. Turner (1900) for pyroclastic rocks and lacustrine sedimentary rocks that occupy parts of Clayton, Big Smoky, and Fish Lake Valleys north and west of the Magruder Mountain area.

The age of the Esmeralda Formation in its type area is Miocene and Pliocene based on evidence from fossil fish, leaves, and mollusks (Turner, 1900, p. 204). A more recent age assignment, based on flora, for part of the type section near Coaldale, Nev., is early Pliocene (Axelrod, 1940). Recent potassium-argon dates from units at or near the type locality range from 12.7 (Evernden and James, 1964, p. 970–971) to 6.9 million years (R. J. Moiola, oral commun., 1964), which agrees closely with the age assignment based on fossils. Because of the difficulty in correlating these upper Tertiary rocks from one outcrop to another, the term Esmeralda Formation has come to mean upper Miocene to Pliocene tuffaceous sedimentary rocks of various thicknesses, facies, and ages in Esmeralda County and adjacent parts of Nevada. It includes parts of such units as the Siebert Lake beds, described by Ransome (1909) from the Goldfield district, and the Siebert Tuff of the Tonopah district described by Spurr (1905).

In the extreme south end of Fish Lake Valley, tuffaceous sandstone, siltstone, and conglomerate with interbeds of basalt and rhyolitic tuff form a continuous section more than 2,400 feet thick (fig. 14). This section is assigned to the Esmeralda Formation. The sequence thins abruptly toward the bordering mountains to the northeast with facies

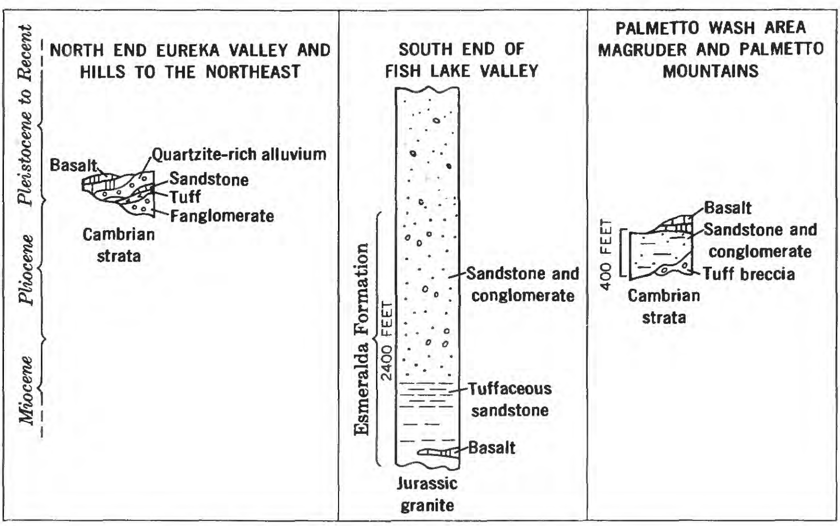

FIGURE 13.—Composite Cenozoic stratigraphic columns.

changes from siltstone to boulder conglomerate. Coarse detritus, which is distributed through the entire section, consists almost entirely of granitic material, including granite boulders or cobbles in the conglomerate lenses and orthoclase, quartz, and biotite grains in the sandstone units. The entire sequence in southern Fish Lake Valley dips away from Sylvania Mountain, a granitic body that borders the valley on the southeast and acted as the major source of detritus in the south end of the valley.

The upper three-fifths of the section in the south end of Fish Lake Valley consists largely of buff friable arkosic sandstone, which intertongues with granitic conglomerate. The upper part of the section seems to merge imperceptibly into the Recent alluvium in the valley

FIGURE 14.—Esmeralda Formation at south end of Fish Lake Valley. Total thickness is more than 2,400 feet. Lower 1,000 feet is tuffaceous sandstone and siltstone; upper 1,400 feet is arkosic sandstone and conglomerate.

bottom. The lower part of the section contains more pyroclastic rock and green siltstone and looks like the type Esmeralda Formation. A camel bone, identified as *Procamelus* (proximal phalanx) by S. D. Webb (1961), Department of Paleontology, University of California, Berkeley, was found near the base of the sequence and indicates a late Miocene to early Pliocene age, which agrees with the age of the type Esmeralda Formation.

In the axis of the valley, the arkosic tuffaceous sequence unconformably overlies approximately 100 feet of steeply dipping white to purple tuffaceous sandstone and basaltic agglomerate.

