Geology of the Northern Part of the Toquima Range, Lander, Eureka, and Nye Counties, Nevada

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By EDWIN H. McKEE

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GEOLOGY OF THE NORTHERN PART OF THE TOQUIMA RANGE, LANDER, EUREKA, AND NYE COUNTIES, NEVADA

By Edwin H. McKee

ABSTRACT

Rocks exposed in the Toquima Range of central Nevada (centering at lat 39°20′ N. and long 117°50′ W.) are early and middle Paleozoic marine strata, middle and late Mesozoic intrusive igneous rocks, and late Cenozoic volcanic and sedimentary rocks.

The Paleozoic rocks are divided into three assemblages: an eastern, predominately carbonate assemblage; a western, siliceous assemblage consisting mostly of chert and dark shale; and an overlapping assemblage of limestone and conglomerate.

The carbonate assemblage is composed of four sequences that have been brought together on thrust faults with displacements of 10 miles (16 km) or more. Each sequence contains a generally similar yet significantly different series of strata that were deposited across a wide region of central Nevada. The four distinct stratigraphic sequences of correlative carbonate rocks are exposed at localities only a few kilometres apart in the northern Toquima Range. Three of the sequences are stacked in a series of thrust plates in the Ikes Canyon window in the southern part of the area; the fourth is exposed in Petes Canyon window about 12 miles (19 km) to the north. The Ikes Canyon window contains, from lowest to highest, the August Canyon, Mill Canyon, and June Canyon sequences. The August Canyon sequence is presumed to be autochthonous; the others, allochthonous. The Petes Canyon window contains the Petes Canyon sequence. Most of the units of the carbonate assemblage can be correlated with the well-known formations from the Eureka–Antelope Valley region of east-central Nevada, and many of the names used for these Ordovician, Silurian, and Lower Devonian rocks are derived from the classic sections of the Eureka–Antelope Valley region to the east. These formations include the Goodwin Limestone, Ninemile Formation, Antelope Valley Limestone, Hanson Creek Formation, and Roberts Mountains Formation. Formations defined from localities in the Toquima Range are the Tor and McMonnigal Limestones and the Caesar Canyon Limestone, the Gatecliff Formation, and the Masket Shale of Kay and Crawford (1964).

The western, siliceous assemblage, composed of Ordovician rocks, is called the Vinini Formation and is similar to allochthonous silicic rocks in other parts of central Nevada. These strate have been thrust 40 miles (64 km) or more eastward from their site of deposition on the regionally extensive Roberts Mountains thrust.

The overlapping assemblage of conglomerate and limestone lies with angular unconformity on both the older Paleozoic siliceous and carbonate assemblages and is presumed to represent detritus that accumulated near the edges of the allochthonous thrust plates. Kay and Crawford's (1964) name Wildcat Peak Fromation is adopted for rocks of the overlapping assemblage in the Toquima Range. These rocks are of Pennsylvanian age.

The Paleozoic history of the region can be divided into three phases. The first is recorded by strata of Early Ordovician to Early Devonian age deposited in deep water in regions west of central Nevada and in shallow seas in central Nevada and eastward. The Toquima Range lies in the shallow shelf zone near the transition into deeper water. The line marking the abrupt change in slope from shallow to deeper water probably extended northeastward across Nevada and passed 15–25 miles (24–40 km) west of the Toquima Range. Carbonate rocks were deposited east of this line. On the slope between shallow and deep water west of the line, little or no sediment accumulated. Siliceous strata were deposited in deep basins to the west. This general configuration of shelf and ocean basin and the distribution of sedimentary deposits existed with only minor differences until at least the start of Middle Devonian time.

The second major phase in the history of the region is thrust faulting that brought strata deposited west of the Toquima Range into the region and is responsible for the series of stacked thrust plates of carbonate and siliceous rocks in the central part of the range. Inception of thrusting in the west may have coincided with the waning stages of deposition in the eastern shelf region. Thrust faulting in the Toquima Range occurred between Middle Devonian and Early Pennsylvanian time. This age bracket has a possible lower limit set by the age of the youngest rocks cut by thrusting and an upper limit set by the age of rocks that overlap the thrust faults.

The rocks that serve as the upper limit at the time of thrust faulting also record the third major chapter of the Paleozoic history. Pennsylvanian conglomerate and limestone were deposited with marked angular unconformity on both allochthonous and autochthonous lower Paleozoic strata. These Pennsylvanian rocks indicate a return to shallow marine conditions accompanied by accumulation of coarse conglomeratic debris from the tectonically active regions nearby.

Quartz monzonite and granodiorite that yield potassium-argon ages of about 155 m.y. (million years) form plutons in the northern part of the Toquima Range. These ages suggest that the bodies are of Middle Jurassic age, about the same age as other plutons in central Nevada, including the Austin pluton at Austin, Nev. It seems probable that the small plutonic bodies in the northern Toquima Range exposed south of Clipper Gap Canyon and near Spencer Hot Springs are parts of the same large plutonic complex as the Austin pluton of the Toiyabe Range a few miles to the west.

Tertiary rocks in the northern part of the Toquima Range are mostly ash-flow sheets of regional extent. A few flows and sedimentary rocks are of local distribution. The oldest formation, the Northumberland Tuff, has a potassium-argon age of about 32 m.y. and was erupted from a volcanic center on the southern edge and a short distance south of the area. The volcanic center collapsed to form a caldera in which enormous landslide blocks and water-laid strata were deposited near the caldera rim and welded tuff was deposited in the central area. Rocks in this filled caldera are overlapped by a regionally widespread ash-flow sheet here called the tuff of Hoodoo Canyon, dated by potassium-argon at about 30 m.y. A second volcanic center in the northern Toquima Range erupted a local but thick pile of tuff and tuff-breccia informally called tuff of Stoneberger Canyon and dated at about 31 m.y. There is no evidence

of collapse of this volcanic center after eruption—the pyroclastic material simply formed a thick pile of weakly welded to nonwelded debris. Other ash-flow sheets of regional extent are the Pancake Summit Tuff, dated at about 31 m.y.; the Bates Mountain Tuff, comprising three cooling units all about 24 m.y. old; and the tuff of Clipper Gap, which at about 22 m.y. old is the youngest volcanic formation in the area. All the ash flows except the tuff of Hoodoo Canyon are of rhyolitic composition; the tuff of Hoodoo Canyon is quartz latite. The older units including the tuff of Hoodoo Canyon, tuff of Stoneberger Canyon, Pancake Summit Tuff, and Northumberland Tuff are rich in crystals and contain as much as 40 percent phenocrysts; the younger units including the Bates Mountain Tuff and the tuff of Clipper Gap are poor in crystals and contain 10 percent phenocrysts or less. The thickness of the ash flows and the stratigraphic sequence they form vary within the area.

INTRODUCTION

SCOPE AND PURPOSE

Geologic work for this report began in 1966 during a reconnaissance geologic study of Lander County, Nev. Mapping was completed the following year in the Spencer Hot Springs quadrangle (McKee, 1968b), which includes the northern part of the area of this study. During 1968 through 1971 the mapping was extended east to include part of the Hickison Summit quadrangle and to include the Wildcat Peak and the western part of the Dianas Punch Bowl quadrangles.

The study focuses on (1) the stratigraphy and structure of middle Paleozoic rocks in the northern part of the Toquima Range and their relations to a general paleotectonic reconstruction of middle Paleozoic geology of central Nevada and (2) the stratigraphy and distribution of Tertiary volcanic rocks, including rock chemistry and radiometric age. Study of the Tertiary units establishes correlation of widespread ash-flow sheets recognized elsewhere in Nevada and defines eruptive centers within the region.

LOCATION AND ACCESSIBILITY

The Toquima Range trends north-northeast in central Nevada from southeast of Austin to the vicinity of Tonopah (fig. 1). Included in the area of this study is the northern third of the range between lat 39° and 39°20′, long 117° and 116°40′. This area of about 430 square miles (1,100 km^2), mapped at a scale of 1:62,500 (pl. 1) includes all of one and parts of three 15-minute quadrangles outlined in figure 2.

There are no paved roads within the area, although U.S. 50 passes a few miles to the north and Nevada State Route 8A parallels its west edge. A graded dirt road, State Route 82, runs along the east edge of the Toquima Range and is joined by a graded dirt road that crosses the range at Petes Summit and joins 8A to the west. A second graded dirt road crosses the range at Northumberland Canyon about 15 miles (24 km) south

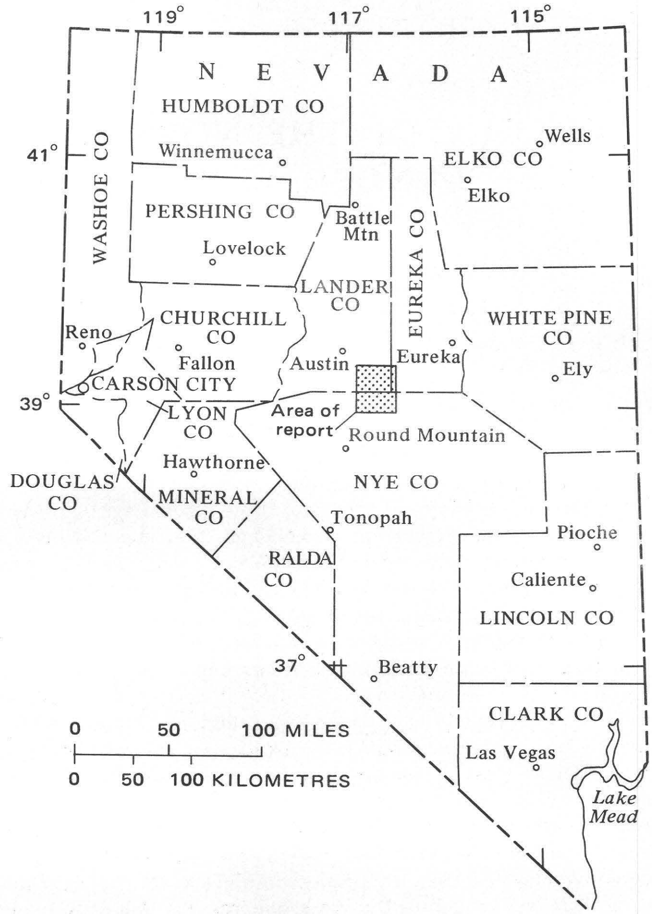
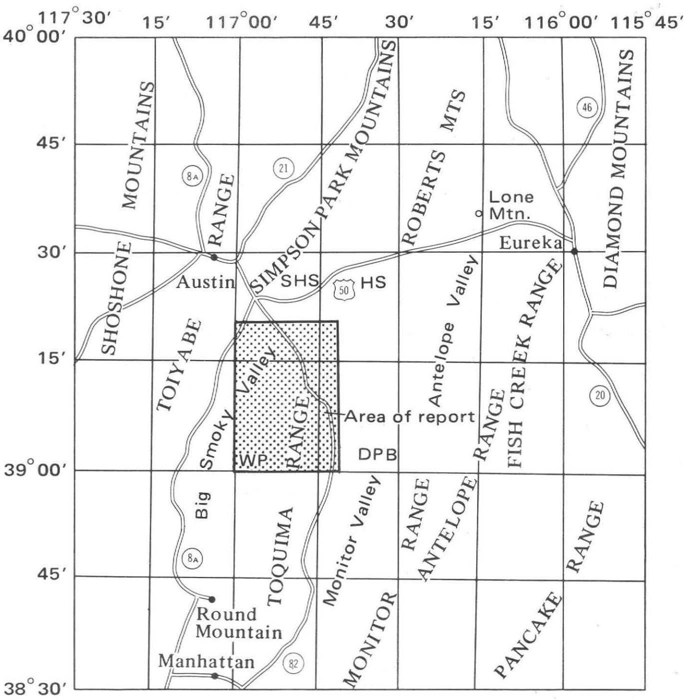


FIGURE 1.—Index map showing location of the area studied for this report.

of Petes Summit. Several minor roads, accessible by truck, reach all major canyons on both flanks of the range, and one road follows most of the crest of the northern part of the range. Access by vehicle is good to most parts of the area mapped.

PHYSICAL FEATURES

The Toquima Range and flanking valleys are typical physiographic features of the Basin and Range province. The range is narrow, linear, and north trending and is bounded by nearly flat alluvial valleys of comparable width. The mountain rises abruptly from the Great Smoky Valley on the west and Monitor Valley on the east along fault line scarps. Alluvial fans spreading from canyon mouths coalesce at valley edges; playas occur in the central parts of the valleys. The highest point in the northern part of the Toquima Range is Wildcat Peak (elev. 10,507 ft; 3,203 m); relief between this peak and Monitor Valley to the east is about 3,000 feet (900 m). Most of the highland of the northern part of the range, however, is a nearly flat surface several kilometers wide and 1,000–2,000 feet



SHS – Spencer Hot Springs quadrangle
HS – Hickinson Summit quadrangle
WP – Wildcat Peak quadrangle
DPB – Dianas Punch Bowl quadrangle

FIGURE 2.—Map of central Nevada showing the 15-minute quadrangles included in the geologic map (pl. 1) and the area discussed in this report.

(300–600 m) lower than Wildcat Peak. This tableland is developed on nearly horizontal Tertiary volcanic rocks. Elsewhere, the range is not capped by the flat-lying Tertiary units and instead consists of narrow ridges, spurs, and peaks.

CLIMATE AND VEGETATION

The climate of the Toquima Range as in most of the Great Basin is arid. The nearest point where annual rainfall has been recorded is Austin, Nev., about 30 miles (48 km) northwest at an elevation of about 6,800 feet (2,073 m). Here the mean annual precipitation is 11.83 inches (30.05 cm). Valley floors at elevations between 5,500 and 6,500 (1,676 and 1,981 m) probably receive about half this amount. The top of the Toquima Range probably receives 12–15 inches (30–38 cm) of rainfall annually. Annual evaporation is greater than precipitation so that in most years standing water disappears from the playas by midsummer. There are only a few small springs and two small streams in the area; in dry years some of these disappear by late summer. Temperatures are moderate in summer, seldom reaching 100°F even in the valleys and about 20° less within the range. Winter temperatures drop below freezing for extended periods, occasionally below 0°F.

Vegetation is typical of the Upper Sonoran and Transition zone and varies from sagebrush, rabbit brush, and sparse grasses at low elevations to piñon, pine, juniper, and mountain mahogany on mountain slopes. At the highest elevations occur stands of limber pine. Willows are common near springs or in valleys with high water tables; aspen groves occupy the watered canyons at higher elevations or cover north slopes where snow accumulates.

ARCHEOLOGICAL RECORD

The archeological record in the northern part of the Toquima Range reveals a long period of occupation by small groups of Indians living a migratory life controlled mostly by seasonal variations in rainfall and temperature. These mobile hunting-gathering people shifted from the valley regions to the mountains depending on the availability of game or edible plants, especially the piñon nut on which the economy was based. The hunt was probably the controlling event in the existence of these people and at times brought them together for animal drives. Evidence based partly on petroglyphs at scattered localities in the Great Basin suggests that at the large hunts petroglyphs were drawn at a certain site by a shaman as part of the hunting ritual.

Three petroglyph sites are known in the northern part of the Toquima Range within the area of this study: the Petes Summit site described by Heizer and Baumhoff (1962), the Gatecliff site in Mill Canyon presently being studied, and the Jean Spring site discovered during the course of this study and as yet undescribed. Aside from the three sites each marked by numerous petroglyphs, a number of individual slabs of limestone with scratched glyphs have been found in the area and are described by McKee and Thomas (1972). The exact significance of these inscribed rocks is not known but is presumed to be related to hunting magic.

HISTORY OF INVESTIGATION

Until recently the geology of the northern part of the Toquima Range remained unmapped and little known. Nearby regions favored by mining camps and rich mineral deposits understandably drew the attention of most early-day geologists; by the 1920's parts of nearly all of central Nevada, with the exception of the northern Toquima Range, had been described in mining surveys and reports.

In 1927 an Ordovician fauna collected near Ikes Canyon, in the northern part of the Toquima Range, was described by R. C. Bassler (1927; see also R. S. Bassler, 1941); in subsequent years many geologists visited this canyon to collect fossils or study the stratigraphic relations of the Paleozoic rocks. A brief description of the stratigraphy at the mouth of Ikes Canyon by Kirk (1933) focused on the Pogonip Limestone and overlying Silurian rocks, noting the absence of the Eureka Quartzite. Collections of brachiopods, many of which figure prominently in Ulrich and Cooper's (1936, 1938) and Cooper's (1956) faunal lists of the Lower and Middle Ordovician, were collected from Ikes Canyon. Fossil collections made later by Kay (1962) were from measured stratigraphic sections compiled during the course of geologic mapping (Kay and Crawford, 1964). In 1970 Ross published a comprehensive study of the Ordovician stratigraphy of eastern and central Nevada, including a discussion of the Ordovician rocks near the mouth of Ikes Canyon from which he collected and described a large fauna. A second and different section of Ordovician rocks in Ikes Canyon is described by McKee, Ross, and Norford (1972). For students of the Ordovician System in the Basin and Range province, the area around Ikes Canyon has become a classic locality.

The first geologic mapping in the northern part of the Toquima Range was conducted during the 1950's by students at Columbia University under the direction of Professor Marshall Kay. As a result of this work, Kay (1952, 1957) recognized the structural juxtaposition of siliceous and carbonate facies of lower and middle Paleozoic rocks, a relationship previously described elsewhere in central Nevada but unrecognized in the Toquima Range. In 1964 a more comprehensive analysis by Kay and Crawford of the structural geology and stratigraphy showed that in the vicinity of Ikes Canyon lower and middle Paleozoic carbonate rocks that were once widely separated are juxtaposed in three thrust slices. An additional sequence of carbonate rocks was found by McKee and Ross (1969) about 15 miles (24 km) north of Ikes Canyon.

GEOLOGIC SETTING

The lower and middle Paleozoic carbonate rocks from the northern part of the Toquima Range were deposited near the lateral transition from miogeosyncline (shelf) on the east to eugeosyncline (deep basin) on the west. Because these carbonate rocks were deposited near this transition zone, they occupy a key position with regard to reconstruction of the lower and middle Paleozoic paleotectonic pattern of the Great Basin. Formations that show little facies change across wide regions east of the Toquima Range here exhibit marked changes within very short distances. In addition to the lower and middle Paleozoic strata deposited in the region of the Toquima Range, at least three tectonic plates containing rocks of the same age have been superimposed by thrusting here. When the thrust plates are unraveled, tectonically telescoped stratigraphy representing deposition across areas tens of kilometres wide is revealed.

Tertiary volcanic rocks that cap much of the northern part of the Toquima Range consist of ash-flow sheets that can be correlated across a large region of central Nevada. These rocks were erupted before the present ranges were blocked out by Basin and Range faulting and occupy perched and isolated positions high in the ranges. In addition to these widespread formations, local volcanic rocks can be traced to several eruptive centers within the range. One eruptive site at Northumberland Canyon offers a unique opportunity to study features of a collapse caldera.

ACKNOWLEDGMENTS

The cooperation of many U.S. Geological Survey colleagues aided greatly in achieving the objectives of this study. In particular, Reuben J. Ross, Jr., visited the area many times to collect fossils and study stratigraphic relations in the field. Ross identified Ordovician fossils and offered stimulating discussion. Thomas E. Mullens, Forrest G. Poole, Robert L. Foster, John H. Stewart, and Chester T. Wrucke, all familiar with problems of Nevada stratigraphy, helped greatly in discussions of regional geology. Charles W. Merriam studied many of the Silurian and Devonian fossil collections, identified the coral-brachiopod fauna, and added invaluable insight into the relations suggested by these collections and those from elsewhere in Nevada. John W. Huddle and Jonathon C. Matti of the U.S. Geological Survey and Raymond L. Ethington, University of Missouri, identified conodonts, and W. B. N. Barry, University of California, Berkeley, identified Silurian and Devonian graptolites. The author expresses thanks to Marshall Kay of Columbia University for his interest and many helpful suggestions.

PALEOZOIC ROCKS

Paleozoic rocks exposed in the northern part of the Toquima Range are Ordovician, Silurian, Devonian, and Pennsylvanian. The pre-Pennsylvanian carbonate rocks broadly resemble the classic section near Eureka and Antelope Valley, Nev. (Merriam and Anderson, 1942; Merriam, 1963), and therefore are part of the eastern carbonate assemblage as described by Merriam and Anderson (1942), Roberts, Hotz, Gilluly, and Ferguson (1958), Gilluly and Gates (1965), and many other geologists. The siliceous rocks of the same age belong to the western assemblage, transported eastward into central Nevada in the upper plate of the Roberts Mountains thrust. The Pennsylvanian deposits are limestones and conglomerates that unconformably overlie the older rocks.

Rocks of the carbonate, eastern assemblage are exposed in two windows in the Roberts Mountains thrust. In the northern one, Petes Canyon window (McKee and Ross, 1969), the name Petes Canyon sequence is applied to all the eastern assemblage. In the southern one, Ikes Canyon window, three sequences of carbonate rocks are exposed, separated by two thrust faults. These three sequences in the Ikes Canyon window have been named the August Canyon, Mill Canyon, and June Canyon sequences by Kay and Crawford (1964). Figure 3 illustrates the thrust relations and the distribution of rock sequences in the northern Toquima Range, and figure 4 shows the correlation of the formations in the sequences.

UPPER PLATE OF THE ROBERTS MOUNTAINS THRUST; VININI FORMATION

DISTRIBUTION AND NATURE OF OUTCROPS

The Vinini Formation crops out chiefly in three parts of the area (see fig. 3; pl. 1): a wide irregular band across the central part of the range at Petes Summit, a narrow band across the southern part of the area near Wildcat Peak, and small areas at the north end of the range.

The strata near Petes Summit are the most revealing because they contain a series of graptolite zones and are relatively undeformed; thus a stratigraphic succession can be established. The formation in the southern area is generally too deformed to establish a stratigraphic sequence, and no fossils were found during this study. The relation between the two areas of outcrop is not known, and it is possible that they are separate thrust plates.

In general the Vinini Formation in this region comprises an unknown thickness of thin-bedded black chert (fig. 5), quartzite (fig. 6*A*), red to black siltstone, dark limestone, and in a few places thin flows of pillow basalt (fig. 6*B*). These strata resemble rocks of the western to transitional assemblage farther north and west in central Nevada. Because of the monotonous nature of this assemblage and the extreme complexity of structure (fig. 6), few satisfactory stratigraphic relations have been established within the formation. If the entire Vinini in the region is considered, there may be as much as 5,000–6,000 feet (1,500–2,000 m) of strata; a thickness of about 1,000 feet (300 m) can be demonstrated in the Petes Summit area.