OLDER ALLUVIUM

Poorly consolidated fanglomerate dipping toward Eureka Valley is exposed along the southwest side of the hills northwest of the Last Chance Range approximately 2.5 miles southeast of Piper Mountain. The rock units are lenticular and range from a few feet to 150 feet in thickness. They are composed of poorly sorted rubble made up of angular fragments of all sizes from boulders to silt. Most of the silt is red. Detrital fragments large enough to identify have been derived from the Cambrian rocks exposed in the nearby hills. These fanglomerates are now being dissected and eroded away and their outcrops are incised by small canyons so that they form low ridges on the southwest side of the range. Their age is unknown, but is presumed to be late Tertiary, as indicated by their unconsolidated character and their stratigraphic position below tuffs containing partially devitrified glass. They also underlie basalt flows that have been elevated several hundred feet by Basin-Range faulting, which formed the present topography.

Rhyolite tuff, associated with the fanglomerates along the northeast edge of Eureka Valley, crops out in a few small areas. In most places it lies on top of the red fanglomerate and is capped by basalt flows or alluvium. The number and thickness of the tuff beds are variable, but in most places only one bed, approximately 20 feet thick, is present.

Thin lenticular beds of sandstone occur locally between or on top of the tuff beds. In contrast to the tuff, this sandstone is well stratified and shows horizontal, parallel layering, crossbedding, and local channeling.

Capping the peaks and ridges at the north end of Eureka Valley is alluvium composed exclusively of quartzite boulders and cobbles derived from the Cambrian Harkless Formation. In some places this unconsolidated sedimentary deposit, composed of angular rubble, lies directly on the basement Cambrian strata, and in other places it overlies the tuffaceous beds or fanglomerate described above. Basalt seems to cap the alluvium locally.

The age of this quartzite-rich alluvium is unknown, but the alluvium is older than the present physiographic stage of the range and is also probably older than the basalt. It is probably early Pleistocene.

The depression between the Sylvania and Palmetto Mountains is largely underlain by weakly consolidated tuffaceous sandstone and conglomerate. In general this is a broad, flat area; it covers approximately 30 square miles in the northern part of the Magruder Mountain quadrangle and extends southward to the ruggedly dissected mountain slopes at the north end of Death Valley. This physiographic surface is an old pediment—the friable tuffaceous strata beneath have been gently tilted, beveled, and covered by a thin veneer of more recent alluvium. Palmetto Wash and its tributaries are now actively dissecting this old surface (fig. 15), and in places they have cut as much as 300 feet into the underlying strata. Exposures along the stream valleys reveal a slight angular discordance between the alluvial veneer at the surface and the tuffaceous strata beneath. The two can be readily distinguished by composition, for the older strata contain detritus derived from the Palmetto Mountains exclusively, whereas the alluvial cover is a mixture of debris from local sources. The tuffaceous rocks unconformably overlie Precambrian and Paleozoic rocks in most of this area, but near Pigeon Springs they seem to overlie a rhyolite tuff-breccia. The age of the biotite in this tuff-breccia was determined

FIGURE 15.—Pediment dipping away from Palmetto Mountains. Weakly consolidated sedimentary deposits are several hundred feet thick under part of this surface.

(J. D. Obradovich, oral commun., 1961) by the potassium-argon method as $4.5 \pm 0.2 \times 10^6$ years, which suggests middle Pliocene. Therefore, the gently tilted tuffaceous sandstone and conglomerates and the pediment surface that bevels them must be post-middle Pliocene in age.

The tuffaceous sandstone and conglomerate, as well as the pediment surface, dips southwestward away from the Palmetto Mountains (fig. 15) and overlaps the northeast flank of the Sylvania Mountains. Detritus in the conglomerate is composed of rock fragments from distinctive Ordovician and Middle Cambrian formations found only in the Palmetto Mountains, and facies distribution of the conglomerate lenses bears out the inference that the Palmetto Mountains were the source of detritus. Semiconsolidated alluvium from this sequence of rocks forms a mantle on Cambrian strata at altitudes of 8,000 feet in the pass near Excelsior Spring, Palmetto Mountains. Small patches of this alluvium are also to be seen on the northeast side of this pass approximately 3 miles northeast of Excelsior Spring.

EXTRUSIVE IGNEOUS ROCKS

OLIVINE BASALT

Olivine basalt is exposed discontinuously over the hills at the north end of the Last Chance Range. It caps Piper Mountain, the highest peak in the western part of the area mapped, and forms a broad, flat surface at the top of the range 4 miles southeast of Piper Mountain. Smaller exposures are scattered along the north side of Eureka Valley, on the west end of Magruder Mountain, and on the north side of the Sylvania Mountains.

Basalt of the Last Chance Range is a single flow that presumably was once continuous over an area of approximately 10 square miles. It ranges in thickness from a few feet to about 60 feet, but in most places is no more than 20 feet thick. The basalt rests directly on Jurassic granitic rocks at Piper Mountain, on Cambrian sedimentary rocks and possibly on the high-level early quartzite-rich alluvium in the hills northwest of the Last Chance Range, and on rhyolitic tuff or fanglomerate along the north side of Eureka Valley. In the Sylvania Mountains and at Magruder Mountain, basalt lies on Precambrian and Cambrian strata, on granitic rocks, or on tuff-breccia dated (potassium-argon method) as middle Pliocene (J. D. Obradovich, oral commun., 1961).

INTRUSIVE IGNEOUS ROCKS

Several large granitic bodies lie within the area mapped, and numerous small intrusions of aplite, felsite, diorite, and andesite cut the Precambrian and Paleozoic sedimentary rocks.

The largest granitic bodies in the area belong to the southern part of the Inyo batholith, which has been studied by Anderson (1937a), Emerson (1966), and others. These rocks extend into the area from the northwest and form the peaks at the south end of the White Mountains. At the south end of Fish Lake Valley, the Sylvania Mountains, another large granitic complex, extend southeastward for 15 or more miles and terminate at the north end of Death Valley. The only other large granitic body in the area mapped is the Palmetto pluton, a small portion of which extends into the north-central part of the area. This body was described by H. W. Turner (1909).

Portions of the Inyo batholith in the western part of the Soldier Pass quadrangle are quartz monzonite and granodiorite. The main part of the Sylvania Mountains is quartz monzonite that is identical in appearance with the quartz monzonite in the western part of the Soldier Pass quadrangle. The southeastern part of the Sylvania granitic complex, at the north end of Death Valley and in Tule Canyon south of Magruder Mountain, is granite rather than quartz monzonite.

QUARTZ MONZONITE

At the south end of the White Mountains, the coarse-grained porphyritic quartz monzonite has been intruded into Lower Cambrian sedimentary rocks. The strata have been contorted and metamorphosed near the contact, but the contact is sharp. In one place, a long, narrow pendant of quartzite and phyllite can be traced for several miles into the plutonic body.

In hand specimen the rock is uniformly porphyritic with large (one-half inch) euhedral phenocrysts of pink orthoclase in a gray medium-grained hypidiomorphic-granular groundmass. The minerals of the groundmass are plagioclase, quartz, hornblende, biotite, and sphene, listed in decreasing order of abundance.

A potassium-argon date of 162 ± 4 million years was obtained by G. H. Curtis (oral commun., 1961) from biotite in a sample of the quartz monzonite collected approximately 2.5 miles north of Piper Mountain. This date corresponds to those obtained for the adjacent granodiorite and the quartz monzonite in the Sylvania Mountains, 10 miles to the southeast, and suggests a Middle Jurassic age.

GRANODIORITE

South of the quartz monzonite and separated from it by a zone of metamorphic rocks 100–300 feet wide is a fine- to medium-grained gray to white granodiorite. This rock is well exposed on Piper Mountain and southward along the spur of the Inyo Range that bounds the east side of Deep Spring Valley. Unlike the nearby quartz monzonite, this granodiorite is variable in both composition and texture. Areas of mafic enrichment occur within the granodiorite, and zones of dioritic rock containing as much as 50 percent hornblende occur locally along much of its contact with Cambrian sedimentary rocks. The contacts are sharp against contorted and thermally metamorphosed Precambrian and Cambrian country rock, as are those of the quartz monzonite. A few aplite and felsite dikes 10–100 feet thick cut across the sedimentary rocks near the contact.

Biotite from a sample of the granodiorite collected near Deep Spring School by G. H. Curtis (oral commun., 1961) was dated by him, using the potassium-argon method, as 170±5 million years old, or of Middle Jurassic age.

The Sylvania Mountains, which comprise approximately 35 square miles in the central and southeastern parts of the area mapped, are largely a complex plutonic body. Most of the pluton is a porphyritic quartz monzonite that looks, both in hand specimen and in thin section, like the quartz monzonite at the south end of the White Mountains (p. H27). Smaller amounts of medium-grained quartz monzonite and granodiorite crop out locally. In Tule Canyon, on the southeast edge of the mapped area, much leaching and alteration have destroyed the mafic minerals, but thin sections show that the rocks were once typical quartz monzonite. In some places near the north edge of the pluton, the quartz monzonite is a medium- to fine-grained nonporphyritic rock, but it is petrographically similar to the main porphyritic quartz monzonite.