Petes Summit area.—In the Petes Summit area the stratigraphy of the Vinini Formation in a series of imbricate thrust plates can be pieced together using several marker beds and intervals of distinctive lithologic types. A quartzite bed (fig. 6*A*) that is overlain and underlain by pastel graptolite-bearing shale provides a

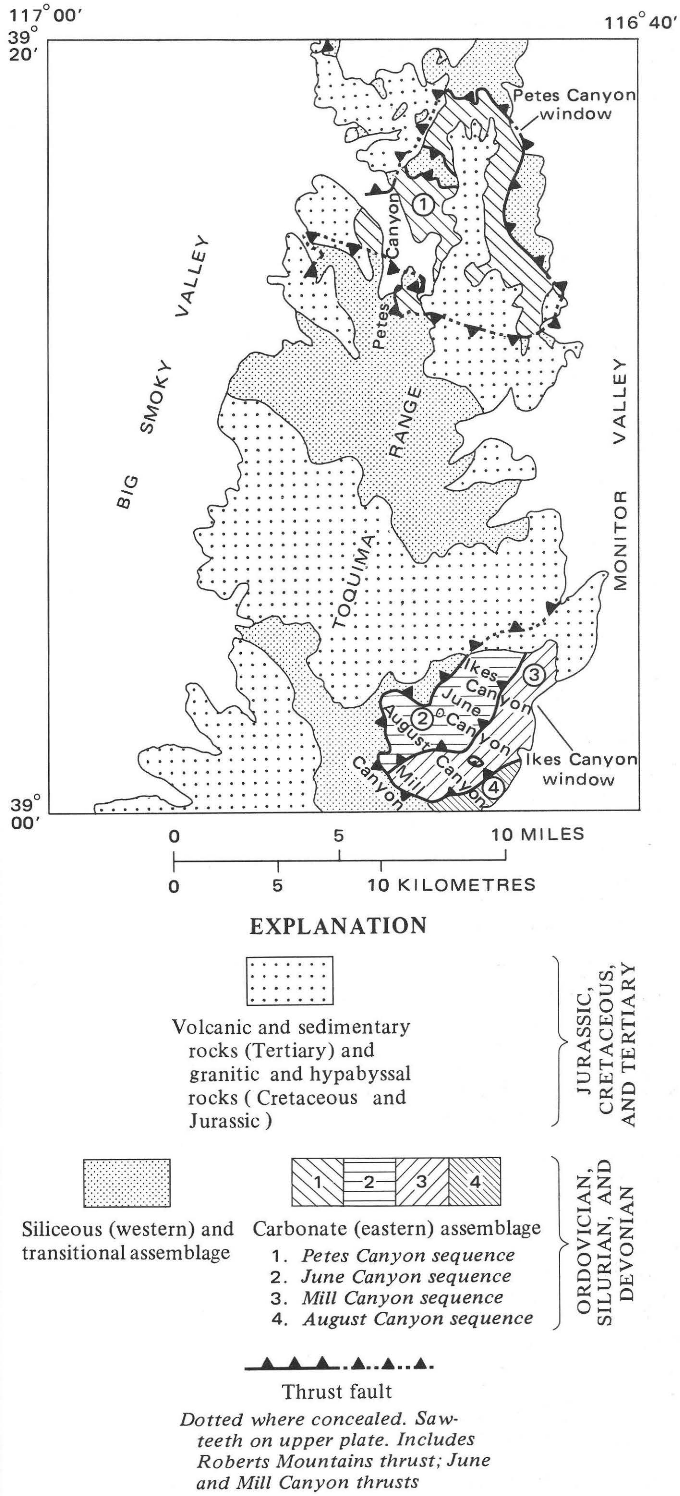
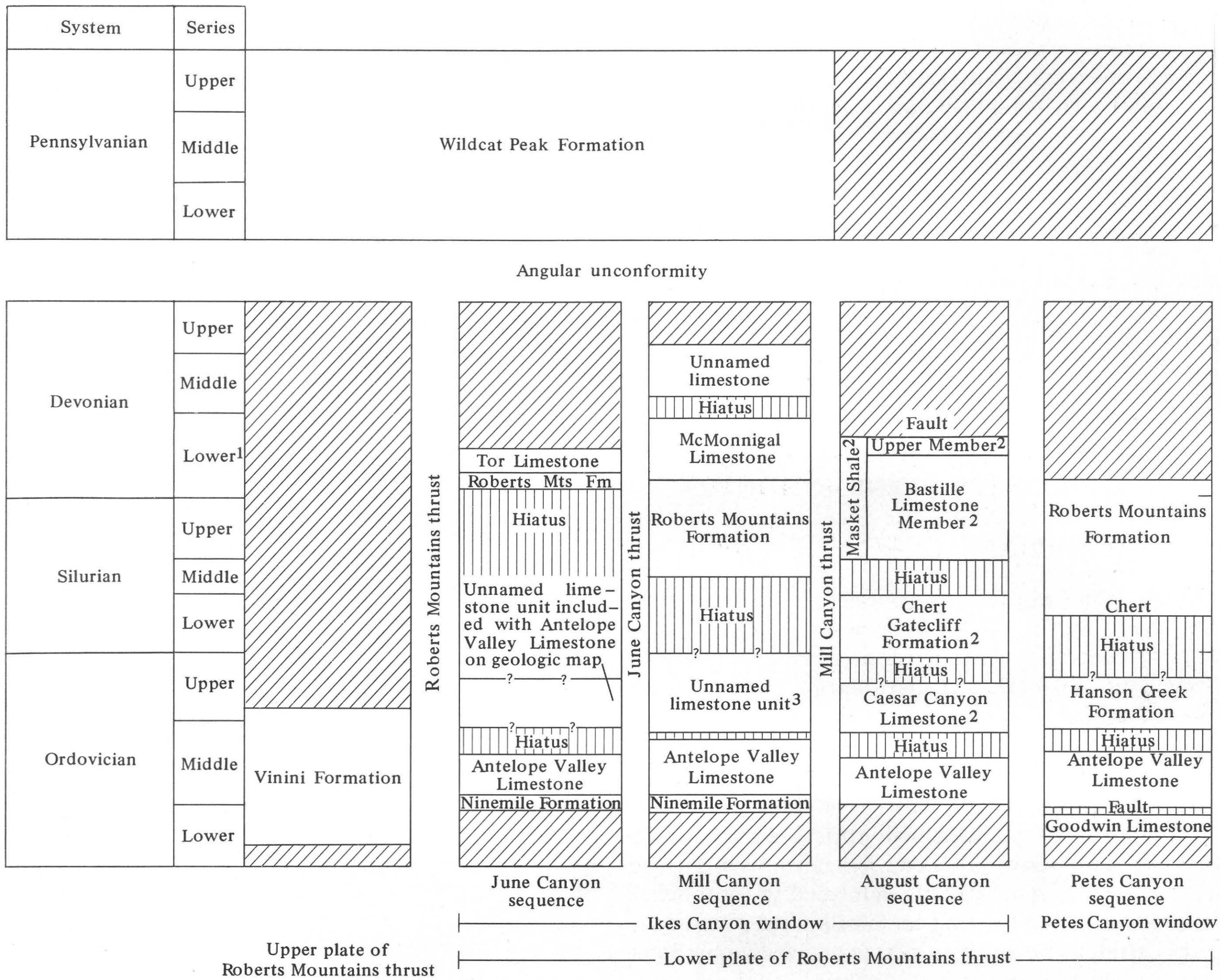


FIGURE 3.—Simplified geologic map of the northern part of the Toquima Range showing distribution of the eastern and western assemblages. The eastern assemblage comprises four distinctive sequences, two of them presumed to have been thrust into the area.

distinctive stratigraphic marker in this area. This bed forms a resistant wall along the top of a series of north-trending ridges where the section is cut by im-



[1] Includes Silurian coral zone E of Merriam (1973) and *Monograptus uniformis* zone (Devonian)
[2] Of Kay and Crawford (1964)
[3] Includes Diana Limestone of Kay and Crawford (1964)

FIGURE 4.—Correlation of Paleozoic formations in the northern part of the Toquima Range. Each column represents a different sequence of strata.

bricate faults that generally dip to the west, and it is repeated at least six times near Petes Summit (see cross section *C–C′*, pl. 1). The formation is divided into an upper and lower part in this report, the quartzite bed marking the basal unit of the upper part. A composite section of the formation in the area of the imbricate faults is shown in figure 7. Most of this section is unmeasured; thicknesses are estimated from one or more localities that contain less than 300 feet (100 m) of unfaulted strata. The upper third of this composite section is the most precisely known because it contains the quartzite marker bed as well as other markers. Most of the graptolite-bearing beds are in this stratigraphic interval. From this part of the section also and near Petes Summit, Kay (1960, 1962) and Kay and Crawford (1964) defined four new formations (Charcoal Canyon, Petes Summit, Sams Spring, and Joes Canyon) comprising their Clipper Canyon Group. In the present study the name Vinini Formation is retained for all these rocks because they resemble the Vinini Formation in neighboring parts of Nevada and because Kay and Crawford's formations are not recognizable map units elsewhere.

Northern Area.—The Vinini Formation in the northern part of the map area is composed of similar lithologies as in the Petes Summit area; however, it is

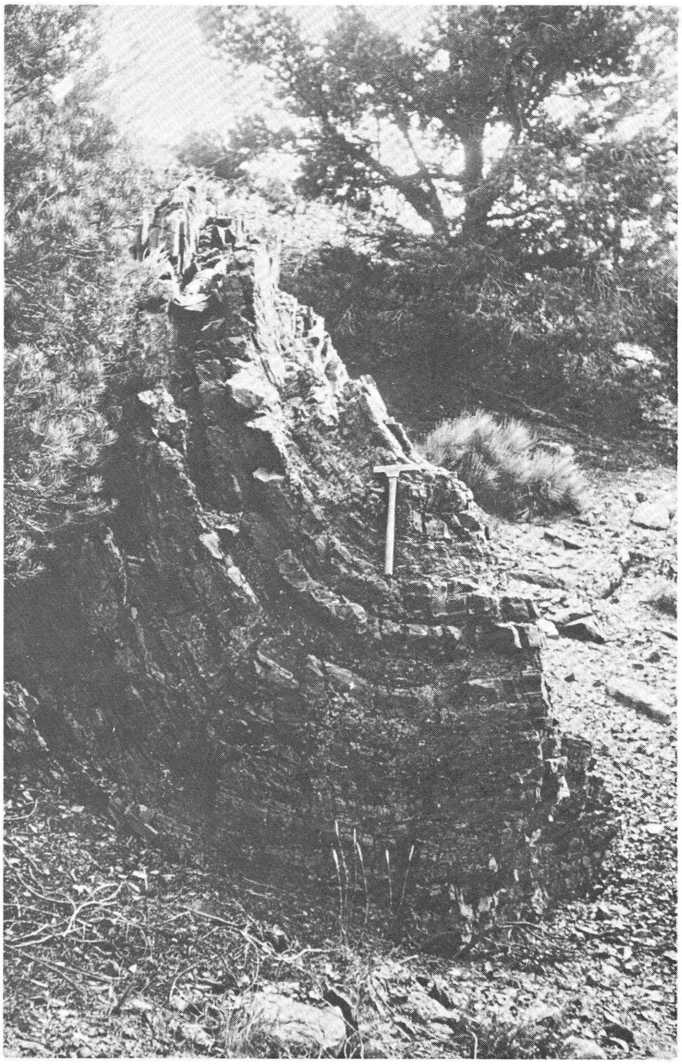


FIGURE 5.—Thin-bedded black chert and shale interbeds typical of most of the Vinini Formation. Degree of deformation is also typical of this formation in this area.

structurally more complex and is not here subdivided into upper and lower parts. The quartzite marker bed, key to unraveling the structure around Petes Summit, is not a conspicuous unit in the north although it or similar quartzites occur at scattered places. Dark to pastel siltstone usually with shaly partings and with a few graptolites is the major lithologic unit; thin-bedded dark chert is common, and thin (as much as 2 feet or 0.6 m thick) black or dark-gray limestone beds are a third significant lithology. The graptolites from this area are all types found near Petes Summit.

Southern area near Wildcat Peak.—Rocks mapped as the Vinini Formation in the southern part of the area are mostly thin- to medium-bedded dark chert and some dark shale now extremely deformed. Deformation in this area has contorted the strata to such an extent that no attempt was made to establish a strati-

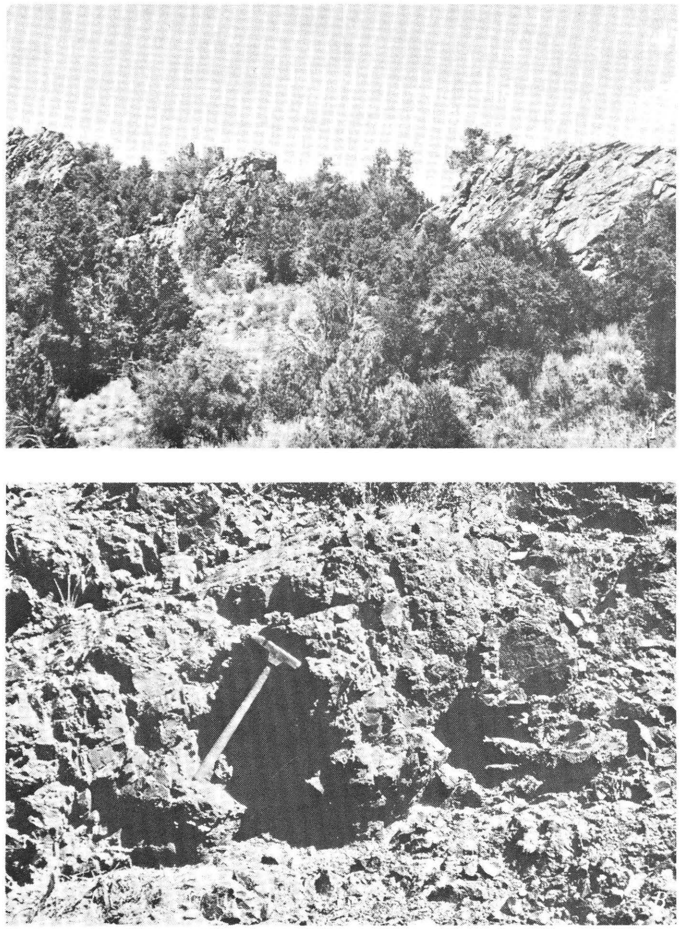


FIGURE 6.—Vinini Formation. *A*, Quartzite marker bed at the base of the upper member of the formation; here it is about 30 feet (10 m) thick. *B*, Pillow lava in the lower member of the formation.

graphic succession. Small- and large-scale faults with every possible orientation and displacement, as well as folds with a variety of amplitudes and trends, are seen in some places; elsewhere outcrops consist of a rubble of broken fragments. None of the marker beds or distinctive lithologic units of the Petes Summit area were recognized in the southern area, and no fossils were found. The relation of these rocks to those in the north is not known, and it is possible that they are an entirely separate and unrelated thrust plate and may be a different age. These strata are called the Vinini Formation because they most resemble parts of that formation from its type area at Vinini Creek in the Roberts Creek Mountains.

FAUNA, AGE, AND CORRELATION

Graptolites are common in shaly siltstones at numerous stratigraphic horizons in the Vinini Formation. Accompanying the graptolites are *Caryocaris,* phosphatic inarticulate brachiopods, and rodlike structures that are probably bryozoan. Conodonts occur in

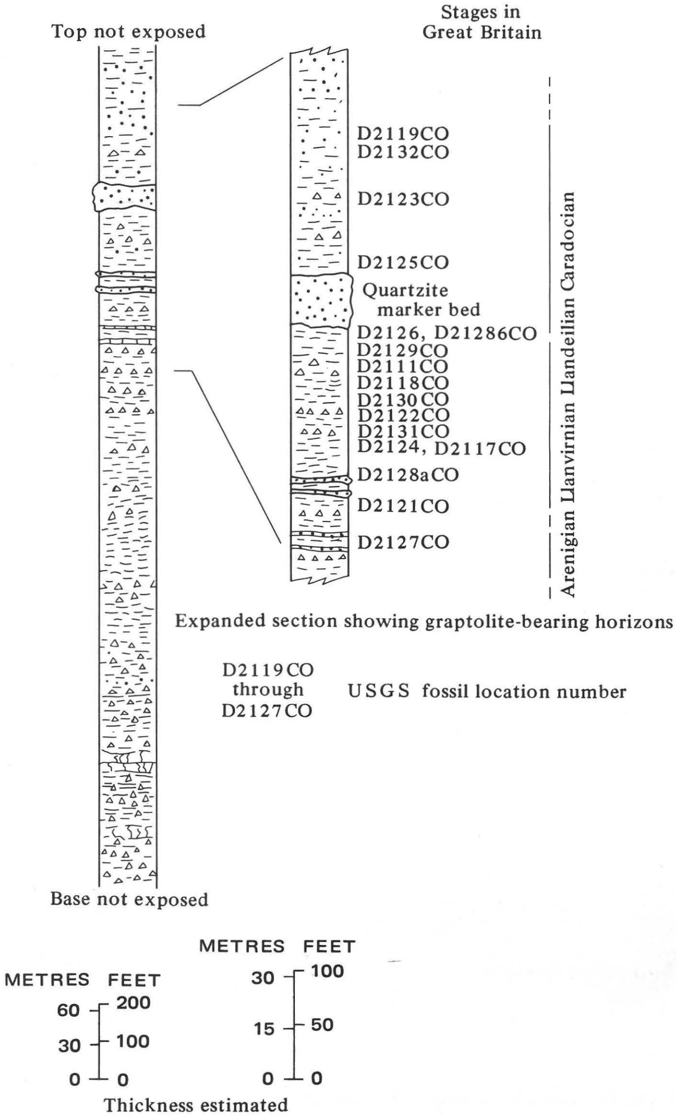


FIGURE 7.—Composite section of the Vinini Formation in the area around Petes Summit.

the limestone beds. No fossils have been found in the cherts, quartzites, or in areas where deformation has destroyed the original shaly partings characteristic of the siltstones.

In the Petes Summit area graptolite beds occur through about 400 to 600 feet (100–200 m) of the upper part of the section but have not been found in the topmost 100 (50 m) or lower 1,000 feet (300 m) of this section. A few graptolites from the area north of Petes Canyon have been collected, but they are not associated with the quartzite marker bed and hence remain to be fitted into a stratigraphic section. No fossils have been found in the southern part of the area. Collections keyed to the composite section near Petes Summit (fig. 7) are listed below in ascending stratigraphic order. The graptolites represent zones spanning most of the Middle and part of the Upper Ordovician Series (Arenigian, Llanvirnian, Llandeilian, Caradocian Stages). The conodonts that occur at about the same horizon as the lowest graptolites suggest a Middle Ordovician age. The stratigraphically lowest collections from the Vinini Formation below the quartzite marker unit (see fig. 7) are as follows (all graptolite identifications by R. J. Ross, Jr.):

USGS colln. D2127 CO; Nevada coordinates, central zone:
E. 463,700 ft., N. 1,622,000 ft.
Didymograptus cf. *D. extensus*
Tetragraptus fruticosus (four branched)
Didymograptus sp.

USGS colln. D2121 CO; Nevada coordinates, central zone:
E. 464,000 ft., N. 1,621,650 ft. (strat. position approx.)
Phyllograptus cf. *P. anna*
Didymograptus cf. *D. simulans* Elles and Wood
Didymograptus sp.
Climacograptus cf. *C. riddellensis* Harris
Isograptus caduceus cf. var. *maxima* Harris

USGS colln. D2128a CO; Nevada coordinates, central zone:
E. 454,950 ft., N. 1,619, 658 ft. (strat. position approx.)
Didymograptus two spp.
Tetragraptus quadribranchiatus
Glossograptus? sp.
Paraglossograptus sp.
Amplexograptus two spp. (one cf. *confertus*)
Glyptograptus sp. (very slender)

USGS colln. D2117 CO; Nevada coordinates, central zone:
E. 467,000 ft., N. 1,602,150 ft. (strat. position approx., about the same as colln. D2124 CO)
Glyptograptus cf. *G. dentatus* (Brogneart)

USGS colln. D2124 CO; Nevada coordinates, central zone:
E. 465,750 ft., N. 1, 595,800 ft. (about the same strat. position as colln. D2117 CO)
Glyptograptus n. sp. or may be *Diplograptus amplexograptoideus* Ross and Berry
Glossograptus sp.
Dicellograptus sp. (very small)

USGS colln. D2131 CO; Nevada coordinate, central zone:
E. 464,600 ft., N. 1,613,200 ft.
Didymograptus sp.
Phyllograptus sp.
Glossograptus sp.
Pterograptus sp. or *Pseudobryograptus*?
Amplexograptus cf. *A. confertus* as interpreted by Harris and Thomas in Australia

USGS colln. D2122 CO; Nevada coordinates, central zone:
E. 464,000 ft., N. 1,621,650 ft.
Didymograptus sp. (very wide stipes)
Amplexograptus sp. (at last two species)
Paraglossograptus sp.
Glossograptus sp.
Pseudoclimacograptus? sp.

USGS colln. D2130 CO; Nevada coordinates, central zone:
E. 467,250 ft., N. 1,597,500 ft. (strat. position approx.)
Tetragraptus sp.
Trichograptus aff. *T. immotus* Harris and Thomas
Cardiograptus crawfordi Harris
?*Glossograptus acanthus* Elles and Wood
Paraglossograptus etheridgei Harris
Diplograptus cf. *D. docratus* var. *amplexograptoideus* Ross and Berry

Amplexograptus arctus Elles and Wood
Amplexograptus confertus (Lapworth)
USGS colln. D2118 CO; Nevada coordinates, central zone: E. 454,050 ft., N. 1,614,900 ft.
Isograptus sp.
Glyptograptus sp. (very small)
USGS colln. D2111 CO; Nevada coordinates, central zone: E. 462,300 ft., N. 1,607,650 ft.
Glossograptus hincksi Hopkinson
Glyptograptus dentacus (Brogniart)
Glyptograptus sp.
USGS colln. D2129 CO; Nevada coordinates, central zone: E. 461,980 ft., N. 1,613,200 ft.
Didymograptus sp. (extensiform)
Dicellograptus intortus Lapworth
Dicellograptus sp.
Dicranograptus rectus Lapworth
Dicranograptus sp.
USGS colln. D2128b CO; Nevada coordinates, central zone: E. 454,950 ft., N. 1,619,650 ft. (strat. position same as colln. D2126 CO)
Climacograptus eximius Ruedemann
Orthograptus? sp.
Glyptograptus aff. *G. teretiusculus* (Hisinger)
Duranograptus nicholsoni
USGS colln. D2126 CO; Nevada coordinates, central zone: E. 462,750 ft., N. 1,623,050 ft. (strat. position same as colln. D2126 CO)
Dicellograptus sextans
Amplexograptus cf. *A. arctus* Elles and Wood
Orthograptus? or *Retiograptus*?