A potassium-argon date of 155±4 million years was obtained by J. D. Obradovich (oral commun., 1961) for biotite in the main porphyritic quartz monzonite in the southern part of the Sylvania Mountains. This indicates a Middle Jurassic age and corresponds, within the limit of error, to the date obtained from the quartz monzonite in the southern part of the Inyo batholith. The Sylvania Mountain pluton should probably be regarded as part of the Inyo batholith.

A coarse-grained granitic rock crops out in the highest part of the Palmetto Mountains and extends over approximately 40 square miles along the southwest edge of the Silver Peak Range. This pluton, like the Sylvania Mountain pluton and Inyo batholiths, is composed of several types of granitic rock ranging from granite to granodiorite, but porphyritic quartz monzonite is most common. Only a small part of this pluton is within the mapped area; it consists of a dikelike offshoot (500 ft. wide) extending south from the main plutonic body.

DIORITE DIKE ROCKS

Numerous small dark intrusive bodies cut the Precambrian and Paleozoic sedimentary rocks in many parts of the area studied. Most of these occur as swarms of five to 10 dikes, each 100–500 feet long. The individual dikes have no obvious preferred trend or pattern and some bodies are nearly equidimensional in outcrop plan.

Andesite and fine-grained hornblende diorite are the major rock types comprising these dark dikes. The average composition is approximately 60–70 percent plagioclase, 10–20 percent hornblende, 5–10 percent orthoclase, and 0–10 percent magnetite. Quartz, apatite, and biotite are accessories.

FELSITE DIKE ROCKS

Aphanitic light-colored dikes, most of which are altered, crop out in the area mapped, but they are less abundant than the dark andesite and diorite dikes described above. A swarm of felsite dikes occurs on the north side of Palmetto Mountain, and numerous similar dikes occur in the adjoining area to the northeast. On Magruder Mountain and in the hills northwest of the Last Chance Range, light-colored dikes are rare.

METAMORPHIC ROCKS

Two types of metamorphism can be distinguished in the area mapped: (1) contact metamorphism related to the plutonic intrusions, and (2) low-grade regional metamorphism not apparently associated with intrusive bodies.

Almost all the Precambrian and Cambrian sedimentary rocks show some degree of metamorphism. Most siltstone is weakly foliated, and the argillaceous material is largely converted to very fine grained biotite and chlorite; foliation completely obscures bedding locally. Carbonate rock is commonly recrystallized, and many fossils in limestone have been distorted beyond recognition or destroyed. Sandstone units are dense quartzite containing completely interlocking and intergrown grains and a small amount of recrystallized biotite and chlorite in the matrix. All the quartzite has well-developed joints normal to bedding so that it breaks into rectangular blocks.

No definite pattern is apparent in the distribution of low-grade regionally metamorphosed rocks. Slightly altered sedimentary rock grades along strike into slate and phyllite, then back into apparently unaltered rock. Such a relationship suggests that temperatures and pressures were only locally strong enough to produce foliation. Bedding is the dominant planar feature in the rock throughout the region, and, because of this, the rocks can be mapped according to sedimentary-stratigraphic methods. No attempt has been made to analyze the metamorphic structure or fabric.

Spotted hornfels, marble, and schist are developed at or near the contacts of all the large plutonic bodies. The widest contact aureole is adjacent to the Sylvania Mountain quartz monzonite near Log Cabin Spring. Here, metamorphic rocks extend for approximately a mile from the granitic outcrops. In other places the contact aureole is not as wide. The contact metamorphic zone around the south end of the Inyo batholith is less than a quarter of a mile wide, as is the zone of hornfelses and skarn south of the Palmetto Mountain pluton. The mineralogy, fabric, and structure of the metamorphic rocks is also somewhat different around all the plutonic bodies.

Metamorphic rocks south of the Inyo batholith are highly distorted by both folding and faulting, and the pelitic rocks have a schistose texture. Limestone was apparently the first rock type to become metamorphosed, for it is recrystallized to a coarse-grained marble at distances of approximately one-fourth mile from the granitic rock. Closer to the pluton, pelitic rocks have been altered to hornfelses that are increasingly schistose toward the pluton. However, rocks immediately adjacent to the pluton are hornfelses. Quartzose rocks show little or no sign of metamorphism except where concentrations of coarse biotite flakes occur on some of the bedding surfaces. Schists and highly deformed calc-silicate hornfelses crop out in some places near the granitic rock, and limestone units thicken or thin disproportionally.

Samples of all the principal rock types were collected across the zone of contact metamorphism. Mineralogically these rocks were restricted to two facies of contact metamorphism—the albite-epidote hornfels facies and the hornblende hornfels facies (Fyfe and others, 1958, p. 203–205).