The collections from the upper part of the Vinini Formation (pl. 1) above the quartzite marker unit (see fig. 7) are as follows:

USGS colln. D2125 CO; Nevada coordinates, central zone: E. 464,100 ft., N. 1,614,950 ft.
Dicranograptus nicholsoni var. *whitianus*
Dicranograptus cf. *D. nicholsoni*
Dicellograptus sp.
Orthograptus cf. *O. truncatus*
Climocograptus sp.
USGS colln. D2123 CO; Nevada coordinates, central zone: E. 454,950 ft., N. 1,608,850 ft.
Climacograptus cf. *C. caudatus* Lapworth
Climacograptus spiniferus Lapworth
Orthagraptus cf. *O. truncatus intermedius*
Pseudoclimacograptus scharenbergi
Glyptograptus? cf. *G. dentatus*
USGS colln. D2132 CO; Nevada coordinates, central zone: E. 452,100 ft., N. 1,635,800 ft.
Glyptograptus sp.
Orthograptus truncatus var.?
Orthograptus? *quadrimucronatus*
USGS colln. D2119 CO; Nevada coordinates, central zone: E. 452,650 ft., N. 1,608,450 ft.
Orthograptus quadrimucronatus

Conodonts collected from thin limestone beds below the quartzite marker unit (see fig. 7) and identified by John W. Huddle include:

Acontiodus n. sp. (five)
Acodus sp. (two)
Distacodus cf. *D. stola* Lindström (two)
Drepanodus suberectus (Branson and Mehl) (three)
Oistodus sp. (seven)
Scolopodus striolatus Harris and Harres (four)
Drepanodus sp. (one, hooded form)
Oistodus venstus Stauffer (four)
Phragmodus? or *Periodon*? sp. (three)

Correlation of the Vinini Formation in the Petes Summit area with other rocks in central Nevada is possible because of its large graptolite fauna. The formation in the southern parts of the area is less well correlated because it has no fossils, and it is called Vinini Formation solely because it bears a lithologic similarity to that formation elsewhere.

The graptolite succession from the rocks near Petes Summit spans Middle Ordovician and extends into Late Ordovician time (fig. 7), from Arenigian into the Carodocian. Collections of graptolites from similar-looking rocks in the Simpson Park Mountains about 22 miles (35 km) north of Petes Summit and mapped as Vinini Formation by McKee (1968a) include genera diagnostic of some of the middle zones (Lanvirn) at Petes Summit, verifying the lithologic correlation. Strata mapped as Palmetto Formation (Ferguson and Cathcart, 1954) in the southern part of the Toiyabe Range in the vicinity of Wall Canyon and about 40 miles (65 km) from Petes Summit bear a striking lithologic similarity to the Petes Summit rocks and a comparable graptolite succession (F. G. Poole, oral commun., 1972). The Palmetto Formation in the southern Toiyabe Range and the Vinini Formation in the Petes Summit area of the Toquima Range are correlative, as their very similar lithologies suggest.

Direct correlation of the Vinini Formation of Petes Summit with eastern assemblage carbonate rocks is possible because of the presence of diagnostic graptolites in limestone of the Antelope Valley Limestone in Ikes Canyon. The species *Climacograptus riddelensis* Harris, indicative of the Middle Ordovician *Paraglossograptus etheridgei* zone, occurs in the Antelope Valley Limestone, and this zone is established in the Vinini Formation by the presence of *Trichograptus* aff. *T. immotus* Harris and Thomas, *Paraglossograptus etheridgei* Harris, and species of *Diplograptus, Amplexograptus,* and *Glossograptus*. The younger parts of the Vinini Formation at Petes Summit containing graptolites representative of the Carodocian correlate with graptolite-bearing rocks mapped as Hanson Creek Formation in the Petes Canyon sequence. Correlation is also probable with part of the Caesar Canyon Limestone of Kay and Crawford (1964) in the August Canyon sequence and the unnamed limestone units above the Antelope Valley Limestone of the Mill and June Canyon sequences.

LOWER PLATE OF THE ROBERTS MOUNTAINS THRUST

ORDOVICIAN SYSTEM

POGONIP GROUP

GOODWIN LIMESTONE

Limestone assigned to the Goodwin Limestone crops out on the east side of Henry Meyer Canyon in the northern part of the map area and is part of the carbonate eastern assemblage exposed in the Petes canyon window (Petes Canyon sequence). About 400 feet (120 m) of strata make up the formation here, but the total thickness of the unit is unknown because the lower contact is not exposed and the upper contact is a fault. The lowest 100 feet (30 m) of strata is thin-bedded light-gray very fine grained subporcellaneous cherty limestone (about 10 percent of the rock is chert) with a few distinctive beds of intraformational flat-pebble limestone conglomerate. The chert occurs as dark elongate blebs parallel to bedding, and in places stringers of chert define bedding. Above this 100+ foot 30+ m) interval is about 200 feet (60 m) of thin- to medium-bedded fine-grained gray fossiliferous limestone, some of which contains abundant concentric algal structures. The upper 100 feet (30 m) of the section is thin- to medium-bedded limestone characterized by abundant spherical algae as much as 2 inches (5 cm) in diameter. The formation as a whole is well bedded; beds vary from less than an inch (2.5 cm) to about 2 feet (0.6 m) thick.

The lowest occurrence of fossils is more than 100 feet (30 m) above the exposed base. Here a conodont collection contains *Cordylodus angulatus* Pander and *Cyrtoniodus prion* Lindström (identified by R. L. Ethington). These forms are dominant in the lower part of the type House Limestone (Lower Ordovician) of Hintze (1951) of western Utah; they are not found in the basal part of the Goodwin Limestone in the Antelope Range of central Nevada, which is generally considered correlative with the House Limestone (R. L. Ethington, written commun., 1967). In the middle part of the formation about 200 feet (60 m) above the base, the trilobites *Kainella* sp., *Apatokephalus* sp., *Trinodus* sp., and *Shumardia*? sp. have been collected (USGS colln. D1919 CO) and are equivalent to the fauna from the lower part of the Goodwin Limestone of the Antelope Valley area to the east (Merriam, 1963). Conodonts from about the same horizon, collected and identified by R. L. Ethington, include *Paltodus, Scolopodus, Oneotodus,* and *Oistodus forceps* Lindström, which are present also in the lower part of the Goodwin Limestone in the Antelope Range (R. L. Ethington, written commun., 1967). A collection (USGS colln. D1920 CO) from slightly higher in the middle part of the formation includes *Apatokephalus* sp., *Rossaspis* sp., and *Parabolinella* sp. and is probably equivalent in age to the middle of the Goodwin Limestone in the Antelope Valley area. The upper part of the formation is characterized by abundant spherical algae but contains no diagnostic fossils.

NINEMILE FORMATION

The Ninemile Formation crops out at several places in the Ikes Canyon window, where it is the lowest unit of the Mill Canyon and June Canyon sequences. Especially good exposures of the formation in the Ikes Canyon window are about 1 mile (1.6 km) west of the mouth of June Canyon, south of Ikes Canyon near hill 8937, and the north slope of Sawlog Ridge about 1 mile (1.6 km) southeast of Stoneberger Basin (pl. 1). At these places it consists of olive-green to brown siltstone with shaly partings and a few buff limestone beds 1–4 inches (2–10 cm) thick. It forms poor exposures and is commonly covered by a distinctive brown soil. The shale beds are usually contorted and sheared, and the thin limestones are broken and discontinuous. The formation is cut out by faults in the Petes Canyon window and does not occur above the Goodwin and below the Antelope Valley Limestones.

In the Ikes Canyon area of the Toquima Range, the formation was named the Stoneberger Shale by Kay (1960), and it was more fully described by Kay and Crawford (1964, p. 431) but not shown on their geologic map. In this study the Ninemile Formation and Kay's Stoneberger Shale are considered synonyms, and the well-established name Ninemile Formation, having priority, is retained.

The base of the Ninemile Formation has not been found in the Toquima Range. The upper contact with the Antelope Valley Limestone is gradational across a stratigraphic interval of about 50 feet (15 m), and the boundary is placed at the horizon above which limestone forms more than 50 percent of the section. A complete section of the formation does not occur at any single locality, but a thickness of about 400 feet (120 m) occurs on the south side of the upper reaches of June Canyon. On the basis of these exposures and the thickness of 550 feet (170 m) at the type locality in the Antelope Range (Merriam, 1963), it is estimated that the formation is 500–700 feet (170–215 m) thick in the Toquima Range. An estimate of 1,000 feet (348 m) or more by Kay and Crawford (1964, p. 431) seems excessive. There is no significant difference in thickness or lithology of the formation in the Mill Canyon and June Canyon sequences. The August Canyon sequence of the Ikes Canyon window does not expose rocks as old as the Ninemile Formation.

Graptolites, trilobites, brachiopods, ostracodes, bryozoans, and conodonts have been collected at several horizons in the Ninemile Formation in the Ikes Canyon window. The lowest of these collections, probably about 200 feet (60 m) below the Antelope Valley Limestone in the June Canyon sequence, includes (USGS colln. D2254 CO) trilobites, *Paraspis erugata* Ross and *Periscoclonus* sp.; (USGS colln. D2278 CO) Olenid, new genus and species, *Endymionia* sp., *Peraspis* sp., and *Carolinites* sp.; graptolites, *Didymograptus* cf. *D. extenus, Glyptograptus* sp., and *Isograptus caduceus* cf. var. *divergens;* and (USGS colln. D2280 CO), *Glyptograptus* cf. *G. austrodentatus* and a very questionable *Cryptograptus* sp. Collections (USGS colln. D2279 CO, D2162 CO) in the transition interval between the Ninemile Formation and Antelope Valley Limestone contain the trilobites *Carolinites* sp. (tuberculata), *Peraspis* sp., and *Endymionia* sp., the graptolites *Glyptograptus* cf. *G. austrodentatus, Glyptograptus* sp., Dichograptids (at least two genera and several species), *Amplexograptus*? sp., and *Diplograptus* sp. (cf. *D. decoratus* var. *amplexograptoideus* Ross and Berry). Conodonts from the thin buff limestone beds in the formation include (identified by J. W. Huddle), *Acontiodus* sp. (two), *Cordylodus* sp. (one), *Drepanodus contractus* Lindström (one), *Drepanodus* sp. (three), *Oistodus venstus* Stauffer (four), and *Phragmodus* sp. (one); also included are (USGS colln. 7014 CO) *Acontiodus* sp. (one), *Cordylodus* sp. (one), *Drepanodus* sp. (seven), *Periodon* sp. falodiform element, and *Sagittodontus*? sp. (one).

A collection (USGS colln. D2281 CO) from the formation in the Mill Canyon sequence about 100 feet (30 m) below the Antelope Valley Limestone contains the trilobite *Paraspes*? and the graptolite *Phyllograptus* aff. *P. typus* Hall or *P. ilicifolius* var. *grandis* Elles.

All fossils collected during this study from the Ninemile Formation in both thrust sequences (June and Mill Canyons sequences) indicate an early Middle Ordovician age, but these fossils are only from the upper part of the formation. The lower part of the formation is presumed to be Early Ordovician, as fossils of this age are reported by Kay and Crawford (1964) from these rocks. Graptolites suggest the zone of *Isograptus caduceus* or *Paraglossograptus etheridgei*. The trilobites suggest the *Orthidiella* zone and have been described in the lower part of the Antelope Valley Limestone of the Mill Canyon sequence near the mouth of Ikes Canyon (Ross, 1970). The conodonts are considered Middle Ordovician by J. W. Huddle (written commun., 1970).

Placement of the Ninemile Formation of the Toquima Range in the Early and Middle Ordovidian extends its age upward from the Early Ordovician age as defined in its type area to the east (Merriam, 1963, p. 23). This distinctive mappable unit has transgressed time in a westerly direction. Lower parts of the formation may contain Lower Ordovician fossils as reported by Kay and Crawford (1964), but they were not found during this study.

ANTELOPE VALLEY LIMESTONE

The Antelope Valley Limestone crops out in the Ikes Canyon window along the range from north of Ikes Canyon to south of June Canyon and in the upper reaches of these canyons as far south as Mill Canyon. In most places it forms cliffs or steep slopes and is geomorphically the most conspicuous formation in the window. Outcrops in the Petes Canyon window are on the north side of Henry Meyer Canyon and east of this canyon on the east side of the range.

The formation is recognized in all the stratigraphic sequences (August Canyon, Mill Canyon, and June Canyon) of the Ikes Canyon window as well as in the Petes Canyon sequence, and in all sequences it is generally similar and easily distinguished from other limestone units in the region. Variation of the formation between sequences involves relative amounts of silty partings between limestone beds, the average thickness of beds, and overall thickness of the unit. The formation is generally thin- to medium-bedded grayish-blue limestone with irregular patches that weather yellow orange. The mottled yellow-orange-on-gray appearance is characteristic of the formation, particularly where bedding is thin. Beds less than a few centimetres thick and shaly partings also tend to weather a yellowish color that contrasts with the uniform gray of more massive beds. Much of the rock weathers to form rough, pitted surfaces on which silicified brachiopod fragments are locally abundant. In other places finer grained gray limestone weathers to smooth, almost lithographic surfaces, and fossils in it, especially trilobites, appear as jet-black outlines. Diagnostic of the formation are *Girvanella, Receptaculites,* and large low-spired gastropods (*Pallisaria* and *Maclurites*).

The thickest sections of this formation are in the Mill Canyon sequence and are found near the mouth of Ikes Canyon and June Canyon. One section, probably the most complete in the area, is on the north side near the mouth of Ikes Canyon and is measured at more than 900 feet (275 m) (Ross, 1970). Further up Ikes Canyon the Antelope Valley Limestone in the June Canyon sequence is estimated to be about 700–750 feet (215–230 m) thick (McKee and others, 1972). Elsewhere only partial sections of the formation are found: in the Petes Canyon window (Petes Canyon sequence) it is estimated to be about 800 feet (245 m) thick (McKee and Ross, 1969); the August Canyon sequence (in the Ikes Canyon window) contains only about 100 m of the formation, with no base exposed.

Petes Canyon sequence

The Antelope Valley Limestone in the Petes Canyon window is exposed as several partial sections, none more than about 400 feet (120 m) thick. The total thickness of the formation in this sequence is probably about 800 feet (245 m). The basal contact is a fault, below which are cherty limestones of the Goodwin Limestone or, at one place, a small amount of sheared shale of the Ninemile Formation. The upper contact is depositional with the overlying Hanson Creek Formation. No angular discordance is evident between the Hanson Creek and Antelope Valley Formations, but a hiatus exists on the basis of a distinct age difference between the fauna of the two formations. Fossils from the Antelope Valley Limestone of the Petes Canyon sequence include (USGS colln. D1886 CO) the brachipod *Orthambonites minuscules* (Phleger) and the trilobite *Nileus* sp. indicative of the *Orthidiella* zone; (USGS colln. D1884 CO) the brachiopods *Orthambonites* cf. *bifurcatus* Cooper, *Syndielasma biseptatum* Cooper, *Taphrodonta parallela* Cooper, and *Leptellina occidentalis* Ulrich and Cooper; the trilobites *Basilicus mckeei* Ross, *Bathyurellus fietleri, Kawina* sp., and *Nileus hesperafiinis* Ross; and the ostracodes (identified by Jean M. Berdan) *Eoleperditia* sp. aff. *E. bivia* (White), *Leperditella* sp., and *Schmidtella* sp. The latter collection is representative of the *Anomalorthis* zone.

August Canyon sequence

The Antelope Valley Limestone near the range front north of Mill Canyon is the oldest formation exposed in the August Canyon sequence. The exposed section is only about 100 m thick and is faulted to such an extent that detailed stratigraphic studies and thickness measurements have not been made. The basal contact is not exposed, but the contact with the overlying Caesar Canyon Limestone of Kay and Crawford (1964) is well exposed at several localities. This contact appears conformable, but faunal evidence indicates that the formations are separated by a hiatus of considerable duration (fig. 4).

Fossils have been collected from the upper part of the formation in beds about 50 feet (15 m) beneath the Caesar Canyon Limestone. These include (USGS colln. D2288 CO) the brachiopod *Orthidiella costellato* Cooper, the trilobite *Ectenonotus* sp., and the following conodonts:

Beolodina n. sp. (one)
Belodina n. sp. (three)
Drepanodus suberectus (Branson and Mehl) (two)
Multioistodus cf. *Oistoidus longiramus* Lindström (one)
Periodon aculeatus Hadding
 cordylodian element (four)
 ozardodinian element (seven)
 falodian element (four)
"*Oistodus*" cf. *O. abundans* Branson and Mehl (six)
"*O.*" sp. (two)
"*O.*" sp. (one)
Scolopodus sp. (one)

The brachiopod and trilobite are indicative of the *Orthidiella* zone or temporally equivalent to the lower part of the Antelope Valley Limestone at its type area. The overlying *Anomalorthis zone* seems to be missing. The conodont assemblage is a new fauna including distinctive new species of *Belodina* and of *Multioistodus*. *Periodon aculeatus* is known from several horizons in the Antelope Valley Limestone in Nevada, but its range is uncertain. The collection suggests an early Middle Ordovician age (J. W. Huddle, written commun., 1972).

Mill Canyon sequence

The best known section of Antelope Valley Limestone in the Toquima Range is in the Mill Canyon sequence. Excellent outcrops of Antelope Valley Limestone exposed on both sides of Ikes Canyon near its mouth have attracted geologists for a number of years, and fossils from this locality have figured prominently in studies of Ordovician shelly faunas of the Great Basin. This section has been described by Ross (1970), who gave a detailed stratigraphically controlled faunal list and discussed the correlation of the Antelope Valley Limestone in the Mill Canyon sequence with sections of the formation east of the Toquima Range. He measured a section 950 feet (290 m) thick that extends from thin-bedded limestone probably less than 100 feet (30 m) above the Ninemile Formation to the contact with an unnamed Ordovician limestone. Thus the formation in the Mill Canyon sequence is about 1,000 feet (300 m) thick. The basal contact with the Ninemile is gradational across an interval of about 50 feet (15 m) and is drawn at the horizon above which limestone is the main rock type and siltstone is subordinate. The upper contact appears conformable and is recognized by the presence of a thin lenticular bed of phosphatic sandstone separating massive dark-gray limestone of the overlying unnamed unit from medium-bedded gray and yellow Antelope Valley Limestone below. Faunal evidence indicates that a hiatus exists between these two limestones (fig. 4, Mill Canyon sequence).

Fossils (see Ross, 1970, for faunal lists) indicate that the same faunal zones are represented in the Antelope Valley Limestone of this sequence as at its type area—the *Orthidiella* zone and overlying *Anomalor-*

this zone—although in the type area the Antelope Valley Limestone includes some Lower Ordovician strata as well. Since the upper part of the Ninemile Formation and the overlying Antelope Valley Limestone of the Toquima Range are both Middle Ordovician, it is evident that the Ninemile–Antelope Valley Limestone contact is time transgressive from Early to Middle Ordovician between the Antelope Valley region and the Toquima Range.

June Canyon sequence

About a mile and a half (2.4 km) from the mouth of Ikes Canyon, the Antelope Valley Limestone of the June Canyon sequence forms steep slopes and cliffs on both sides of the canyon. The predominantly thin-bedded character of the formation in this sequence contrasts with the thicker bedded, more massive cliff-forming limestone of the Mill Canyon sequence at the mouth of Ikes Canyon. A complete section of the Antelope Valley Limestone crops out 2 miles (3.2 km) above the mouth of Ikes Canyon. The basal contact with the underlying Ninemile Formation occurs on the south side of the canyon, but most of the formation crops out on the north side. A total thickness of about 700–750 feet (215–230 m) is estimated for the formation. The basal contact is gradational across a stratigraphic interval of about 50 feet (15 m) as in the Mill Canyon sequence, and the upper contact is drawn on the geologic map at the base of the Tor Limestone. A series of medium- to thin-bedded gray limestone beds (see p. 56) totaling about 20 feet (6 m) in thickness overlying the formation contains a distinctive Upper Ordovician fauna similar to that of the Caesar Canyon Limestone of Kay and Crawford (1964). These beds are mapped with the Antelope Valley Limestone in the June Canyon sequence because they are impossible to separate at the small scale of the map, but they are not considered part of the formation. A hiatus spanning a large part of the Middle Ordovician exists between these limestone beds and underlying 700 feet (215 m) of Antelope Valley Limestone (fig. 4).

Fossils collected from the Antelope Valley Limestone on the north side of Ikes Canyon and reported by McKee, Ross, and Norford (1972) include brachiopods, trilobites, conodonts, ostracodes, and graptolites. They constitute a significant collection, as they allow comparison of shelly and graptolitic zones. The shelly fauna through the entire Antelope Valley Limestone (and the underlying Ninemile Formation) in the June Canyon sequence is representative of the *Orthidiella* zone; possibly a small amount of the *Anomalorthis* zone is present at the top of the formation. Graptolites found with the brachiopods and trilobites of the *Orthidiella* zone include *Climacograptus riddelensis* Harris (a guide to the zones of *Paraglossograptus etheridgei*) and *Glyptograptus teretinusculus,* which previously had been correlated with the shelly *Anomalorthis* zone (Ross and Berry, 1963). It is evident that the graptolitic *P. etheridgei* and *Didymograptus bifidus* and *D. artus* zone beneath it correlate with the *Orthidiella* zone in central Nevada. Ostracodes that are locally abundant include (USGS colln. D2276 CO) *Eoleperditia bivia* (White), *Eobromidelia*? sp., *Tsitrella*? sp., Ectoprimitia sp., *Schmidtella*? sp., *Baltonotella* sp., *Macrocyproides* sp., and smooth ostracodes, (indet.). With the present incomplete knowledge of ostracode ranges in the Antelope Valley Limestone of the Great Basin, little can be said about the significance of these forms except that they suggest an age somewhat younger than most of the Antelope Valley collections so far studied by J. M. Berdan (written commun., 1973).

Age and correlation

The Antelope Valley Limestone in the four thrust sequences of the northern part of the Toquima Range is of Middle Ordovician age. The four sequences, however, are not identical; two of them contain upper Middle Ordovician strata of the *Anomalorthis* zone and lower Middle Ordovician rocks of the *Orthidiella* zone, and two contain only the lower zone. The type section, about 20 miles (30 km) to the east, contains upper Lower Ordovician strata and the two Middle Ordovician brachiopod zones as well. The Petes Canyon and Mill Canyon sequences contain both brachiopod zones and are about equivalent in age to the type section (less a small amount of upper Lower Ordovician), that is, equivalent to the Middle Ordovician Whiterock Stage of Cooper (1956, chart 1). The August Canyon and June Canyon sequences contain only the *Orthidiella* zone, which makes up the lower part of the type section or the lower part of the Petes Canyon and Mill Canyon sequences.