Mineral assemblages, according to rock type, are:

Pelitic	*Calcareous*	*Quartzose*
Muscovite-biotite-quartz-andalusite	Calcite-talc	Quartz-biotite-muscovite-andalusite
	Calcite-quartz-talc	
Muscovite-biotite-quartz	Calcite-dolomite-forsterite	Quartz-albite-biotite-muscovite
	Quartz-calcite-diopside	Albite-epidote-quartz

Metamorphic rocks along the north edge of the Sylvania Mountain pluton are poorly exposed, but where found they are typically hornfelses. Most of the rocks in contact with the pluton are types that do not tend to show much metamorphism. The massive coarse-grained Reed Dolomite, where it is in contact with the granitic rocks, looks the same as elsewhere. Hornfelses and dolomites of the Wyman Formation also show little difference in character near the pluton. A few small isolated pendants of metamorphic rock occur within the Sylvania Mountain pluton. This rock is mostly garnet-epidote-diopside skarn, and its outer 3–4 feet is a zone of complex granitic injection. The mineral assemblages of the carbonate and quartzo-feldspathic rocks is typical of the albite-epidote hornfels facies (Fyfe and others, 1958, p. 203). A summary of their mineralogy is:

Pelitic and quartzo-felspathic hornfelses	*Limestone or marble*	*Skarn*
Quartz-biotite-muscovite	Calcite-quartz	Garnet-epidote-diopside-microcline-plagioclase
Quartz-albite-biotite-epidote		
Quartz-albite-andalusite (altered to sericite)		Garnet-epidote-calcite-diopside-actinolite(?)

Dark shale, chert, and limestone of the Palmetto Formation are in contact with the Palmetto Mountain pluton. In most localities where this formation is not near granitic rocks, it is greatly deformed; no additional amount of deformation is apparent in the contact aureole. Outcrops consist of a rubble of small rectangular blocks of black chert interfingered with dark siliceous siltstone. Within about a quarter of a mile of the pluton, zones of pure garnet skarn or garnet epidote skarn occur. Pods several feet wide, which contain more than 25 percent hematite are exposed locally. Near the granitic contact, siltstone and shale are less abundant and chert is more abundant, which suggests silicification. Local zones of gnarled black chert and white quartzite are common in such places. No rocks other than silicified types are within approximately 100 yards of the pluton.

The great amount of silicic and iron-rich rock associated with the Palmetto Mountain pluton contrasts with rock of the contact zones north of the Sylvania Mountain pluton and south of the Inyo batholith. In the latter two zones obvious metasomatic addition is very slight. Rocks around the Palmetto Mountain pluton lack the metamorphic fabric, either hornfelsic or schistose, that is typical of the other two contact aureoles.

STRUCTURAL GEOLOGY

The structure is believed to have developed from at least two episodes of deformation. An earlier episode or episodes is recorded by folds, thrusts, and high-angle faults involving the Precambrian, Cambrian, and Ordovician rocks, and a later period of deformation is indicated by the Tertiary to Recent block faults of Basin and Range type which have displaced rocks of all ages in the mapped area and are responsible for the existing topography.

The pre-Basin and Range structure differs greatly in style from north to south across the area. The Palmetto Mountains in the northeast are characterized by thrust faults, recumbent folds, and steeply dipping faults that form a continuation of the pattern in the adjoining Silver Peak Range. This structural complexity contrasts strongly with the structure of the Precambrian and Paleozoic rocks in the hills northwest of the Last Chance Range and in Magruder Mountain, which is characterized by open folds and numerous small steeply dipping faults. Similar structure is also typical of the Precambrian and Cambrian strata of the White and Inyo Mountains to the west. Thus, the region becomes more complex structurally from south to north (fig. 16).

The structural geology of individual mountains within the area is described separately, as each mountain presents different and contrasting characteristics. These subareas, from southwest to northeast, are (1) the hills at the southwest end of Fish Lake Valley; (2) the Sylvania Mountains; (3) Magruder Mountain; and (4) the Palmetto Mountains. The Death Valley–Furnace Creek fault zone, a conspicuous linear structural feature in eastern Califronia, which passes diagonally across the southwest corner of the area between the Sylvania Mountains on the north and the Last Chance Range on the south, is considered separately.

HILLS AT THE SOUTHWEST END OF FISH LAKE VALLEY

The oldest structures in the hills at the south end of Fish Lake Valley, which are a northern extension of the Last Chance Range, are several open northwest-trending anticlines and synclines in Precambrian and Cambrian rocks. These have been displaced along numerous steeply dipping faults, most of which trend east-west and can be traced for distances of less than a mile. These faults appear to be older than, and unrelated to, the Basin and Range faults that outlined the present range. Most of these older faults show displacements of 200 feet or less, but one, approximately in the center of the range, is a large reverse fault (or thrust) that brought Middle Cambrian limestone on the south against, or on top of, Lower Cambrian quartzite and siltstone to the north. The fault surface is very irregular; its dip generally ranges from 70° to 20°. The fault contact is marked by a zone several hundred feet wide of intensely wrinkled and isoclinally folded limestone, but the Lower Cambrian quartzite north of the fault is not severely deformed. Possibly the entire thickness of Middle and Upper Cambrian limestone is allochthonous.

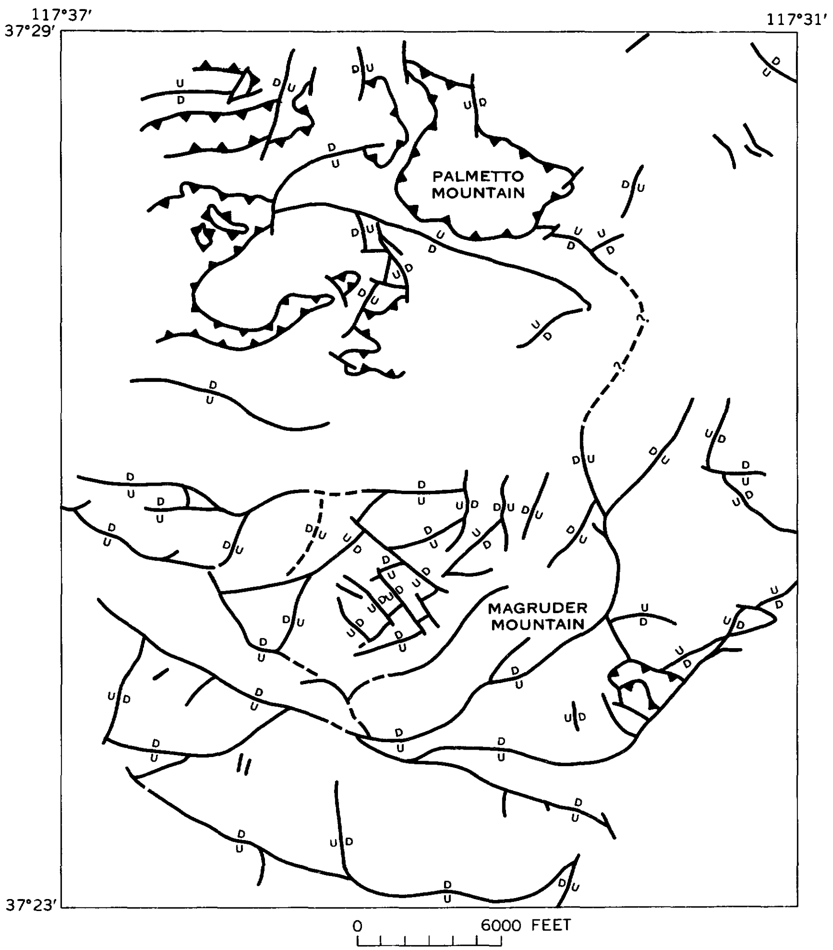

FIGURE 16.—Faults in the eastern part of the Magruder Mountain area. The Palmetto Mountains are characterized by thrust faults cut by steeply dipping faults; Magruder Mountain is characterized by steeply dipping faults.

SYLVANIA MOUNTAINS

The Sylvania Mountains, like the other mountains in the region, resulted from Basin and Range faulting. They are separated from the Last Chance Range on the south by the Death Valley–Furnace Creek fault zone, and they are probably bounded on the north side also by a fault zone now obscured by the Palmetto Wash pediment. Alteration and shearing within the granitic rocks along the north side of the Sylvania Mountains are the only evidence of faulting. Faults are difficult to recognize in most parts of the Sylvania Mountains, and the displacement along them is uncertain because of the homogeneous nature of the granitic rock. Zones of alteration and leaching within the pluton are common, but they may not indicate displacements.

The Sylvania pluton is of Middle Jurassic age and is in contact with thermally metamorphosed Precambrian and Cambrian sedimentary rocks. Most of the faults in the Sylvania Mountains postdate the intrusion, and most of the recognizable structure is Tertiary to Recent Basin and Range structure. One fault, however, seems to be unrelated to the block faulting described above. This fault, the Palmetto-Oasis thrust fault (Page and Wright, 1942, p. 3), is at the northwest end of the mountains near the Sierra talc mine. It is a reverse fault that dips 15°–25° NE and brings granitic rock over the Precambrian Reed Dolomite. The lack of physiographic expression and the low angle of the fault suggest that this fault is older than the steep Basin and Range faults that flank the range. Along the fault a thin zone of talc has formed both in the dolomite and for several inches into the granite. The altered granitic rock retains the coarsely porphyritic texture of the unaltered rock, but it now consists of quartz and altered orthoclase in a matrix of chlorite, talc, and fine-grained quartz. The talc mineralization seems to be younger than the faulting and probably resulted from percolation of hydrothermal solutions along the fault plane.