HANSON CREEK FORMATION

The Hanson Creek Formation in the northern part of the Toquima Range occurs only in the Petes Canyon window. It crops out on the north side of Henry Meyer Canyon and near the east margin of the range. Outcrops are poor, but the formation shows clearly as a dark zone on aerial photographs. The formation varies in thickness from about 35 to 75 feet (10 to 25 m) and although angular discordances at the base or top are not obvious, some of this variation probably reflects unconformities at both horizons. Faunal evidence clearly indicates breaks in the stratigraphic column at both the top and bottom of the unit.

The lowest beds in the formation are massive dark cherty limestones. These limestones are succeeded by thinner bedded cherty limestone and dark-red to green shale in the middle part of the formation and thin-bedded limestone, cherty limestone, and some lenticular chert beds in the upper part. The upper part of the formation seems to grade into thin-bedded black chert of the basal part of the Roberts Mountains Formation, but minor faulting at places where this contact is best exposed obscure the relations.

Graptolites and conodonts have been collected from the Hanson Creek Formation in the Petes Canyon window. The graptolites, which include *Climacograptus, Glyptograptus*? sp., *Orthograptus* sp., *Dicellograptus*? sp., and possible fragments of *Dicranograptus* (McKee and Ross, 1969) are considered to be Late Ordovician and are similar to the fauna from the Hanson Creek Formation and possibly the upper member of the Copenhagen Formation on the east side of the Monitor Range. Conodonts include *Acontiodus, Oistodus forceps, Ozarkodina*?, and dichognathid and phragmodid types (R. L. Ethington, written commun., 1967). The conodont fauna most closely resembles that from the upper member of the Copenhagen Formation in the Monitor Range, and some of the forms are common in the Caesar Canyon Limestone of Kay and Crawford (1964) in the August Canyon sequence of the Ikes Canyon window (R. L. Ethington, written commun., 1967). The Hanson Creek Formation in the Petes Canyon window also correlates at least partly with the unnamed limestone above the Antelope Valley Limestone in the Mill Canyon sequence, and the upper limestone bed mapped with the Antelope Valley Limestone in the June Canyon sequence. Fossils have not been found in the upper cherty part of the formation to indicate whether it is Early Silurian, as is the upper part of the Hanson Creek Formation in Eureka County (Mullens and Poole, 1972).

CAESAR CANYON LIMESTONE OF KAY AND CRAWFORD (1964)

The name Caesar Canyon Limestone was first used on two illustrations by Kay in 1960, and in 1964 Kay and Crawford defined it as argillaceous limestone beds above the Antelope Valley Limestone and beneath their Gatecliff Formation. The Caesar Canyon Limestone of Kay and Crawford (1964) occurs only in the August Canyon sequence. It was correlated with the upper part of the Copenhagen Formation in the Antelope Valley region (Kay, 1960; Kay and Crawford, 1964). The Caesar Canyon Limestone in this report includes the same rocks as those described by Kay and Crawford (1964).

The formation crops out in fault blocks along the east front of the range north of Mill Canyon and south of June Canyon. Exposures are poor, but enough partial sections include both contacts to establish the general stratigraphy. The lower 40 to 50 feet (12–15 m) is medium-bedded gray to black argillaceous limestone characterized by concentrations of phosphatic debris and phosphatic sand grains. Especially distinctive in the phosphatic accumulations are unflattened (three-dimensional) fragments of graptolites and delicately pitted cephelons of cryptolithoid trilobites. Both types of fossils show extremely sharp morphologic detail that is accentuated by their jet-black color. The upper 40–50 feet (10–15 m) of the formation is thin- to medium-bedded argillaceous and cherty limestone containing thin interbeds of black chert and black shale. The contact with the underlying Antelope Valley Limestone appears conformable on a local scale as does the upper contact with the Gatecliff Formation of Kay and Crawford (1964). Fossil evidence indicates that an unconformity exists at the basal contact and probably the upper contact as well.

Graptolites, trilobites, brachiopods, bryozoans, and conodonts have been collected from the Caesar Canyon Limestone of Kay and Crawford (1964). The most diagnostic of these forms are the graptolites, which include (USGS colln. D2181 CO and D2182 CO) diplograptid graptoloids, *Climacograptus* sp. and *Amplexograptus* sp.; brachiopods, *Orthambonites*? sp. and *Skenidiodes* sp.; and trilobites, *Cryptolithoides* sp., *Ampyxina salmoni* Churkin, *Primaspis*? sp., and *Robergea* sp.

The fauna listed by Kay and Crawford (1964, p. 433) indicated to them correlation with the upper part of the Copenhagen Formation. This correlation seems to be verified by the additional collections reported here and by studies of conodonts made by R. L. Ethington (written commun., 1967). In addition, the graptolites indicate a correlation with the lower part of the Hanson Creek Formation. The Caesar Canyon Limestone of Kay and Crawford (1964) correlates with the unnamed limestone above the Antelope Valley Limestone in the Mill Canyon sequence and with the upper limestone bed included on plate 1 with the Antelope Valley Limestone in the June Canyon sequence.

UNNAMED LIMESTONE IN MILL CANYON SEQUENCE

From 30 to 100 feet (10–30 m) of dark-gray argillaceous limestone overlies the Antelope Valley Limestone and underlies the Roberts Mountains Formation in the Mill Canyon sequence. This unit is difficult to differentiate from the underlying Antelope Valley Limestone in places, but the presence of a thin lenticular phosphatic sandstone serves as a key marker between the formations. This sandstone is as much as 3 inches (8 cm) thick and fills depressions or forms a local scattering of phosphatic debris and isolated pebbles and sand grains directly above the Antelope Valley Limestone. Finely preserved, delicately sculptured pieces of cryptolithoid trilobites and segments of unflattened graptolites, reminiscent of the Caesar Canyon Limestone of Kay and Crawford (1964), occur in the phosphatic sandstone. Phosphatic sandstone also is locally interbedded with black or dark-red shale at the base of the formation. The remainder of the formation is medium- to thin-bedded dark limestone locally containing a massive light-gray cliff-forming limestone about 30 feet (10 m) thick in the upper part. This cliff-forming rock was distinguished by Kay (1960) and named the Diana Limestone, and it was described and shown on the geologic map of Kay and Crawford (1964, pl. 1). Unfortunately, printing errors on the map have led to confusion as to the location and even the stratigraphic position of this rock. In this report all the rocks above the Antelope Valley Limestone and below the Roberts Mountains Formation are mapped as the unnamed limestones, including the Diana Limestone of Kay and Crawford (1964). The upper contact with the Roberts Mountains Formation seems concordant, but the Diana Limestone of Kay and Crawford is missing in places, suggesting that this contact is an unconformity. Fossil evidence supports the stratigraphic evidence for an unconformity.

Trilobites and conodonts collected from the unnamed limestone are considered to be late Middle or Late Ordovician (Ross, 1970). Graptolite collections from the phosphatic sand bed at the base of the unit are also late Middle or Late Ordovician, namely graptolite zone 13 of Berry (1960). These collections include (USGS colln. D1901 CO) *Climacograptus typicalis, Diplograptus*? sp., *Orthograptus truncatus* var. *strigosus, Orthograptus truncatus* cf. var. *intermedius, Orthograptus truncatus* var. *pertenius*; (USGS colln. D2136 CO) *Climacograptus* sp., *Climacograptus scharenbergi* Lapworth?, *Glyptograptus* sp., *Orthograptids* of the *O. truncatus group; (USGS colln. D2253 CO) Climacograptus* cf. *C. typicalis, Orthograptus* cf. *O. truncatus intermedius, Orthograptus* cf. *O. truncatus* var., *Dicellograptus*? or *Dicranograptus*? fragments; (USGS colln. D2180 CO) *Climacograptus* sp., the trilobites *Robergia* major Raymond and *Raymondella* sp., and corals questionably identified as *Streptelasma* sp., and favositid types. Conodonts include USGS colln. 7015 CO, 7016 CO, D2180 CO) the following:

Acontiodus sp. (one)
Cordylodus sp. (one)
Dichognathus sp. (one)
Drepanodus suberectus (Branson and Mehl) (sixteen)
Falodus (one)
Eoplacognathus sp. (forty-three)
Ambalodus sp. (eighteen)
Polyplacognathus sp. (ten)
Prioniodus? sp. (five)
Paltodus sp. (two)
Ligonodina sp. (one)
Scolopodus cf. *S. insculptus* (Branson and Mehl) (eighty)
Amorphognathus sp. (one)
Oistodus parallelus Pander (three)
Acodus? sp. (one)
Trichonodella sp. (one)

The lenticular massive upper limestone, the Diana Limestone of Kay and Crawford (1964), was considered to be Silurian in age on the basis of *Halysites* sp. and *Favosites* sp. (Kay and Crawford, 1964, p. 439 and 455). This age has been questioned by other geologists (see Berry and Boucot, 1970, p. 208; Ross, 1970, p. 21) who considered it to be Ordovician in age. This difference in interpretation results from mislabelling of the Diana Formation on the map of Kay and Crawford (1964, pl. 1), an error that caused subsequent workers to misidentify the formation and sample rocks other than those intended to be Diana by Kay and Crawford. A collection (USGS colln. D276S D) made during this study near Kay and Crawford's (1964, pl. 1) Diana Peak yielded the brachiopods *Eostropheodonta* sp. and *Gypidula*? sp., which suggest an Early Silurian age (J. T. Dutro, Jr., written commun., 1970) and which seem to verify the age designation of Kay and Crawford. However, the trilobites, graptolites, and conodonts from the unit that suggest a late Middle or Late Ordovician age (see p. 15) are considered by me to more accurately date the unit as Ordovician.

Most of the unnamed limestone in the Mill Canyon sequence correlates with the Caesar Canyon Limestone of Kay and Crawford (1964) in the August Canyon sequence (fig. 4). However, no upper massive bed (Diana Limestone of Kay and Crawford) is known in the Caesar Canyon Limestone of Kay and Crawford (1964), and this part of the formation most likely correlates with the basal part of the Gatecliff Formation of Kay and Crawford (1964). The unnamed limestone probably correlates also with the Hanson Creek Formation of the Petes Canyon sequence and part of the Hanson Creek, plus the upper part of the Copenhagen Formations of the Antelope Valley region east of the Toquima Range.

LIMESTONE OVERLYING THE ANTELOPE VALLEY LIMESTONE IN JUNE CANYON SEQUENCE

About 20 feet (6 m) or less of medium- to thin-bedded gray argillaceous limestone above the Antelope Valley Limestone and beneath the Roberts Mountains Formation in the June Canyon sequence may belong to the unnamed limestone of the Mill Canyon sequence. This limestone is mapped with the Antelope Valley Limestone on the geologic map (plate 1) because it is too thin to show on the scale of 1:62,500. In most places the upper contact is a fault with Tor Limestone; the thin-bedded to shaly Roberts Mountains Formation that lies between these more massive units has been cut out by faulting. Only at one locality on the north side of Ikes Canyon, about a mile and a half (2.4 km) from the mouth of the canyon, has any Roberts Mountains Formation been found. Here less than 10 feet (3 m) of yellowish-weathering platy to shaly argillaceous limestone containing monograptids separates the unnamed limestone (top of the Antelope Valley) from the Tor Limestone.

No diagnostic fossils have been collected in the limestone overlying the Antelope Valley Limestone in this sequence. Its age and correlation with the Middle or Late Ordovician unnamed limestone unit in the Mill Canyon sequence and with the Caesar Canyon Limestone of Kay and Crawford (1964) in the August Canyon sequence are speculative. If it does correlate with these units, it is separated from the underlying Middle Ordovician Antelope Valley by an unconformity.

SILURIAN SYSTEM

GATECLIFF FORMATION OF KAY AND CRAWFORD 1964

The name Gatecliff Formation, which first appeared on two illustrations by Kay in 1960, was described by Kay and Crawford in 1964 and applies to chert and dolomite on the north side of Mill Canyon. The best exposures of the formation are about a mile (1.6 km) west of the mouth of the canyon (fig. 8). This formation is found only in the August Canyon sequence. The cherty part of the Gatecliff Formation resembles the basal chert of the Roberts Mountains Formation of the Petes Canyon sequence and elsewhere in Nevada but is not considered to be part of that formation because the unusual thickness of dolomite and sandy dolomite beneath the chert is unique to the August Canyon sequence. As used in this report, the Gatecliff Formation is the same as described by Kay and Crawford (1964).

The Gatecliff consists of a basal light-gray medium- to thin-bedded dolomite about 60 feet (20 m) thick, a middle massive sandy crossbedded dolomite about 25 feet (8 m) thick (fig. 9*B*), and an upper thin-bedded gray dolomite and black chert (fig. 9*A*) as much as 70 feet (22 m) thick. Chert makes up a large part of the upper half of this upper unit. This cherty section is probably the most easily identified Paleozoic unit in the Toquima Range. It forms a black cliff that can be traced along the east front of the range in the vicinity of Mill Canyon and for about a mile (1.6 km) to the north of this canyon.

FIGURE 8.—Upper part of the Gatecliff Formation of Kay and Crawford (1964). Dark unit is the upper cherty part of the formation.

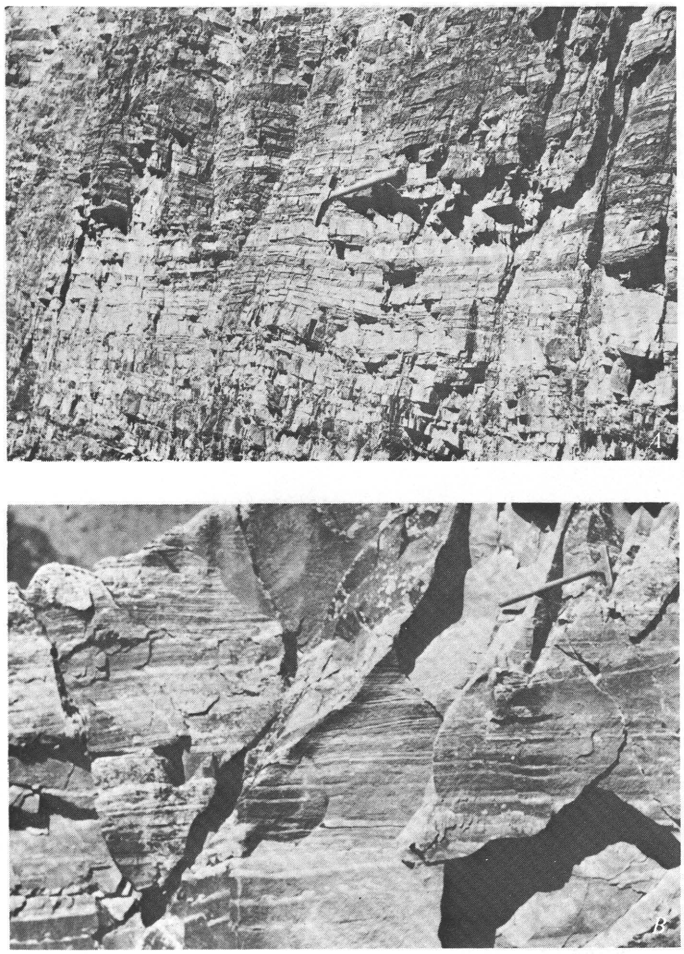


FIGURE 9.—Gatecliff Formation of Kay and Crawford (1964). *A*, Upper cherty unit. *B*, Quartz sand-bearing dolomite beneath the cherty unit.

The lower contact of the formation with the Caesar Canyon Limestone of Kay and Crawford (1964) appears conformable, but fossil evidence in the two units indicates a hiatus of considerable duration separates the formations (see August Canyon sequence, fig. 4). The upper contact shows no physical evidence of disconformable relations but may be an unconformity. This contact is between black chert and gray dolomite of the Bastille Limestone Member of the Masket Shale of Kay and Crawford (1964). The light-gray dolomite characteristic of the basal part of the Gatecliff and interbedded with chert in the upper part of the formation is similar to the dolomite of the bottom part of the Bastille.

Fossils are scarce in the Gatecliff Formation of Kay and Crawford (1964); only two collections consisting of corals and crinoids have been made from this formation in the August Canyon sequence. However, Kay and Crawford (1964) have identified graptolites collected in chert of the Gatecliff Formation in the Prospect thrust sequence of Kay and Crawford (1964) about 6 miles (10 km) south of the area. These graptolites from the Prospect sequence helped in dating the Gatecliff Formation. A collection (USGS colln. D2163 CO) from the lower dolomite beds in the August Canyon sequence on the north side of Mill Canyon yielded the corals identified by W. A.. Oliver and D. R. Budge in 1971: *Cladopora*? sp., *Favosites* (three) spp., *Halysites* or *Cystihalysites* sp., "*Cystiphyllum*" sp., and amplexoid coral, which indicate a Silurian age. Graptolites from chert in the Gatecliff Formation of the Prospect sequence about 6 miles (10 km) south of Mill Canyon and identified by W. B. N. Berry are *Climacograptus scalaris* and *Climacograptus* cf. *C. medius*. These fossils indicate an Early Silurian (early Llandoverian) age.

The upper cherty part of the Gatecliff Formation of Kay and Crawford (1964) correlates with the basal chert beds of the Roberts Mountains Formation present at a number of places in Nevada. These distinctive rocks of earliest Silurian age in the Prospect sequence south of Mill Canyon transgress time in a northeasterly direction (F. G. Poole, written commun., 1971) but remain in the Lower Silurian throughout their extent. The underlying sandy dolomite unit and gray dolomite beds at the base of the formation, also probably Silurian in age, may correlate with the Diana Limestone of Kay and Crawford (1964) in the Mill Canyon sequence and the upper part of the Hanson Creek Formation in Eureka County (Mullens and Poole, 1972). No rocks that are comparable in age or lithology are found in either the June Canyon sequence of the Ikes Canyon or in the Petes Canyon sequence of the Petes Canyon window.

SILURIAN AND DEVONIAN SYSTEMS

The base of the Devonian System as used in this report follows the recommendation of the committee on the Silurian-Devonian boundary (McLaren, 1970) by considering *Monograptus uniformis* to be the lowest Devonian. As so defined, the Lower Devonian in the Toquima Range includes Silurian coral zone E of Merriam (1973).

ROBERTS MOUNTAINS FORMATION

Good exposures of the Roberts Mountains Formation are in both the Petes Canyon and Ikes Canyon windows. The formation is a part of the Petes Canyon sequence and the Mill Canyon sequence, and a very small amount, probably faulted, occurs in the June Canyon sequence. The formation in the northern part of the Toquima Range is predominately thin-bedded platy-splitting silty fine-grained limestone that characteristically weathers to light yellowish tan and forms slopes covered with loose rubble of the weathered rock. Impressions of large monograptid graptolites locally form a crisscross pattern of the surface of thin slabs of the limestone. A few medium to thick graded beds composed mostly of bioclastic debris occur in the middle and upper part of the formation. The basal part of the formation in the Petes Canyon sequence is thin-bedded black chert, a well-known stratigraphic marker throughout central Nevada; the chert is not present in the Mill Canyon sequence. The total thickness of the formation can be measured only in the Mill Canyon sequence where it is bracketed by more massive limestones. Here it is about 350 feet (110 m) thick. It is probably more than twice this thick in the Petes Canyon sequence.

The formation is known from only one locality in the June Canyon sequence. On the north side of Ikes Canyon about 2 miles (3.2 km) from its mouth, a section about 10 feet (3 m) thick of thin-bedded yellowish-weathering argillaceous limestone crops out on top of the Antelope Valley Limestone and beneath the Tor Limestone. Elsewhere in this sequence the formation is missing and is presumed to have been cut out by faulting. The abnormal thinness of the existing section may be the result of faulting, but the possibility also exists that only a small amount of the argillaceous limestone was deposited unconformably on the Antelope Valley Limestone. Fossils indicate that only the uppermost part of the formation is present in this sequence. The upper contact with the Tor Limestone is conformable.

Graptolites are the most common fossil found in the Roberts Mountains Formation in the Toquima Range. They occur mostly in the thin-bedded platy-splitting limestone typical of the formation in the area. Corals occur locally in graded bioclastic medium- to massive-bedded parts of the formation, and conodonts are abundant in the thicker beds. The graptolites in the lower and middle part of the formation are monograptid forms typical of the middle part of the Silurian (equivalent to upper Llandovery, Wenlock, and lower Ludlow, W. B. N. Berry, written commun., 1968). *Monograptus* aff. *M. praehercynicus* from high in the formation indicates the *Monograptus uniformis* zone, considered by most geologists to be the lowest graptolite zone in the Devonian (McLaren, 1970).

PETES CANYON SEQUENCE

The oldest fossils found in the Roberts Mountains Formation in the Petes Canyon sequence are from the basal cherty unit about 75 feet (23 m) from the base of the formation. These fossils, collected by T. E. Mullens (USGS colln. D328SD), are *Climacograptus* sp., *Retiolites geinitzianus angustibens* Elles and Wood, and *Monograptus convolutus* (Hisinger) and are considered to be upper Lower Silurian (Llandovery) zone 22 or 23 of Elles and Wood (W. B. N. Berry, written commun. to T. E. Mullens, 1971). Another collection from somewhat higher in the formation contains (USGS colln. D272SD) *Crytograptus*? sp., *Monograptus vomerinus* cf. var. *gracilis* Ellis and Wood, *Monograptus* sp. (monograptid of the *M. prioden* group), and proximal ends of a monograptid of the *M. spiralis* group or a cyrtograptid of the *C. murchisoni* group, which suggest an early Wenlock age (W. B. N. Berry, written commun., 1969). Other collections from the platy limestone contain *Monograptus*? *discus, Monograptus dubius* type, *Monograptus colonus* type, and *Monograptus bohemicus* (McKee and Ross, 1969), representative of the middle part of the Silurian.

At one locality several medium to thin lenticular beds of limestone of unknown but presumed high stratigraphic position in the formation or possibly above it contain a coralline fauna of (USGS colln. 8089SD and M–1147) *Favosites* sp. (massive), *Striatopora* sp., *Pleurodictyum* n. sp., *Syringaxon* n. sp. *Australophyllum* sp. r. n. sp., *Levenea subcarinata* subspecies a, *Leptocoelia* sp., and *Kozlowskiellina* n. sp. and also stromatoporoids (massive) and the conodonts (USGS colln. 8837SD) *Belodella triamgularis* (Stauffer) (one), *Icriodus latericrescens* n. subsp. B? Klapper (seven), and *Panderous* sp. Both the shelly fossils and conodonts suggest an Early Devonian age and are most like the fauna from the Rabbit Hill Limestone (C. W. Merriam, 1968, and J. W. Huddle, 1971, written communs.). These limestone beds seem to be in normal stratigraphic sequence with thin argillaceous limestone beds bearing poorly preserved and unidentified monograptids.

JUNE CANYON SEQUENCE

Several fragmentary graptolites identified as *Monograptus* aff. *M. praehercynicus* by W. B. N. Berry (1970) were collected in the thin interval of Roberts Mountains Formation of the June Canyon sequence. This form, indicative of the *Monograptus uniformis* zone, normally occurs near the top of the Roberts Mountains Formation.