MAGRUDER MOUNTAIN

Magruder Mountain is a highly faulted homocline that dips to the southeast. A few small open folds form wrinkles in the otherwise uniformly dipping strata. Many steep normal and reverse faults, having displacements mostly less than 500 feet and trends of N. 55° W. or N. 30° E., offset Cambrian rocks. A large low-angle reverse fault that dips approximately 25° S. trends along the entire south edge of the range. This fault has a stratigraphic separation of several thousand feet. Another steeply dipping fault having large stratigraphic separation lies between Magruder Mountain and the Palmetto Mountains and is covered by Tertiary rhyolite tuff and alluvium.

The Basin and Range faults that account for the topographic expression of Magruder Mountain are mostly covered by Recent alluvium. Several patches of weakly consolidated older alluvium in Tule Canyon south of O'Hara Spring have been tilted as much as 50° S., presumably along a Tertiary to Recent Basin and Range fault. This fault is delineated by an east-west line of springs for approximately 10 miles; named springs along this line are Wild Rose, Walker, Stockage, and Log Cabin Springs.

Evidence of Basin and Range faulting along the north side of Magruder Mountain is not obvious. A linear topographic depression between Magruder Mountain and the Palmetto Mountains is suggestive of a fault zone, but it does not constitute conclusive evidence.

PALMETTO MOUNTAINS

The complex pre-Basin and Range structure of the Palmetto Mountains contrasts strongly with the simpler structures of Magruder Mountain to the south and with those in the hills at the north end of the Last Chance Range to the west (fig. 16). The Palmetto Mountains consist of a mosaic of high- and low-angle faults of variable stratigraphic separation. In the entire 50-square-mile area of these mountains, the writer found only two contacts that are sedimentary—all others are fault contacts. Major folds are obscure because of the small size of most faulted segments, but local zones of intense small-scale folding are common and are presumed to be parts of larger folds. A possible explanation for this maze of faults is that several large thrust plates, one of Middle and Upper Cambrian limestone and another of lower (?) and Middle Ordovician shale, overrode the area and were later offset by steeply dipping cross faults. Only small displaced segments of the original thrust plates are now represented by the apparently concordant contacts.

Evidence that the Palmetto Mountains were elevated along late Tertiary to Recent faults is lacking in the area mapped. Only the southern slope of these mountains lies within the area, and this is a gentle slope rising gradually from Palmetto Wash. It forms a part of the Pliocene and Pleistocene pediment and it does not appear to have been displaced by recent faulting along the Palmetto Mountain front. The rounded, gentle profile of the Palmetto Mountains and the lack of recent faults suggest that this part of the Silver Peak Range was a high on an older surface that existed prior to the culminating episode of Basin and Range faulting, which elevated most of the mountains in the region.

DEATH VALLEY–FURNACE CREEK FAULT ZONE

The Death Valley–Furnace Creek fault zone extends into the southwestern part of the area mapped. It parallels the California-Nevada State boundary for more than 50 miles and is visible from Death Valley in the south to Fish Lake Valley in the north (fig. 17; pl. 1). Along the northeast edge of the Last Chance Range and the south edge of the Sylvania Mountains, the fault zone is well exposed. Here, it is about half a mile wide, and the faulting has brought granitic rock of the Sylvania Mountains against Precambrian and Cambrian sedimentary rocks. All rocks within the zone are granulated, sheared, and leached to such an extent that fresh, coherent specimens are difficult to collect. Where the fault zone is covered by alluvium along the west side of Fish Lake Valley, recent movement along one of the faults has formed a small scarp.

The fault zone must be at least as old as late Miocene. Sedimentary rocks of the Esmeralda Formation (here with a maximum potassium-

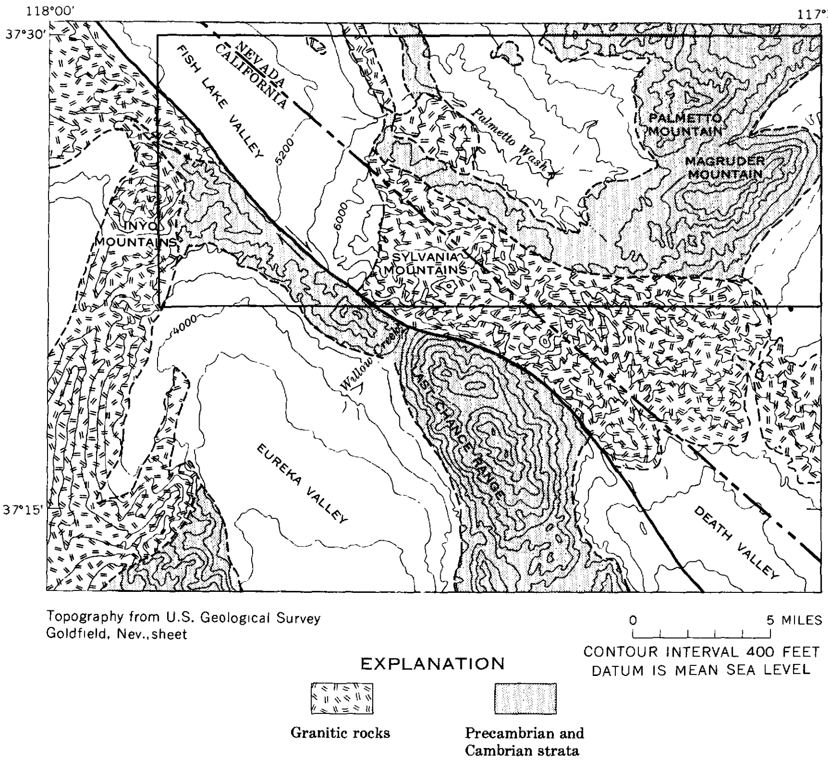

FIGURE 17.—Possible displacement in the Death Valley–Furnace Creek fault zone.

argon age of 13.1 m.y.) fill the south end of the Fish Lake Valley graben, which indicates that this structural trough existed at the time of deposition. Moreover, the facies patterns and the thickness of the Esmeralda Formation within the trough indicate that the Esmeralda deposits in this area were largely confined to this basin.

The amount of movement along the Death Valley–Furnace Creek fault zone is difficult to determine. Displacements have been primarily right-lateral strike slip, although several hundred feet or more of vertical displacement is indicated by the uplift of the Last Chance Range. In the area mapped it is difficult to match features across the fault to determine the amount of strike-slip movement, but the distinctive quartz monzonites at the south end of the White and Inyo Mountains (Inyo batholith) and Sylvania Mountains, which look identical and are the same age (approximately 160 m.y. by potassium-argon dating), may well be offset parts of the same body. If these are displaced segments of the same pluton, they indicate separation of more than 20 miles in a right-lateral sense (fig. 17).

SUMMARY OF STRUCTURE

Two, and probably more, episodes of deformation are documented in the rocks of the Magruder Mountain area. The first episode deformed Precambrian and early Paleozoic rocks and probably included both low- and high-angle faulting. The second episode is documented by high-angle faults of Tertiary to Recent age that gave rise to the present topography. The age of the older structures is uncertain, but it seems likely that one or more of the orogenies of regional extent, recognized in areas to the north, probably also affected the Magruder Mountain area or parts of it. These include the Late Devonian to Early Mississippian Antler orogeny, the Middle or Late Permian Senoma orogeny, and the "Jurassic and Cretaceous orogeny" (Silberling and Roberts, 1962).

The tectonic pattern in the northern part of the area and in regions farther north is a complex mosaic of thrust faults and recumbent folds cut by steeply dipping faults. The rocks involved in the thrusting are thin-bedded dark shale, limestone, and quartzite of the Early(?) and Middle Ordovician Palmetto Formation, which is in fault contact with various Precambrian and Cambrian sandstone, siltstone, and limestone units. It seems probable that the Antler orogenic belt extends into this region as suggested by Roberts (1964, p. 26) and that the outcrops of the Palmetto Formation transitional assemblage) may be klippen of the Roberts thrust. South of the Palmetto Mountains—at Magruder Mountain, the hills north of the Last Chance Range, and westward in the White and Inyo Mountains—the tectonic pattern is characterized by steeply dipping faults and open folds (fig. 16), and the dark graptolitic Ordovician shales are missing. The nearest Ordovician rocks to the south and west, limestone of the Pogonip Group, are in Eureka Valley approximately 20 miles from Magruder Mountain.

The marked difference in tectonic pattern and the proximity of contrasting Ordovician facies strongly suggest that all the Ordovician rocks in the northern part of the Magruder Mountain area are allochthonous, and support Roberts' suggestion that the Ordovician rocks were brought in from the west and northwest. No evidence of the Antler orogeny is found to the south or southwest of the Palmetto Mountains.

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