MILL CANYON SEQUENCE

Graptolites were collected from the base to the top of the formation, and corals are found in the upper half of the unit in the Mill Canyon sequence. A collection of graptolites from the lower 10 feet (3 m) of the formation contains *Monograptus testis* (Barrande) considered late Wenlock by W. B. N. Berry (written commun., 1970). Slightly higher but still near the base of the formation, the forms *Monograptus* cf. *M. pseudodubius* Boucek, *Monograptus* sp. (*M. ludensis* type), and *Gothograptus spinosus* (Wood) are present and represent latest Wenlock (*M. ludensis* zone; W. B. N. Berry, written commun., 1970). Another collection in the lower part of the formation contains *Monograptus colonus* (advanced form), *Monograptus chimaera, Monograptus bohemicus, Monograptus scanicus, Monograptus uncinatus*?, and a plegmatograptid and is early Ludlow, *Monograptus chimaera-M. scanicus* zone (W. B. N. Berry, written commun., 1971). Higher collections contain *Linograptus*? sp., considered by Berry to be latest Ludlow or younger, and in the uppermost part of the formation *Monograptus* aff. *M. praehercynicus* typical of the *M. uniformis* zone is found.

The Mill Canyon sequence has also yielded coral faunas representing zones A and D of the Great Basin coral zones of Merriam (1973). Coral zone A is represented by *Arachnophyllum kayi*, with which *Cyathophylloides fergusoni* and *Neomphyma crawfordi* are associated. Above this horizon coral zone D contains *Styopleura nevadensis, Tonkinaria* sp., and *Verticillopora annulate*.

The Roberts Mountains Formation in the Petes Canyon sequence correlates with the Roberts Mountains Formation in its type section on the west side of Roberts Creek Mountain and with the formation as described from the Antelope Valley region to the east. It includes strata the same age and some slightly older than Roberts Mountains Formation in the Mill Canyon sequence and is equivalent in age to the upper cherty part of the Gatecliff Formation of Kay and Crawford (1964), part of the overlying Bastille Limestone Member of the Masket Shale of Kay and Crawford.

MASKET SHALE OF KAY AND CRAWFORD (1964) (AUGUST CANYON SEQUENCE)

The name Masket Shale was first used by Kay in 1960, and later it was applied by Kay and Crawford (1964) to thin-bedded buff-weathering argillaceous limestone beds above the Antelope Valley Limestone and below the McMonnigal Limestone in the Mill Canyon sequence. These authors also applied the name Masket Shale to strata above their Gatecliff Formation in the August Canyon sequence. The formation in the August Canyon sequence was divided by Kay and Crawford (1964) into a lower member, which they named the Bastille Limestone Member, and an upper unnamed member. This breakdown of the Masket Shale in the August Canyon sequence is retained in this report and is applied to strata above the Gatecliff Formation of Kay and Crawford (1964). In contrast with Kay and Crawford, however, the name Masket Shale in this report is used only in the August Canyon sequence. The term Roberts Mountains Formation is used for partly correlative strata in the other sequences. (See "Roberts Mountains Formation.") The Masket Shale, including the Bastille Limestone Member and the upper unnamed member, crops out near the mouth of Mill Canyon and for about 2 miles (3.2 km) northward along the east front of the range.

BASTILLE LIMESTONE MEMBER OF KAY AND CRAWFORD (1964)

The Bastille Limestone Member consists of about 100 feet (30 m) of light gray medium- to thick-bedded dolomite overlain by about 25 feet (8 m) of medium- to thick-bedded limestone that contains abundant crinoids, corals, and brachiopods. Thin and medium limestone beds containing shelly fossils also occur intermittently in the dolomite sequence. The dolomite is similar in general appearance to dolomite in the underlying Gatecliff Formation, although the Bastille dolomites are faintly to strongly laminated in many outcrops. These laminations are parallel and even, and they resemble lamination in the Roberts Mountains Formation. The contact between the Bastille Limestone Member and the Gatecliff Formation is recognized by the presence of chert in the upper Gatecliff and the absence of chert in the lower Bastille. This contact seems unconformable. However, since no diagnostic fossils have been found in the upper cherty part of the Gatecliff Formation or in the dolomite of the lower part of the Bastille Limestone Member, it is not known if an unconformity exists at this contact. At some other localities in Nevada, an unconformity does occur at the equivalent contact at the top of the basal chert of the Roberts Mountains Formation. The upper contact of the Bastille is gradational and is drawn at the top of medium to thick beds of dolomite and limestone.

DEVONIAN SYSTEM

UPPER MEMBER, MASKET SHALE

The upper member of the Masket Shale consists of several different lithologies. These rock types include (1) medium- and thick-bedded limestone, with some beds having graded lower portions and fine-grained laminated or massive upper portions, (2) thin-bedded platy buff-weathering limestone that resembles the Roberts Mountains Formation at other places in the Toquima Range, and (3) medium- and thick-bedded blocky weathering fine-grained limestone. The first lithology is common in the lower part of the upper member, while the third lithology is dominant in the upper part of the upper member. The platy-weathering lithology, which Kay and Crawford (1964) emphasized in the upper member in the August Canyon sequence, occurs near the middle of the upper member and is the least abundant of the three lithologies.

The thickness of the upper member is unknown because the upper contact is everywhere a fault. As much as 400 feet (120 m) is present on the north side of Mill Canyon.

Megafossils are uncommon in most of the Bastille Limestone Member and the upper member of the Masket Shale. However, diagnostic fossils have been found in the thick limestone beds near the top of the Bastille Limestone Member about 115 feet (35 m) above the base of the formation and from the overlying thin-bedded limestone of the upper member about 300 feet (90 m) above the base of the formation. The collections of fossils from the Bastille Limestone Member contain *Favosites* sp. (massive form), *Coenites* sp., *Kyphophyllum* sp., large crinoid columnals, *Alveolites* sp., and *Altrypa*-like brachiopod fragments. These fossils are suggestive of Silurian coral zone E of Merriam (1973). Conodonts from the same limestone bed include *Spathognathodus inclinatus* (Rhodes) (three), *Spathognathodus* cf. *S. transitans* Bischoff and Sannemann n. subsp. (two), *Spathognathodus* sp. (one), *Gnamptognathus lipperti* (Bischoff) (one), *Hibbardella* sp. (one), *Hindeodella* sp. (one), *Ligonodina* sp. (one), *Ozarkodina* sp. (one), *Plectospathodus* sp. (one), *Prioniodina* sp. (one), and *Trichonodella* sp. (one). These species are considered by J. W. Huddle to indicate most likely an Early Devonian age (written commun., 1970).

Conodonts from the upper member of the Masket Shale about 300 feet (90 m) above the base of the formation and about 200 feet (60 m) above the collection listed above include *Belodella* sp. (two), *Spathog-*

nathodus remscheidensis Ziegler (seventeen), *Hindeodella* sp. (eleven), *Icriodus latericrescens* n. subsp. B Klapper (forty), *Oneotodus beckmanni* Bischoff (two), and *Ozarkodina denckmanni* Ziegler (three). These species are probably Early Devonian in age (J. W. Huddle, written commun., 1970).

The Masket Shale of Kay and Crawford (1964) in the August Canyon sequence is probably correlative with all the Roberts Mountains Formation and all the McMonnigal Limestone in the Mill Canyon sequence and with the entire Roberts Mountains Formation in the Petes Canyon sequence.

TOR LIMESTONE (JUNE CANYON SEQUENCE)

The name Tor was first applied by Kay (1960, figs. 3, 4) to limestone overlying the Antelope Valley Limestone in the June Canyon sequence. Later the name Tor Limestone was described by Kay and Crawford (1964) and shown on their geologic map. Additional study by McKee, Merriam, and Berry (1972) provided new information on the age and correlation of this formation and noted the presence of a small amount of Roberts Mountains Formation beneath it at one place.

The Tor Limestone is a massive light-gray to white coarsely crystalline rock which in places retains its original bioclastic texture. It forms prominent crags, peaks, and cliffs (fig. 10) well exposed in the upper reaches of Ikes and June Canyons and on the ridge between these canyons. It is almost everywhere in fault contact with rocks stratigraphically beneath it—in most places the Antelope Valley Limestone—but at one locality on the north side of Ikes Canyon the formation is in sedimentary contact with the Roberts Mountains Formation.

The original thickness of the Tor Limestone is not known because the upper contact is not found; it is the uppermost unit in the June Canyon thrust plate. About 200 feet (60 m) of the formation is exposed at the locality described above. Elsewhere, Tor Limestone having stratigraphic thicknesses of several hundred feet (about 100 m) or more are exposed in fault blocks, and it is estimated on the basis of the regional stratigraphic thickness of correlative units that the formation is about 500 feet (150 m) thick.

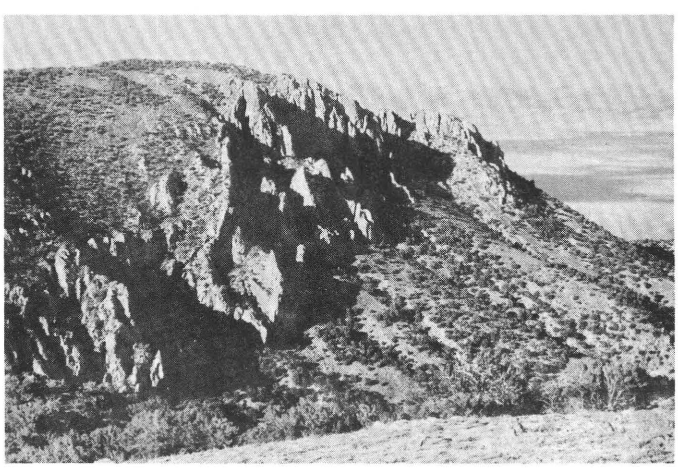


FIGURE 10.—Tor Limestone. This outcrop is on the ridge between Ikes and June Canyons. Here both lower and upper contacts are faults.

Corals collected several metres above the base of the formation include *Favosites* sp. and *Kyphophyllum* n. sp. cf. *K. lindstromi* Wedekind and are considered by C. W. Merriam to represent his Silurian coral zone E (in McKee and others, 1972). Conodonts from the same horizon include a large fauna of 15 genera (McKee and others, 1972) suggestive of an Early Devonian age (J. W. Huddle, written commun., 1970). Graptolites from the Roberts Mountains Formation a few metres beneath the Tor include *Monograptus* aff. *M. praehercynicus* and represent the *Monograptus uniformis* zone considered to be Early Devonian.

The Tor Limestone correlates with the lower beds of the McMonnigal Limestone of the June Canyon sequence and is mostly the same age (Silurian coral zone E of Merriam, 1973) as the upper part of the Roberts Mountains Formation at a number of places in Nevada. Although no megafossils younger than Silurian coral zone E of Merriam (1973) have been found in the Tor Limestone, it seems likely that beds high in the Tor may lie within Devonian coral zone A of Merriam (1974).

McMONNIGAL LIMESTONE (MILL CANYON SEQUENCE)

The name McMonnigal was first used by Kay (1960) for limestone at the top of the Mill Canyon sequence. Later, Kay and Crawford (1964) named this unit the McMonnigal Limestone, briefly described it, and outlined its distribution on a geologic map. More recently the fauna, age, and correlation of the McMonnigal were discussed by McKee, Merriam, and Berry (1972), and a specific type section was designated.

The unit is best exposed on a ridge forming the south side of Ikes Canyon about a mile (1.6 km) west of the range front. Here a 215-foot-thick (65 m) section, designated as the type section by McKee, Merriam, and Berry (1972), overlies the Roberts Mountains Formation. The basal contact is gradational across a stratigraphic interval of about 50 feet (15 m) and is drawn where medium-bedded gray fossiliferous limestone typical of the McMonnigal is dominant over thin-bedded platy-splitting limestone typical of the Roberts Mountains Formation. The upper contact is not exposed because the unit is uppermost in the Mill Canyon sequence. Several stratigraphic sections that can be pieced together suggest that the McMonnigal may have a thickness of approximately 300 to 400 feet (100–130 m).

A shelly fauna reported by McKee, Merriam, and Berry (1972) from the lower part of the formation includes species indicative of the Silurian coral zone E of Merriam (1973). Graptolites from platy limestone interbeds a short distance beneath rocks containing this fauna are *Monograptus* aff. *M. praehercynicus* and indicate the *Monograptus uniformis* zone. Seven conodont species from about the same horizon (McKee and others, 1972) are considered by J. W. Huddle to be of probable Early Devonian age (written commun., 1970). Higher in the formation occur corals and brachiopods indicative of the Devonian coral zone A of Merriam (1974) as well as the graptolite *Monograptus hercynicus,* considered to be Lower Devonian. A large conodont fauna from the same horizon includes 15 species (McKee and others, 1972) and is viewed as Lower Devonian by J. W. Huddle (written commun., 1970).

The lower part of the McMonnigal Limestone is correlative with the Tor Limestone of the June Canyon sequence and probably with upper beds of the Roberts Mountains Formation in the Antelope Valley region of east-central Nevada. Higher parts of the McMonnigal correlate with the Rabbit Hill Limestone of the Antelope Valley region. In areas northeast of the Toquima Range, the name Windmill Limestone (Johnson, 1965) has been applied to strata correlative with the lower part of the McMonnigal.

UNNAMED (DEVONIAN) LIMESTONE

Near the mouth of Ikes Canyon, 100 feet (30 m) or more of thin-bedded dark petroliferous-smelling limestone and black shale in about equal amounts that contains abundant styliolinids crops out in a fault block and may represent rocks that are stratigraphically above the McMonnigal Limestone in the Mill Canyon sequence. These rocks resemble parts of the Middle Devonian Woodpecker Limestone Member of the Nevada Formation in the Eureka region. Conodonts suggest that the Ikes Canyon rocks are of Middle Devonian age (J. W. Huddle, written commun., 1972) and include (USGS colln. 8952–SD) *Eantigognathus*? sp. (one), *Diplododella* sp. (one), *Hindeodella* sp. (four), *Ligonodina* sp. (one), *Panderodus* sp. (one), *Synprioniodina* sp. (two), and *Polygnathus kockeliana* Bischoff and Ziegler (five). The conodont fauna characterized by *Polygnathus kockeliana* is similar to that from the Denay Limestone of Johnson (1966) from the Roberts Creek Mountains.

OVERLAP ASSEMBLAGE

PENNSYLVANIAN SYSTEM

WILDCAT PEAK FORMATION

Limestone and conglomeratic to sandy limestone of the Wildcat Peak Formation lie unconformably on lower and middle Paleozoic rocks of both the western and eastern assemblages in the vicinity of the Ikes Canyon window. The name Wildcat Peak was first applied to these rocks by Kay (1960, fig. 4), and in 1964 Kay and Crawford expanded the name to Wildcat Peak Formation, described three sections with a fauna, and showed the distribution of these rocks on geologic maps (Kay and Crawford, 1964, pls. 1, 6). The thickest section, which occurs in Mill Canyon, was designated as the type section by these authors. The same general section is here considered the type; it lies in sec. 32 (unsurveyed), T. 14 N., R. 46 E., between the 8,800-foot contour and the top of the ridge (elevation point 9,814) on the south side of Mill Canyon (fig. 11*B*) in the

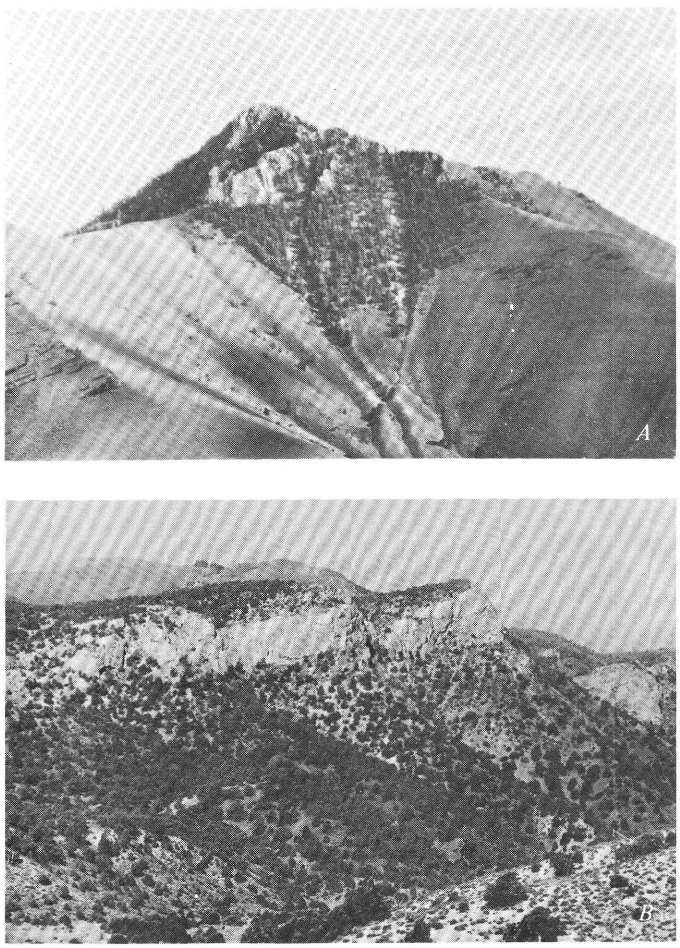


FIGURE 11.—Wildcat Peak Formation. *A*, Forming the peak at Wildcat Peak (reference section). *B*, On the south side of Mill Canyon to ridge top (type section).

Wildcat Peak quadrangle. Other excellent exposures of the formation, designated here as reference localities, include those that form Wildcat Peak (fig. 11*A*), the source of the name, and those on the south side of Mill Canyon near its mouth.

At the type section more than 900 feet (275 m) of limestone and conglomeratic limestone is exposed. The basal contact is an angular unconformity between massive brown-weathering limestone of the Wildcat Peak Formation and contorted thin-bedded shale and chert of the underlying Vinini Formation. This contact has a relief of about a metre in a distance of 100 feet (30 m) or less, and some angular clasts of the Vinini Formation are incorporated in the lowermost limestone of the Wildcat Peak Formation. The general sequence above the Vinini Formation consists of a lower massive fossiliferous limestone interval about 175 feet (55 m) thick, a middle part about 650 feet (200 m) thick of conglomeratic and sandy limestone with lenses of nearly pure sandstone, conglomerate, or limestone, and an upper part at least 150 feet (45 m) thick of limestone with subordinate lenses of sand or pebbles (fig. 12). All the units are lenticular, and sections only a few hundred metres apart vary considerably in thickness and stratigraphic detail.

Pebbles and cobbles as much as a foot (30 cm) across in the conglomerate are angular to subrounded chert and limestone similar to older Paleozoic rocks in the region (fig. 13). The carbonate matrix of the conglomerate weathers to a reddish color in places and at the locality on the north side of Ikes Canyon, the formation consists entirely of reddish soil containing boulders of conglomeratic limestone. Because of the extremely lenticular nature of the formation and the fact that it is found at separate isolated outcrop beds within the formation have not been traced laterally, and detailed stratigraphic relations have not been established for the unit as a whole.

Abundant brachiopods, corals, and bryozoans occur in massive limestone low in the formation at its type section on the south side of Mill Canyon. A representative collection from about 20 feet (6 m) above the base of the formation contains *Anthracospirifer occiduus* (Sadlick), *Neospirifer cameratus* (Morton), *Derbyia* sp., *Ditomopyge*? sp., *Stenoscisma* sp., *Barytichisma* sp., a productid brachiopod, and a fenestrate bryozoan and is considered by C. H. Stevens to be Early Pennsylvanian

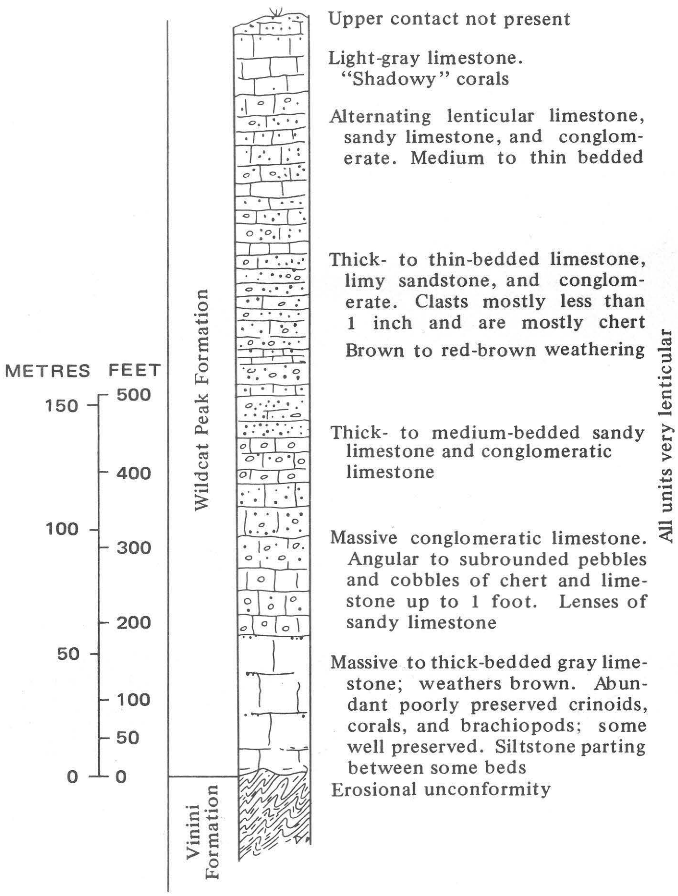


FIGURE 12.—Type section of the Wildcat Peak Formation, south side of Mill Canyon to ridgetop, sec. 32 (unsurveyed), T. 14 W., R. 46 E.

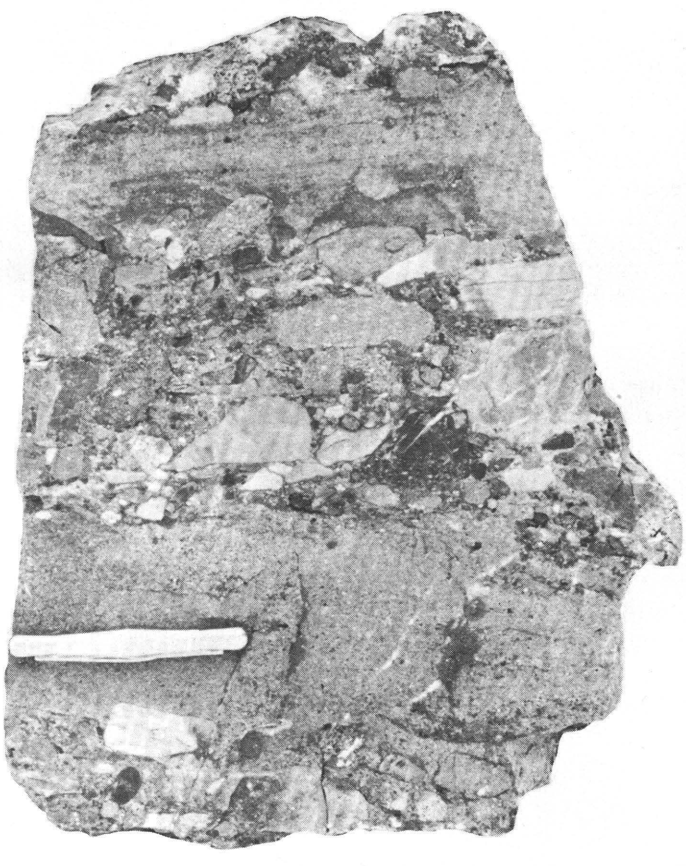


FIGURE 13.—Conglomerate and sandy limestone from the Wildcat Peak Formation. Pebbles are about equal amounts of chert and limestone.

(Atokan?) in age (written commun., 1970). The more than 900 feet (275 m) of strata above the horizon of this collection yielded no diagnostic fossils, although Kay and Crawford (1964, p. 443) reported Upper Pennsylvanian fossils from rocks on the south side of Mill Canyon; presumably these came from higher in the section.

A collection from an isolated outcrop of Wildcat Peak Formation on the ridge at the head of Ikes Canyon yielded the Late Pennsylvanian fusulinid *Triticites* aff. *T. secalicus* (Say) and the sponge *Rhombopara* sp. The stratigraphic relation between rocks of this outcrop and those of the thick Wildcat Peak section in the Mill Canyon area is uncertain. The Wildcat Peak Formation is believed to contain rocks spanning most of Pennsylvanian time.

The Wildcat Peak Formation is considered typical of conglomerates of the overlap assemblage in central Nevada as described by Roberts, Hotz, Gilluly, and Ferguson (1958). Rocks of this overlap assemblage include such conglomeratic units as the Late Pennsylvanian or Permian Brock Canyon Formation in the Cortez Mountains (Gilluly and Masursky, 1965), the Permian Garden Valley Formation of the Sulphur Springs Range (Nolan and others, 1956), and the Late Pennsylvanian and Early Permian Antler Peak Limestone of the Galena Range (Roberts, 1951) and the northern part of the Shoshone Range (Gilluly and Gates, 1965).

LOWER AND MIDDLE PALEOZOIC PALEOGEOGRAPHY

Reconstruction of the Ordovician, Silurian, and Lower Devonian paleogeography in central Nevada is, at best, imprecise and speculative. The discontinuous pattern of outcrops in widely separated ranges offers a minimum of stratigraphic control, and structural complications within the region must be deciphered before thorough paleogeologic interpretations can be made. Moreover, additional difficulties in interpretation result from abrupt facies changes that occur in the vicinity of the Toquima Range because the range is near the margin of the early and middle Paleozoic miogeosyncline. Carbonate units that can be traced with only slight lithologic change for more than 100 miles (160 km) across western Utah and eastern Nevada disappear in central Nevada where they are replaced by shale and chert representing deeper water deposits beyond the edge of the carbonate shelf. In theory the change from shallow to deep water occurs across a distance of only a few tens of kilometres and deposits from the slope separating the shelf from the basin are absent or very thin.

Regional thrusting during the Antler orogeny in Late Devonian and Mississippian time has obscured stratigraphic relations. Transport of siliceous strata many miles to the east of their site of deposition has been documented in central Nevada (Roberts and others, 1958). Tectonic telescoping on the Roberts Mountains thrust and related thrusts is difficult to unravel in more than the most qualitative manner. Some thrust plates moved tens of miles to the east, others may have moved farther (or less); also it is possible that transportation occurred in the opposite direction or from the north or south during thrust faulting younger than the Antler orogeny.

The four described sequences of lower and middle Paleozoic carbonate strata and the siliceous rocks of the same age in the northern part of the Toquima Range have been brought together by thrust faulting. The Petes Canyon sequence closely resembles correlative parts of the stratigraphic section at Antelope Valley about 30 miles (50 km) east of the Toquima Range. It is assumed that the Petes Canyon rocks are in place and not thrust from elsewhere. The August Canyon sequence about 20 miles (30 km) southeast of the Petes Canyon sequence is also assumed to represent rocks that are in place. The two sequences above the August Canyon sequence (the Mill Canyon and June Canyon sequences) are allochthonous and are presumed to have been transported from the west by thrusting. A cross-sectional reconstruction along a general east-west belt in central Nevada (pl. 2) includes the four sequences now found in the Toquima Range restored to their assumed original sites of deposition (pl. 2). The section at Antelope Valley serves as a reference and marks the eastern end of the reconstruction. East of Antelope Valley the strata remain fairly constant in thickness and lithology for long distances. The restored section is about 140 miles (225 km) long, 100 miles (160 km) of this representing shelf or miogeosynclinal rocks, 30 miles (50 km) of basin or eugeosynclinal rocks, and about 10 miles (16 km) representing the slope from shelf to deep basin. This hypothetical reconstruction indicates that eugeosynclinal and miogeosynclinal rocks were deposited only a few tens of kilometres apart and that transitional rocks are virtually absent. This interpretive picture offers an attractive alternative to the concept of extremely long tectonic transport on the Roberts Mountains thrust in all places and explains the failure to discover much truly transitional facies between eastern and western assemblage strata in central Nevada. A gradual westward transition within the carbonate (shelf) sequences is strongly suggested by thinning of rock units and transgression of time. Occasional periods of complete to partial emergence of land are recorded by the gaps in the record. Reef conditions represented by such units as the Tor Limestone developed at different times and places on the shelf. Only a small amount of fine detrital material collected on the broad shelf; some of it was incorporated in the carbonate rocks, and some of it was swept down the slope and is interbedded with the cherty rocks of the basin. It is postulated that the slope between miogeosyncline and eugeosyncline was at the approximate longitude of the Shoshone Mountains (117°15′ to 117°30′) in central Nevada (fig. 14). This hinge line trended in a general northerly direction but was undoubtedly sinuous.

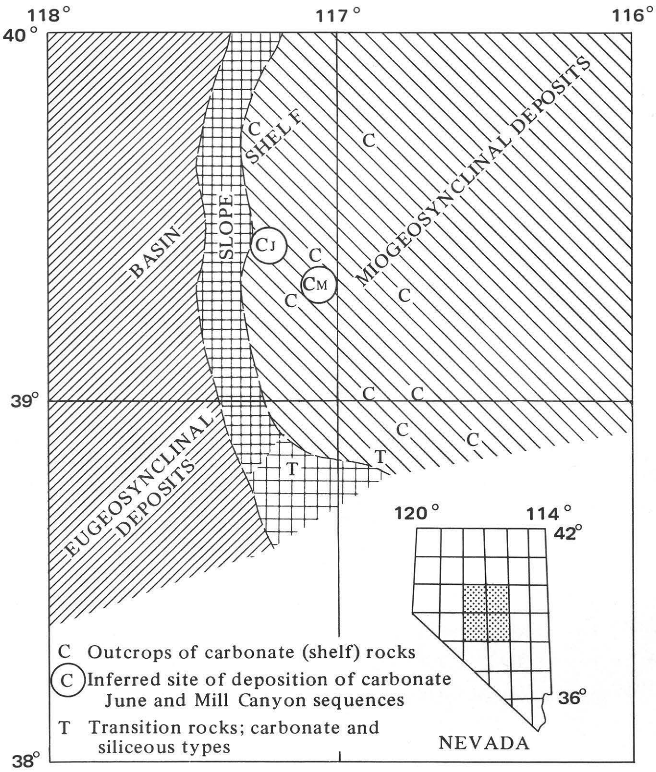


FIGURE 14.—Reconstruction of geosynclinal system in early and middle Paleozoic time in central Nevada. In most places in this area the slope between the miogeosyncline and the eugeosyncline is assumed to have been steep and narrow, so few rocks were deposited on it. In areas where the slope was wider and more gentle, a complete transition between carbonate and siliceous sediments occurred.

MESOZOIC ROCKS

JURASSIC SYSTEM

CLIPPER GAP PLUTON

A granitic body that crops out over about 12 square miles (27 km^2) on the west side of the range south of Clipper Gap Canyon (pl. 1) is here named the Clipper Gap pluton. This pluton has an oval outcrop pattern with a long axis trending north-south. It is exposed through a vertical interval of more than 2,000 feet (600 m) from the top of the range to the floor of Big Smoky Valley. The pluton is not connected at the present level of erosion with any of the large granite bodies in the region such as those in the Toiyabe Range, about 15 miles (24 km) to the west. A small outcrop of granitic rock, undoubtedly part of a larger buried body, protrudes from the alluvium in Big Smoky Valley near Spencer Hot Springs less than 10 miles (16 km) north of the Clipper Gap pluton, and a stock about 10 square miles (25 km^2) in area is located in Northumberland Canyon about 10 miles (16 km) south of the Clipper Gap pluton. These granitic bodies have no visible connection with the Clipper Gap pluton, but their generally similar radiometric ages (see section on "Age") suggest that they may be part of the same intrusive body.

CONTACT RELATIONS

The contact between the pluton and the Ordovician Vinini Formation exposed around about one-half the body on the east side is sharp and steep. Bedding in the Vinini Formation generally is steep and parallels the contact within about one-half kilometre of the pluton; beyond this it dips less steeply, although it retains a general parallelism to the pluton. The pluton seems to have been emplaced parallel to the regional grain of the rocks it intrudes but also seems to have modified and alined the bedding of the rocks adjacent to it. The configuration of pluton and country rock is similar to a knot in a piece of lumber. Pendants or large inclusions of country rock near the margin of the pluton seem to be generally alined with the contact. Dikes of aplite and diorite with various orientations occur in abundance near the contact. Foliation, lineation, shearing, or other obvious textural features that can be related geometrically to the contact are not obvious. The pluton is massive and homogeneous from its margin to the deepest level of dissection. Joints and linear crosscutting veins and dikes do not seem to be related to the margin, which they intersect at various angles.

The country rock adjacent to the pluton shows the effects of thermal metamorphism for a distance of several hundred metres from the contact. Rocks in this zone have hornfelsic to spotted hornfelsic texture. The rocks are extremely hard and recrystallized and in places appear to be silicified either by addition of silica from the pluton or by redistribution and cementation of silica originally in the rock. Pelitic rocks within a few metres of the contact have abundant slightly sericitized porphyroblasts of chiastolite as much as ½ inch (1¼ cm) long in a fine-grained hornfelsic matrix of quartz, graphite, and a little tourmaline. Farther from the contact similar rocks are quartz-graphite-muscovite hornfels. These mineral assemblages suggest a metamorphic grade ranging from the hornblende hornfels facies near the contact to the albite-epidote hornfels facies (Fyfe and others, 1958, p. 204–206) several tens of feet from the pluton.

JOINTS

There are two sets of pervasive joints in the Clipper Gap pluton. The best developed of these throughout the pluton strikes north to northwest and dips 70° to 85° E.; the second set strikes north to northeast and dips less than 20° W. (fig. 15). The steep joints are uniformly spaced a few metres apart; the gently dipping ones are about twice this distance apart. The pattern formed by the joints viewed from above is a series of equally spaced north-trending lines; viewed in an east-west cross section, they outline rows of rectangular blocks with the long dimension vertical. The gently dipping joints give a superficial bedded appearance to the granite in the central part of the pluton where they are especially well developed. The joints have no obvious relation to the shape of the pluton or to its margin except that the gently dipping ones are less pronounced near the margin and the strike of both sets parallels the long dimension of the pluton at its present level of erosion. The two sets mutually cut each other without offset, suggesting that they formed contemporaneously.

PETROLOGY

Most of the pluton is uniformly medium- to fine-grained biotite-hornblende granodiorite. Zones composed of alaskite or diorite and numerous aplite and diorite dikes and quartz veins make up less than 10 percent of the total. Samples collected from widely spaced outcrops representing near-margin and internal parts of the pluton are very similar, differing mostly in grain size. Quartz, plagioclase, potassium feldspar, biotite, and hornblende are recognizable in hand specimen; the microscope reveals magnetite, apatite, and chlorite as accessories. Modal analysis from thin sections and stained slabs show about 25 percent quartz, 40 percent plagioclase, 20 percent potassium feldspar, and about 15 percent biotite and hornblende in equal amounts. It is hypidiomorphic granular.

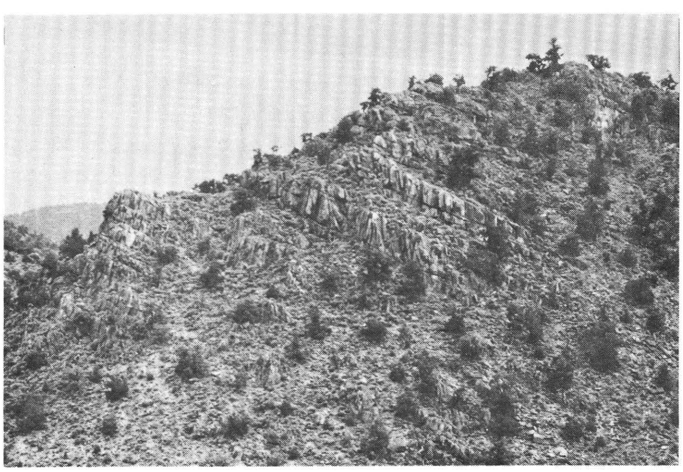


FIGURE 15.—Joints in the Clipper Gap pluton. Two sets are pervasive throughout the body: (1) widely spaced, gently dipping (looks like bedding) and (2) more closely spaced, steeply dipping.

Plagioclase forms euhedral to subhedral crystals, some of which are zoned. Their average composition on the basis of extinction angles is about an_{35} (andesine). Orthoclase occurs as anhedral crystals filling interstices between, or less frequently poikilitically enclosing, the other minerals. Much of the orthoclase is perthitic.

Hornblende and biotite are euhedral to subhedral and appear as ragged splinters or books intergrown with plagioclase or as isolated crystals enclosed by potassium feldspar. Both are pleochroic in differing shades of green, yellow, gold, and brown. The hornblende, which in most places is less altered, has a greater contrast in pleochoic hues.

Inclusions.—Cognate inclusions are rare. Those that were studied appear as spheroidal clots as much as 6 inches (15 cm) across of fine-grained hornblende quartz diorite or hornblende diorite and are scattered randomly throughout the granodiorite. The inclusions grade into the granodiorite host through a distance of about ½ to 1 inch (1–2½ cm). In this interval there is complete gradation in grain size and crystal content from the coarser grained lighter colored granodiorite to the dark fine-grained inclusions. The inclusions are discrete bodies, however, and not merely dark zones in the pluton.

Aplite dikes.—A number of small lenses and dikes of aplite cut the pluton as well as the country rock surrounding it. Most of these bodies are less than a half metre wide and two to three times this long, but some are 3 or more metres long and about a metre across. They consist of quartz, plagioclase, orthoclase, and biotite. Typically the host rock grades to aplite across a zone of several centimetres marked by decreasing hornblende and biotite and a transition in texture from medium- to fine-grained hypidiomorphic to fine-grained allotriomorphic granular. In the larger dikes the center of the aplitic body may be somewhat coarser grained than the margin. The microscope reveals an irregular mosaic of subhedral to anhedral plagioclase, anhedral potassium feldspar, quartz, and scattered jagged plates of biotite. The plagioclase includes subhedral oscillatory-zoned andesine similar to that in the main part of the pluton and gradationally zoned saussuritized albite. The albite, potassium feldspar, and quartz are intimately intergrown and have graphic texture in places.

Dacite dikes.—Dark-brown to gray highly altered dikes common throughout the entire region cut the pluton in places. These bodies are usually 3–30 m long and about a metre wide. Their genetic relation to the large granodiorite pluton is uncertain, but they are clearly younger and probably represent magmatic activity completely unrelated to the granitic intrusion.

CHEMISTRY

A chemical analysis of granodiorite from the Clipper Gap pluton is shown in table 1. This analysis shows that the rock is similar to the average granodiorite of Nockolds (1954, p. 1014, col. III).

AGE

The Clipper Gap pluton has been dated by the potassium-argon method at 151±3 m.y. (million years) (Silberman and McKee, 1971; table 2). This single biotite date suggests that the body is Middle Jurassic in age. Many plutons dated at 150–160 m.y. occur in the region. For example, the pluton at Austin, Nev., 30 miles (50 km) to the northwest is 157±6 m.y. old (Krueger and Schilling, 1971), the stock at Northumberland Canyon 10 miles (16 km) to the south is 154±3 m.y. old (Silberman and McKee, 1971), and the granitic rock that crops out near Spencer Hot Springs is about 160 m.y. old. (See "Granitic Rocks near Spencer Hot Springs.") The single biotite dates on these widely separated plutons are considered to be reliable ages of cooling because elsewhere in central Nevada where mineral pairs of hornblende and biotite have been dated, the ages are concordant (Silberman and McKee, 1971; Smith and others, 1971). In the northern part of the Toquima Range, there is no indication of more than one period of plutonism, and hence no resetting of ages.

TABLE 1.—*Chemical analyses and norms of the Clipper Gap pluton and a Cretaceous dike, Wildcat Peak quadrangle, Nevada*

	Clipper Gap pluton	Cretaceous dike
Chemical analyses (weight percent)		
SiO_2	67.3	67.9
Al_2O_3	15.8	15.8
Fe_2O_3	1.0	.59
FeO	2.2	2.1
MgO	1.6	1.1
CaO	3.1	2.1
Na_2O	3.6	3.9
K_2O	3.9	3.7
H_2O+	.73	1.3
H_2O-	.14	.37
TiO_2	.50	.42
CO_2	<.05	.51
P_2O_5	.14	.13
MnO	.06	.00
Total (rounded)	100	100
CIPW norms (weight percent)		
Q	21.66	25.46
or	23.05	21.86
ab	30.46	33.00
an	14.46	6.35
en	3.99	2.74
fs	2.50	2.68
il	.95	.80
mt	1.45	.86
ap	.33	.31
C	.36	3.05
Sum	99	97

TABLE 2.—*Potassium-argon ages of granitic rocks in the northern part of the Toquima Range*

Source and Rock	Mineral dated	K_2O[1] percent	Ar^{40}rad[2] (mole/g)	$\frac{Ar^{40}rad}{Ar^{40}total}$	Age (m.y.)
Clipper Gap pluton[3]; granodiorite.	Biotite	6.63	1.539×10^{-9}	.90	151±5
Spencer Hot Springs; quartz monzonite.	Biotite	7.68	1.903×10^{-9}	.897	160.7±5
Spencer Hot Springs; biotite quartz diorite porphyry.	Biotite	9.06	2.24×10^{-9}	.921	160.2±5

Constants used: $\lambda\epsilon = 0.585\times10^{-10}yr^{-1}$
$\lambda\beta \times 4.72\times10^{-10}yr^{-1}$
K^{40}/K total $= 1.22\times10^{-4}g/g$

[1]K_2O analyzed by Lois Schlocker using a flame photometer with a lithium internal standard.
[2]Ar analyzed by E. H. McKee by standard isotope dilution methods on a Neir-type, 6-inch 60°-sector mass spectrometer.
[3]See Silberman and McKee (1971).

GRANITIC ROCKS NEAR SPENCER HOT SPRINGS

A few small outcrops of granitic rock occur about 1 mile (1.5 km) east of Spencer Hot Springs in Big Smoky Valley and are undoubtedly part of a larger buried pluton. The main rock type is medium-grained porphyritic biotite quartz monzonite, but porphyritic biotite-hornblende quartz diorite and biotite quartz diorite porphyry are also present. Intrusive relations indicate that the biotite quartz diorite porphyry is the oldest rock and is probably an early phase of the main quartz monzonite pluton. The porphyritic biotite-hornblende quartz diorite is probably the youngest of the rocks and may be related to a wholly different period of intrusion. Because of the small area of outcrop and the much altered condition of the granitic rocks, little can be determined about internal structures or contact relations. Nearby Paleozoic strata (as much as about 100 metres from the granitic rock) have been metamorphosed to hornfelses and skarns exposed best at the Linka mine. These rocks consist of banded (relic bedding) types, many of which contain large crystals of garnet, epidote, and calcite. Thin sections reveal green amphibole, biotite, quartz, diopside, apatite, and probable idocrase. There is abundant opaque material, some of which is molybdenite. Scheelite was mined from these metamorphic rocks.

PETROGRAPHY

Much of the exposed granitic rock is weathered to grus, but several cohesive specimens indicate that the main rock is medium-grained porphyritic biotite quartz monzonite. Stained slabs show that 25 percent of the rock is quartz, 30 percent potassium feldspar, 30 percent plagioclase, and 15 percent biotite. The groundmass is medium-grained hypidiomorphic granular; the phenocrysts, which are not abundant, are large (as much as 5 cm) euhedral crystals of potassium feldspar, some of which are myrmekitically intergrown with quartz.

The porphyritic biotite-hornblende quartz diorite contains about 40 percent phenocrysts of zoned plagioclase, 5 percent biotite phenocrysts, and a few hornblende phenocrysts in a fine-grained allotriomorphic granular groundmass of potassium feldspar, quartz, and biotite.

The biotite quartz diorite porphyry is a distinctive lithologic type characterized by large euhedral books of biotite scattered throughout a very fine grained allotriomorphic granular groundmass of potassium feldspar, biotite, and quartz. Subhedral phenocrysts of zoned plagioclase are about as abundant as the biotite phenocrysts. Total phenocrysts make up about one-quarter of the rock.

AGE

Biotite from the quartz monzonite is 160.7±5 m.y. old as determined by potassium-argon methods (table 2). The biotite quartz diorite porphyry yielded an age of 160.2±5 m.y. on biotite (table 2). These ages suggest that the pluton is of Middle Jurassic age and that the two varieties of rock crystallized at about the same time. Intrusive relations show that the biotite quartz diorite porphyry is older, but the amount cannot be determined from the potassium-argon ages. The Middle Jurassic age of about 160 m.y. is similar to that determined on other plutonic bodies in the region. The significance of the age of these rocks is discussed in the section "Clipper Gap Pluton."

CRETACEOUS SYSTEM (FELSITE DIKES)

Rocks of Cretaceous age are represented in the northern part of the Toquima Range by at least some of the dikes that cut Paleozoic strata and the Jurassic Clipper Gap pluton (fig. 16). They are felsic and range from granodiorite to alaskite and include fine-grained and porphyritic types. These dikes occur in groups or swarms of five or ten, and most dikes within the group are a few metres wide and 3–30 metres long. They are generally steeply dipping and have a northerly trend. One swarm occurs directly north of the Clipper Gap pluton; another crops out on the ridge between North and South Wildcat Canyons. Almost all the dikes are somewhat altered and have an earthy cast, the feldspars being completely altered to sericite and the mafic minerals to limonite. In a few places, however, the dikes are fresh enough to reveal their original composition. A chemical analysis of one specimen is given in table 1, and a fission track age was determined from the same sample. This rock is a biotite granodiorite porphyry consisting of about 40 percent anhedral to subhedral plagioclase phenocrysts, 20 percent anhedral and deeply embayed quartz phenocrysts, and 10 percent biotite phenocrysts in a cryptocrystalline groundmass of potassium feldspar.

FIGURE 16.—Typical dike in the northern part of the Toquima Range (1 mile (1.6 km) northwest of the Branco mine). It is hornblende-biotite diorite to quartz diorite now altered and is about one-half mile (0.8 km) long and 10–15 feet (3–5) thick. Here it intrudes the Vinini Formation.

Apatite from the unaltered dike discussed above (table 1, No. 2) was dated by the fission track method and found to be about 93 m.y. old, suggesting a Cretaceous age. Analytical techniques used in sample preparation and track counting are similar to those described by Naeser and Dodge (1969, p. 2202–2204). Data obtained and used for the age calculation are flux on thermal neutron dose=1.28×10^{15}, number of spontaneous (fossil) tracks/cm^2=5.3×10^5, and number of induced tracks/cm^2=4.4×10^5. The uncertainty, α, of the age on the basis of the number of spontaneous, induced, and flux tracks counted is about 6 m.y.

The felsite dikes cut the Jurassic (152 m.y. old) Clipper Gap pluton and are overlain by Oligocene (31–33 m.y.) volcanic rocks, and so a general age bracket is available. The 93-m.y. age lies within this bracket and is a time of emplacement of many intrusive bodies throughout Nevada (see Smith and others, 1971; Silberman and McKee, 1971). For this reason the single Cretaceous fission track age is considered to be the age of emplacement of this and probably many other dikes in the area.

CENOZOIC ROCKS

TERTIARY SYSTEM

Tertiary rocks in the northern part of the Toquima Range are mostly welded ash-flow sheets. Lava flows, shallow intrusive bodies, and sedimentary units make up only a small percentage of the Tertiary section. Of the eight welded tuff units distingushed on the geologic map (pl. 1), two are considered to have been erupted from within the map area, and the other six originated from volcanic centers elsewhere in central Nevada. Many of these ash-flow sheets have been mapped in areas tens of kilometres apart and are recognized by lithology and relative stratigraphic position. Correlation of the sheets is substantiated here by radiometric age determinations, chemistry, and directions of natural remanent magnetization.

In general, most of central Nevada was a region of relatively subdued relief at the time of the first Tertiary volcanism, and ash flows spread across this region blanketing the existing topography. In places some of the flows were blocked or ponded, forming anomalously thick piles of tuff; in areas of high relief they are thin or were not deposited.

The succession of ash-flow eruptions in this region lasted for about 10 m.y. but was punctuated by long periods of quiessence. At such times erosion tended to strip the higher areas of their relatively thin cover of tuff. The unconformities thus formed coupled with the local volcanic and sedimentary units between the widespread tuff sheets create a complex stratigraphy that is difficult to summarize except in local areas.

The Tertiary volcanic sequence on the east side of the Toquima Range differs markedly from that on the west although at least one conspicuous welded tuff is common to both areas (figs. 17, 18). The central and northern part of the range is capped by the youngest Tertiary units in the region; they lie on several different older Tertiary rocks as well as on the Paleozoic basement in various places.

TUFF OF IRON SPRING

The remnants of a welded ash-flow sheet, here informally called the "tuff of Iron Spring," crops out in two places along the west margin of the range. One area of outcrops begins half a mile (0.8 km) east of Iron Spring and extends a mile and a half (2½ km) east along the north side of an unnamed canyon; the other outcrops occur about a mile (1.6 km) north of Iron Spring. The tuff, as much as 150 feet (45 m) thick, forms a wedge-shaped remnant beneath the overlying tuff of Hoodoo Canyon.

LITHOLOGY

The tuff of Iron Spring is mostly densely welded, although the basal 5 to 10 feet (1.5–3 m) is moderately to weakly welded, and locally the upper metre or so consists of weakly welded vapor-phase altered tuff. The densely welded rock is pink to gray lavender and knubbly weathering and has elongate gas cavities. It is poor in crystals, containing a few percent sanidine, quartz, and plagioclase. Lithic fragments of chert and andesitic types are ubiquitous and make up about 5 percent of the total rock. The tuff is probably a rhyolite, but chemical data are not available. The remanent paleomagnetic direction of this unit, measured in the field with a fluxgate magnetometer, is in the northern hemisphere (normally magnetized).

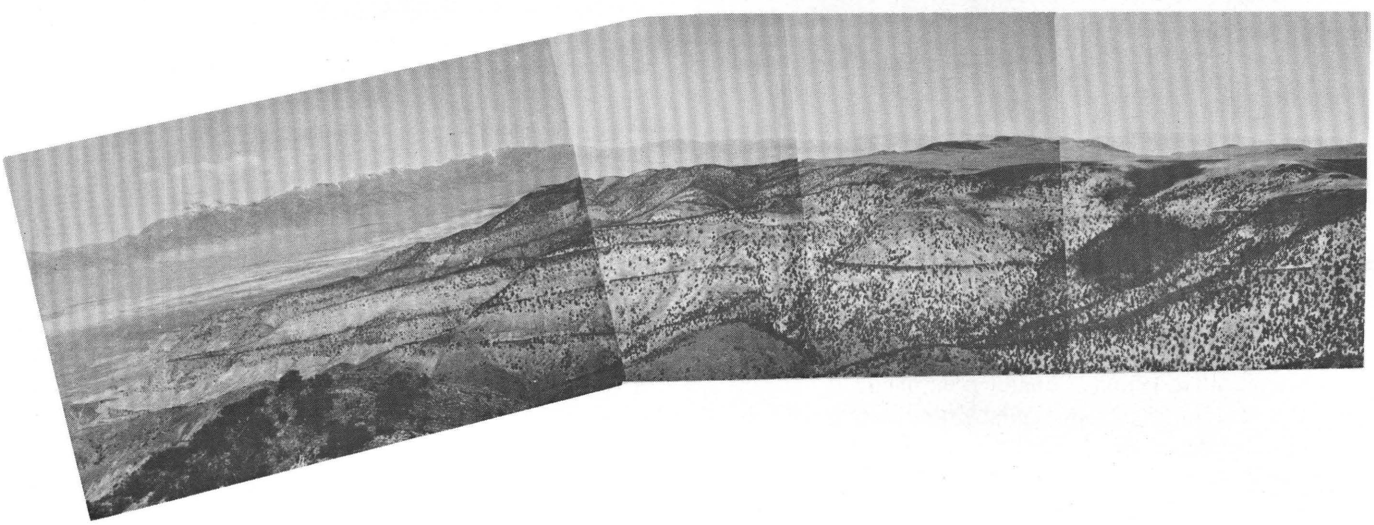


FIGURE 17.—Tertiary sequence on west side of Toquima range. Toiyabe Range in far background. Maximum thickness of the section is about 700 feet (215 m) and includes at least seven ash-flow cooling units. The upper three are units of the Bates Mountain Tuff; the lower four biotite-bearing tuffs include the tuff of Hoodoo Canyon.

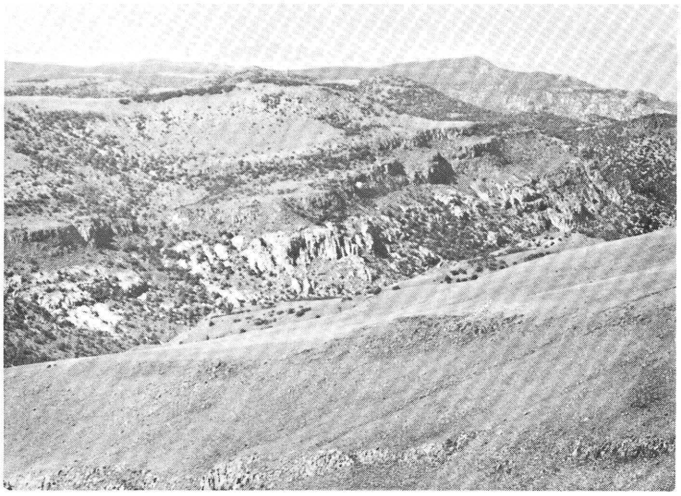


FIGURE 18.—Tertiary sequence on east side of Toquima range looking northeast down Stoneberger Canyon. Maximum thickness of the section is about 900 feet (300 m) and includes six ash-flow cooling units. The basal unit, the tuff of Stoneberger Canyon, is overlain by the tuff of Hoodoo Canyon, three cooling units composing the Bates Mountain Tuff, and the tuff of Clipper Gap.

AGE AND CORRELATION

The tuff of Iron Spring may be the oldest formation of Tertiary age in the northern Toquima Range, and it is certainly one of the oldest on the basis of it's position beneath the tuff of Hoodoo Canyon. Other units beneath the tuff of Hoodoo Canyon elsewhere are not in contact with the tuff of Iron Spring, and so stratigraphic relations cannot be seen. Compared with other remnants of tuff beneath the tuff of Hoodoo Canyon, the tuff of Iron Spring is the thinnest and may be the oldest. In addition it strongly resembles tuff of Dry Creek dated at 34 m.y. (Stewart and McKee, 1968), the oldest ash flow in the southern part of the Simpson Park Mountains, 20 miles (32 km) north of Iron Spring. It is possible that the tuff of Iron Spring and the tuff of Dry Creek are parts of the same unit.

NORTHUMBERLAND TUFF

The Northumberland Tuff crops out at the southwestern corner of the area and extends south at least 10 miles (16 km). It consists of a thick pile of welded rhyolite tuff (fig. 19) and a few rhyolite lava flows erupted from a volcano of Oligocene age (McKee, 1974) centered a few miles south of the area. This volcano developed a caldera, the north edge of which is well exposed in Northumberland Canyon. Present in the tuff are blocks of Paleozoic rock as much as a mile (1.5 km) across (fig. 20) that lie on the Northumberland Tuff and are remnants of large landslide masses that slid into the caldera while eruptions were taking place.

Rhyolite extending several hundred feet (about 100 metres) beneath the landslide blocks is highly altered and contains a veinwork of limonite-stained cracks (fig. 21). The degree of leaching, silicification, and iron staining is greatest directly beneath the Paleozoic blocks and has so altered the rock that it is difficult to recognize it as a crystal-rich rhyolite. At several

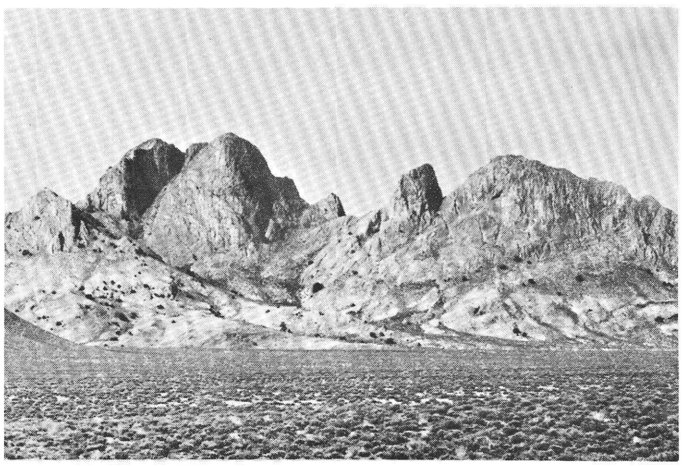


FIGURE 19.—Northumberland Tuff. View south from the mouth of Northumberland Canyon. The tuff here is more than 1,000 feet (300 m) thick.

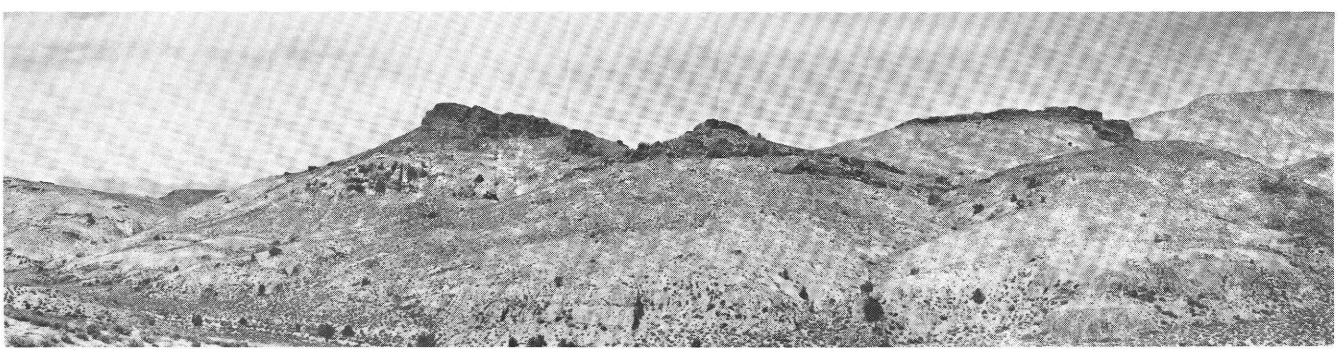


FIGURE 20.—Landslide blocks of Paleozoic rocks on top of Northumberland Tuff near the caldera edge. These blocks of bedded chert and shale (dark outcrops on the skyline) are about a square mile (2 km^2) or less in size and rest with parahorizontal contact on rhyolite of the Oligocene Northumberland Tuff. The rhyolite directly beneath the Paleozoic rocks and for about 200 feet (60 m) downward is much altered.

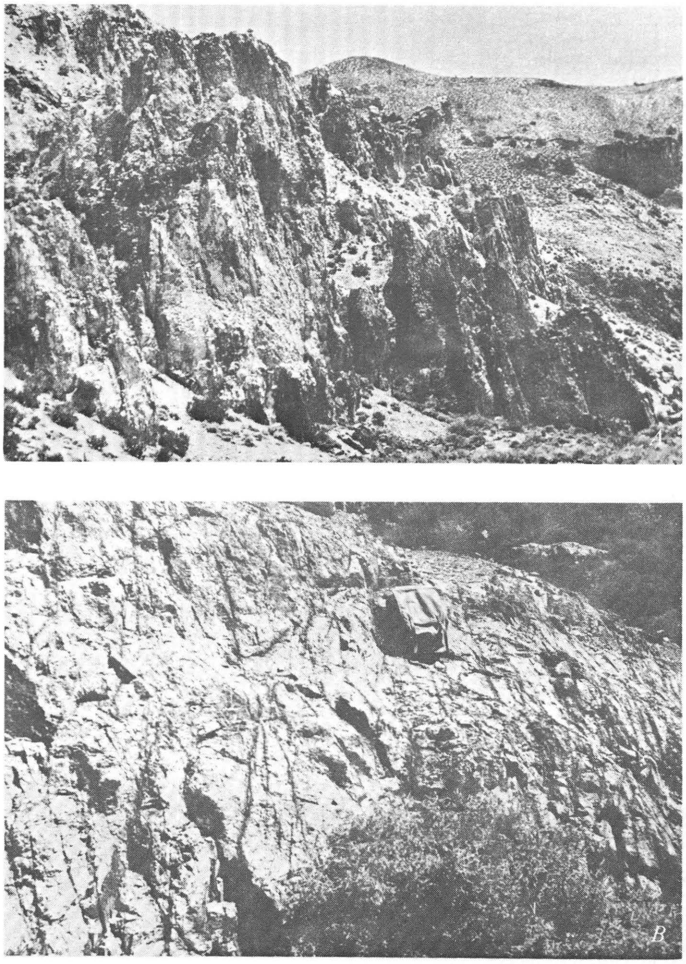


FIGURE 21.—Altered Northumberland Tuff. *A*, beneath a large block (fig. 22) of Vinini Formation. Vertical ribs are iron-rich zones, possibly columnar joints in the tuff. Chemical analysis of this altered tuff in table 3. *B*, Honeycomb veinwork of limonite alteration in Northumberland Tuff about 50 feet (15 m) beneath a large block of Vinini Formation (note backpack for scale).

localities where the landslide blocks of Paleozoic strata are separated like unjoined and somewhat disoriented pieces of a jigsaw puzzle, rhyolite has flowed onto the blocks and has filled cracks between them. On top of the Northumberland Tuff or directly on the Paleozoic landslide blocks where the rhyolite flows are missing, a sequence of stratified water-laid tuffaceous sedimentary rocks formed as a result of ponding of water within the collapse structure. These strata pinch out against the scarp or rim of the ancient caldera.

LITHOLOGY

The Northumberland Tuff is gray to white crystal-rich rhyolite, most of which is welded tuff of ash-flow origin. Tuff-breccia, in places rich in pumice, and laminated lavas are less common. Thin sections reveal that the tuff contains between 20 and 35 percent phenocrysts of quartz and sanidine in a devitrified glassy groundmass. This groundmass is composed of shards, some of which are bent around the phenocrysts, but most are oriented to form eutaxitic texture. In most specimens this texture is only apparent in thin section; megascopically the rock appears structureless. The microscope also reveals that most of the phenocrysts are cracked, and many are fragments of larger crystals. The quartz is typically resorbed. Biotite is uncommon but locally makes up as much as a percent or two of the total rock. Plagioclase was noted in several thin sections but is uncommon. Rock fragments, mostly a centimetre or less in size, are ubiquitous and numerous in the rock and are mostly Paleozoic chert or argillite.

CHEMISTRY

Chemical analyses (table 3) of the Northumberland Tuff reveal, like the modal content, that the rock is rhyolite. The analyses are similar to the "average calc-alkali rhyolite and rhyolite-obsidian" of Nockolds (1954), although they are poorer in total iron oxides and magnesia. The triangular plot of normative albite, anorthite, and orthoclase (fig. 22) shows that the three samples are within the rhyolite part of the diagram as defined by O'Conner (1965, fig. 3) for volcanic rocks from southern Nevada. Within the rhyolite field, a subfield delineated by eight analyses of two rhyolite welded tuffs given by O'Conner is shown. Two samples of the Northumberland Tuff lie within this subfield; the other, which is probably enriched in potassium, lies toward the orthoclase part of the rhyolite field. A fourth analysis of highly altered Northumberland Tuff from about 30 feet (10 m) beneath a large block of Ordovocian Vinini Formation resting on the tuff is included to show the loss or enrichment of various oxides near the top of the thick composite ash-flow sheet. A more than tenfold increase in total iron in the form of

TABLE 3.—*Chemical analyses of four samples of Northumberland Tuff. One sample is of altered tuff from beneath a large landslide block of Vinini Formation*

	Unaltered samples			Altered sample	Change between unaltered and altered samples
	No. 1	No. 2[1]	No. 3		
SiO_2	73.2	75.4	75.6	65.5	Loss (probable).
Al_2O_3	14.3	12.6	12.9	10.1	Do.
Fe_2O_3	.27	.07	.29	11.8	Gain.
FeO	.40	.12	.26	.28	Do.
MgO	.14	.11	.19	.18	
CaO	1.1	.47	.94	1.2	
Na_2O	3.4	1.5	3.0	1.0	Loss.
K_2O	5.4	7.4	4.9	4.2	Do.
H_2O+	.69	1.2	.49	3.4	Gain.
H_2O-	.51	.23	.71	.57	
TiO_2	.22	.20	.25	.25	
CO_2	.05	.05	.05	.58	Do.
P_2O_5	.03	.02	.00	.10	Do.
MnO	.02	.00	.00	.00	

[1]This sample is probably enriched in potassium.

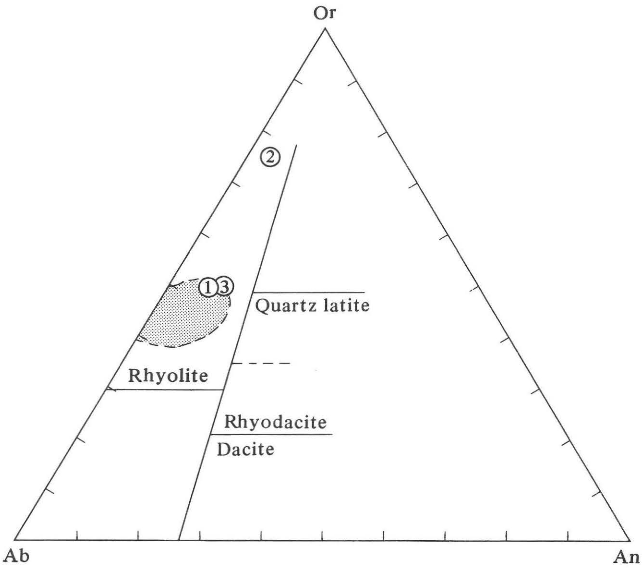


FIGURE 22.—Normative molecular albite, anorthite, and orthoclase. Normative classification of O'Conner (1965, fig. 3). Shaded area is the approximate field defined by eight samples of two rhyolite ash-flow tuffs from southern Nevada shown in O'Conner. Numbers 1, 2, and 3 are Northumberland Tuff. Number 2 shows potassium enrichment.

Fe_2O_3 is the greatest change, total alkalies are depleted by almost half and SiO_2 and Al_2O_3 are probably depleted as well. The bottom of the block of Vinini Formation and smaller isolated pieces of this formation are strongly bleached and stained by the vapors responsible for the alteration of the tuff.

AGE

A potassium-argon age of 32.3 m.y. (table 4) determined on sanidine separated from a sample of the Northumberland Tuff suggests that the unit is of Oligocene age. This sample, from the north side of the caldera, was probably emplaced about midway in the eruptive cycle of the volcano, and its age is considered average for the rocks from the center. The initial ash flows were probably erupted a million years earlier or less, and the last volcanism about the same amount of time later. A duration of vocanism on the order of 1 million years seems reasonable considering the known age spread of different parts of ash-flow sheets (Kistler, 1968; McKee, 1970) or the age of genetically related flows from a volcanic center (for comparison see Noble and others, 1968; Byers and others, 1968; Lipman, and others, 1970; Christiansen and Blank, 1972). A limiting upper age is furnished by the overlying tuff of Hoodoo Canyon, which yielded an average potassium-argon age of 30.4 m.y. (table 4). This tuff lies directly on the Northumberland Tuff in a few places but is separated from the latter unit by varying amounts of sedimentary strata in most places.

NATURAL REMANENT MAGNETIZATION

Measurement of the remanent magnetic direction in the field with a fluxgate magnetometer indicates that at least part of the unit has reverse magnetization, that is, the natural remanent direction is toward the southern hemisphere. The rock is extremely weakly magnetized in many places, however, and confirming measurements were ambiguous. It is possible that different parts of the tuff pile have different remanent magnetic directions because of the probable length of time required for emplacement of the composite unit.

ERUPTIVE AND TECTONIC HISTORY

Eruption about 33 m.y. ago of low-energy ash flows from vents in a shallow magma chamber located immediately south of Northumberland Canyon formed a volcanic pile and spread for about 15 miles (5 m) in a southerly direction, marking the start of volcanism at this volcanic center (fig. 23*A*). After a considerable volume of tuff had erupted, collapse of the volcanic

TABLE 4.—*Potassium-argon ages with analytical data of Tertiary volcanic rocks in the northern part of the Toquima Range*

Name of unit	Location (see fig. 1)	Mineral dated	K_2O percent	Ar^{40}rad (mole/g)	Ar^{40}rad / Ar^{40}total	Apparent age (m.y.)[1]
Tuff of Clipper Gap	1	Sanidine	8.17	2.69×10^{-10}	0.67	22.2
Bates Mountain Tuff	2	do	10.47	3.69×10^{-10}	.84	23.7
Unnamed rhyolite dome	6.7 miles (10.7 km) south of Northumberland Canyon; long 116°56′42″ E., lat 38°54′48″ N.	Biotite	8.30	3.42×10^{-10}	.78	27.7
Tuff of Hoodoo Canyon	3	do	7.68	3.50×10^{-10}	.71	30.6
do	4	do	7.58	3.39×10^{-10}	.54	30.1
Tuff of Stoneberger Canyon	5	Sanidine	9.96	4.61×10^{-10}	.85	31.1
Pancake Summit Tuff	6	do	12.09	5.63×10^{-10}	.91	31.3
Northumberland Tuff	7	do	10.20	4.91×10^{-10}	.92	32.3

Constants used:

$\lambda_\epsilon = 0.585\times10^{-10}yr^{-1}$

$\lambda_\beta = 4.72\times10^{-10}yr^{-1}$

K^{40}/K total $= 1.22\times10^{-4}g/g$

[1]Analytical uncertainty is between 2 and 3 percent of the age. This value represents the cumulative uncertainties in Ar and K analyses and involves such factors as Ar^{38} spike calibration, measurement of isotopic ratios, and flame photometer standards. (See Cox and Dalyrmple (1967) for discussion.)

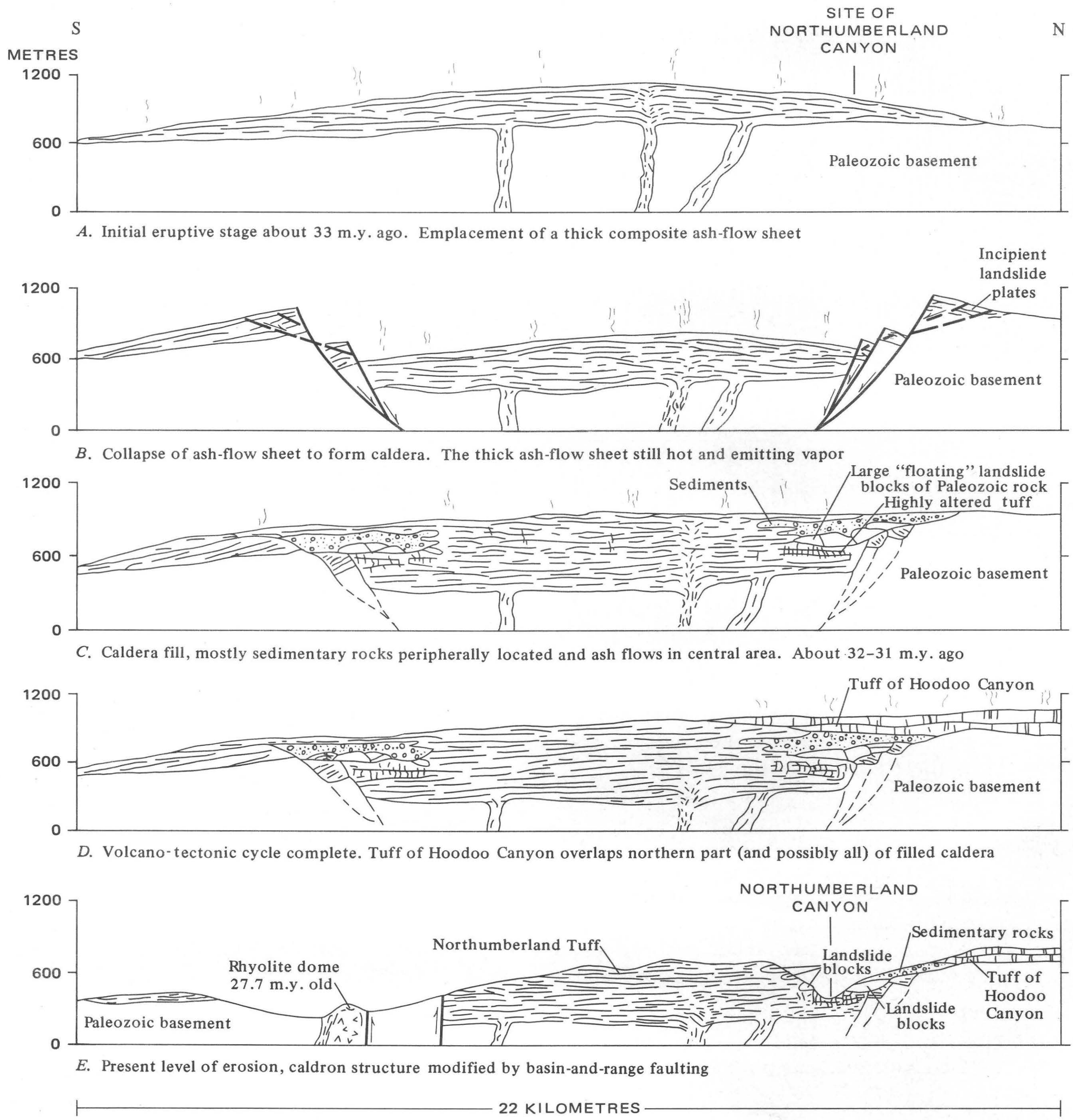


FIGURE 23.—Schematic history of the Northumberland Canyon volcanic center and caldera. Horizontal and vertical scale approximate. *A*, Initial eruptive stage about 33 m.y. ago. Emplacement of a thick composite ash-flow sheet. *B*, Caldera collapse and landsliding at rim. The thick ash-flow sheet still hot and emitting vapor. *C*, Caldera fill, mostly sedimentary rocks in moat area and ash flows in central area. About 32–31 m.y. ago. *D*, Volcano-tectonic cycle complete. Tuff of Hoodoo Canyon overlaps northern part of filled caldera about 30.5 m.y. ago. *E*, Present level of erosion.

center resulted in formation of a caldera (fig. 23*B*). Blocks of Paleozoic strata as much as a mile (1.6 km) across broke from the caldera rim and slid onto the hot ash-flow sheet inside the caldera, some partly sinking into the tuff. Beneath these barriers of relatively impermeable rock, hot vapor migrating upward caused alteration of the tuff and leaching, discoloration, and silicification of the bottom part of the blocks. Continued volcanic activity filled the central part of the caldera and added to the wedge of clastic detritus, including the landslide blocks, that filled the peripheral part of the caldera (figs. 23*C*, 24). Postcollapse ash

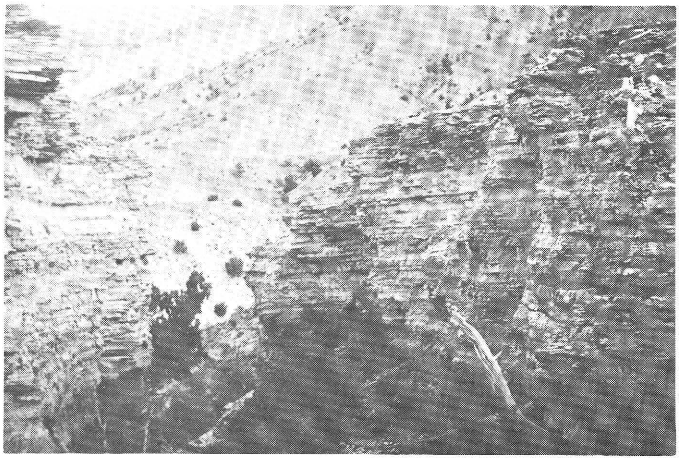


FIGURE 24.—Tuffaceous sedimentary strata that accumulated in the north edge of the collapse caldera after eruption of the Northumberland Tuff.

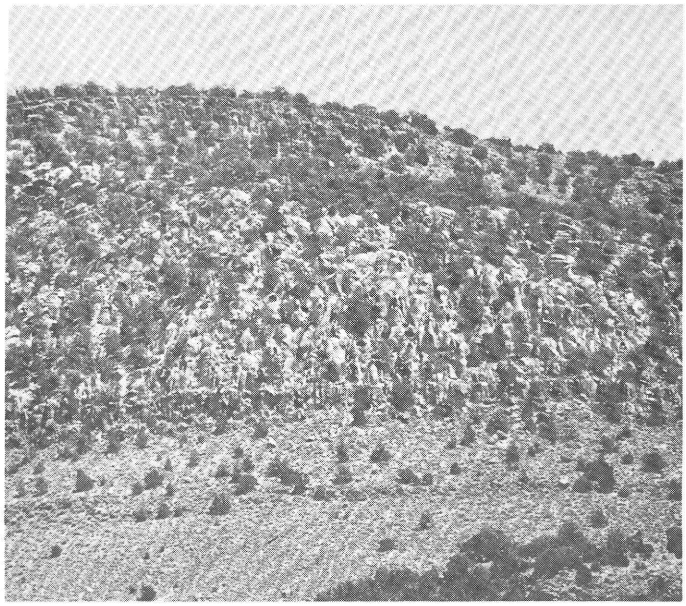


FIGURE 25.—Pancake Summit Tuff on the east side of the range north of the road from Big Smoky Valley to Monitor Valley.

flows accumulated to a thickness of about 1,000 feet (300 m) in the center of the caldera, and sedimentary rocks to a thickness of several hundred feet (about 100 m) around the margin.

PANCAKE SUMMIT TUFF

The Pancake Summit Tuff occurs mainly northeast of Petes Summit in the north-central part of the area, but it also crops out 3 to 5 miles (5–8 km) southeast of Petes Summit (pl. 1). Especially picturesque outcrops of this welded ash flow are found on the east side of the range north of the road from Big Smoky Valley to Monitor Valley (fig. 25). Here the unit is as much as 600 feet (200 m) thick, but it is absent less than 4 miles (6.5 km) both north and south of this thick section. The lenticular nature of the tuff is due primarily to erosion of the soft nonwelded upper part before deposition of younger formations. Most of the Pancake Summit Tuff is hard, densely welded, and resistant to erosion, but basal parts locally are weakly welded and soft; half or more of the unit is soft moderately welded tuff on the west side of the range near Henry Meyer Canyon.

This ash-flow sheet is widespread in central Nevada, extending from the Toquima Range on the west to the White Pine Range east of Eureka. An estimate of its maximum lateral extent by Grommé, McKee, and Blake (1972) is 120 km, and it covers approximately 5,400 km^2. Its estimated volume of 750 km^3 (Grommé and others, 1972, p. 1621) places it as one of the largest ash-flow sheets in Nevada.

LITHOLOGY

The Pancake Summit Tuff is light gray to pink, crystal rich, and made up of about equal amounts of smoky quartz, sanidine, and plagioclase in a groundmass of devitrified shards. About a third of the rock consists of phenocrysts. Biotite is ubiquitous but accounts for no more than about 1 percent of the total rock. A few small rock fragments were noted in several thin sections. Textural features typical of welded tuffs are not obvious in hand specimens, which appear homogeneous and structureless. Thin sections, however, reveal well-developed eutaxitic texture, angular and cracked quartz, feldspar crystal fragments, and shredded biotite flakes alined in the eutaxitic fabric and bent around the angular quartz or feldspar crystals in a manner similar to the shards. Gas cavities are locally present but are not characteristic of this tuff. In some places where the tuff is not densely welded, it contains slightly flattened to nonflattened pieces of pumice.

CHEMISTRY

The Pancake Summit Tuff is rhyolite on the basis of one whole-rock analysis from the Toquima Range (table 5) and several analyses of the rock from localities south of Eureka, Nev. (M. C. Blake, Jr., unpub. data). It compares closely with the average calc-alkali rhyolite and rhyolite obsidian of Nockolds (1954) and is similar to many of the rhyolite ash-flow tuffs of central and eastern Nevada. Chemically and lithologically it is most similar to the Windous Butte Formation of eastern Nevada, which is also about the same age. These two ash-flow sheets are so similar in most aspects that they were not distinguished until the detailed remnant magnetic study by Grommé, McKee, and Blake (1972) indicated that two distinct flow units exist.

TABLE 5.—*Chemical analyses, in weight percent, of Tertiary ash-flow tuffs from the northern part of the Toquima Range*

[Bates Mountain Tuff comprises units B, C (lower and upper), D, and the tuff of Clipper Gap; all samples were collected at Clipper Gap Canyon]

	Pancake Summit Tuff	Tuff of Stoneberger Canyon	Tuff of Hoodoo Canyon	Unit B	Unit C		Unit D	Tuff of Clipper Gap
					Lower cliff	Upper cliff		
SiO_2	74.9	76.0	65.1	71.0	73.6	75.0	73.3	75.3
Al_2O_3	13.2	12.7	16.2	12.9	12.9	13.2	13.0	12.4
Fe_2O_3	.35	1.0	1.3	1.4	.85	.76	1.6	1.2
FeO	.68	.06	1.8	.60	.32	.40	.52	.40
MgO	.13	.27	1.2	.23	.16	.19	.08	.36
CaO	.85	.60	3.0	.50	.50	.91	.31	.75
Na_2O	3.6	3.3	3.4	2.9	3.0	3.3	4.2	3.3
K_2O	5.1	4.6	5.1	5.8	4.8	4.8	5.2	4.7
H_2O^+	.22	.90	2.2	.53	.21	.70	.49	.72
H_2O^-	.65	.02	.55	3.5	2.8	.40	.71	.68
TiO_2	.13	.07	.47	.16	.11	.10	.19	.12
CO_2	<.05	.05	<.03	<.05	<.05	<.05	<.05	<.05
P_2O_3	------	------	.10	-----	----	.06	.04	.03
P_2O_5	.02	------	------	-----	----	----	----	----
MnO	.04	------	.05	.20	.06	.06	.05	.00
Total (rounded)	100	100	100	100	99	100	100	100

STRATIGRAPHIC RELATIONS

In most stratigraphic sections that contain the Pancake Summit Tuff, it is the basal Tertiary unit in the northern Toquima Range. Only at one place south of Stoneberger Canyon is a small area of Tertiary dacite lava beneath this ash flow. Elsewhere the ash-flow sheet rests unconformably on a variety of rock units, mostly formations of early Paleozoic age. Deposition on an irregular erosion surface accounts for some variation in the thickness of this tuff, but most of the apparent thinning of the formation is due to erosion before emplacement of later Tertiary ash flows on top of the formation. The complex stratigraphy resulting from vigorous erosion, local subareal deposition, and periodic emplacement of widespread ash-flow sheets is generalized in the stylized cross section of figure 26, which illustrates relations that could not be shown at the scale of the geologic map (pl. 1).

Correlation of the formation from the Toquima Range in the west eastward to localities in the western part of White Pine County is based primarily on potassium-argon ages and correspondence of the remanent magnetic direction (Grommé and others, 1972). The Pancake Summit Tuff is easily confused with the Windous Butte Formation, but these tuffs can be distinguished with certainty by the direction of their natural remanent magnetism. This direction is toward the southern hemisphere for both ash-flow sheets, differing by about 20° inclination and 30° declination (Grommé and others, 1972). To distinguish this difference requires data from many oriented samples, elimination of secondary magnetic components by demagnetization, and precise measurement of the primary natural magnetic direction. In most places where only standard field mapping supplemented by modal and chemical analyses are available, the Pancake Summit

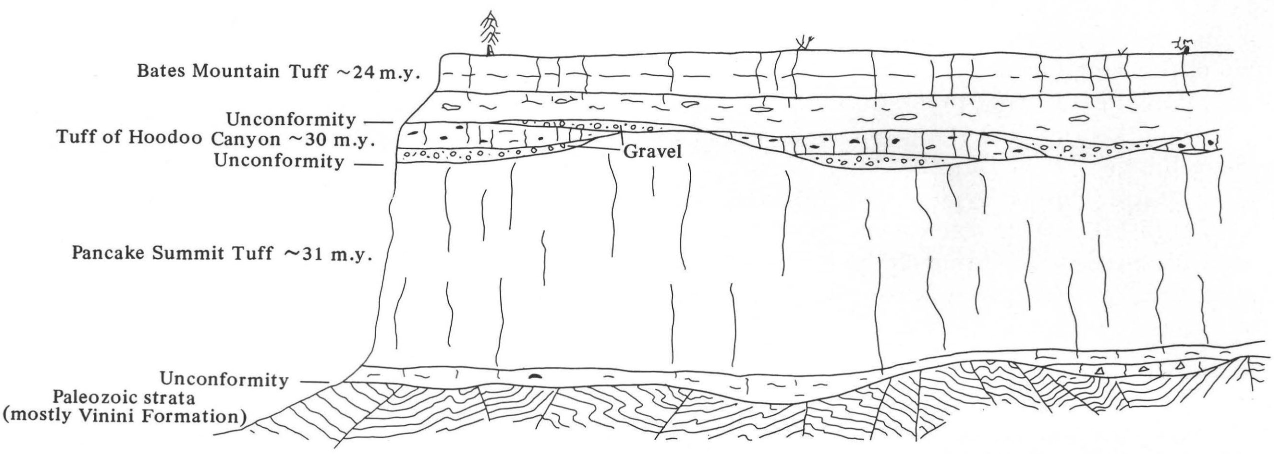


FIGURE 26.—Stylized cross section of the Tertiary rocks in the area near Petes Summit.

Tuff and Windous Butte Formation cannot be differentiated.

A potassium-argon age determination on sanidine from a devitrified densely welded sample of the Pancake Summit Tuff is 31.3 m.y. (table 4). On the basis of this and several other radiometric dates reported by Grommé, McKee, and Blake (1972, table 5) that average 32 m.y., it is concluded that this ash-flow sheet is of Oligocene age.

TUFF OF STONEBERGER CANYON

Tuff and tuff-breccia that form strikingly scenic cliffs at the mouth of Stoneberger Canyon are here informally called the "tuff of Stoneberger Canyon." This formation occurs only between Sams Canyon and Ikes Canyon on the east side of the range and is about 800 feet (250 m) thick at the mouth of Stoneberger Canyon and somewhat thicker a few miles to the east in the upper reaches of Ikes Canyon. It is an extremely lenticular body of rock that was erupted from within this area and accumulated near its site of eruption. As such it constrasts with the regionally extensive ash-flow sheets that spread across large parts of central Nevada.

The formation is typically composed of massive weakly to nonwelded punky tuff-breccia; however, it is surprisingly resistant to erosion and forms cliffs and bold outcrops. The thicker exposures exhibit columnar jointing, and variations in hardness from compaction or welding give the unit a slightly stratified aspect. No lithologic breaks, cooling breaks, zones of vitrophyre, vapor-phase alteration, or significant differences in welding are present to indicate that more than one ash flow is present. In places, as much as half the formation consists of angular pieces of different lithic types, mostly chert of the Vinini Formation, the dominant basement rock in the vicinity, and locally the tuff is composed of a jumble of lithic debris, angular pumice, and fragments of crystal tuff in a tuff matrix. These breccia zones may outline vents from which the tuff was erupted.

The unit thins abruptly westward and pinches out in the area that is now the crest of the range west of White Rock Mountain. On the east side of the range, the formation is the basal unit of the Tertiary section and is overlain by tuff of Hoodoo Canyon. West of the range crest the formation is absent and the tuff of the Hoodoo Canyon is the basal unit. Clearly, a divide located at about the present range crest existed at the time of eruption and acted as a barrier on the west. Figures 17 and 18, photographs taken from the present crest on the ancestral divide, show the different sections of Tertiary units on the west and east sides of the range. Similar thinning, although not so obvious or pronounced, occurs in a north-south direction from the thickest part of the tuff pile.

LITHOLOGY

Tuff of Stoneberger Canyon is typically a white to pale-lavender structureless crystal-rich lithic-bearing tuff-breccia. It is punky and fairly soft in most places but locally is fairly dense and hard. Crystals, which make up about 25 percent of the rock, are smoky quartz, sanidine, and some plagioclase. All are cracked and broken fragments, some of which are as large as 10 mm but most about half this size. The large smoky quartz crystals set in the white groundmass give the rock a distinctive speckled appearance. The groundmass consists of devitrified shards that show very faint to no eutaxitic orientation. Lithic fragments are as much as 10 cm across, but most are 2 cm or less in size; they are angular fragments of dark chert and dacite lava. This xenolithic debris is ubiquitous, and is locally very abundant, making up as much as half of the total rock. Large fragments of pumice, mostly unflattened or only slightly so, also occur.

CHEMISTRY

A chemical analysis of a sample of moderately hard and slightly welded tuff of Stoneberger Canyon is shown in table 5. The analysis, like the phenocryst content, indicates that the tuff is a rhyolite. The rock contains significantly more silica and less total iron than the average calc-alkali rhyolite and rhyolite-obsidian of Nockolds (1954) and can best be described as a high silica rhyolite.

AGE AND NATURAL REMANENT MAGNETIZATION

A potassium-argon age of sanidine from a moderately dense and hard sample of the tuff of Stoneberger Canyon is 31.1 m.y. (table 4), which is Oligocene on the standard geologic time scale of Harland, Smith, and Wilcock (1964). This age is the same within analytical uncertainty as that of the underlying Pancake Summit Tuff and of the overlying tuff of Hoodoo Canyon.

Fluxgate magnetometer measurement of the remanent magnetic direction from four samples at different locations indicates that the tuff of Stoneberger Canyon has reverse (southern hemisphere) natural remanent magnetization.

TUFF OF HOODOO CANYON

A distinctive ash-flow tuff, here informally called "tuff of Hoodoo Canyon," crops out widely in the southern part of the area and forms excellent exposures in the canyon after which it is named. It is an extremely useful marker in the Tertiary section as the rock contrasts with other ash-flow tuffs in the area by containing as much as 10 percent biotite. It is probable, however, that there are several biotite-rich ashflows in the northern part of the Toquima Range at about the same relative stratigraphic position.

The unit described here as tuff of Hoodoo Canyon can be traced almost continuously in the southern part of the area with thickest outcrops on the west side of the range in Wildcat and Hoodoo Canyons (fig. 27). In this area it is more than 400 feet (125 m) thick but thins abruptly in all directions, and it is represented by a uniform sheetlike body less than 100 feet (30 m) thick in the central part of the range and discontinuous pockets and lenses elsewhere in the region. The area in which it reaches its maximum thickness may be an eruptive source or an ancestral basin in which it accumulated to a great thickness. The thin and discontinuous remanents of the unit are due mostly to erosion before emplacement of the next younger ash-flow sheet, the Bates Mountain Tuff.

The tuff of Hoodoo Canyon is a typical welded ash-flow sheet. The basal part of the sheet is soft and nonwelded; the middle, which comprises about half of the total thickness of the unit, is densely welded devitrified tuff with a discontinuous band of nondevitrified black glass near the base; and the upper part of the sheet is soft and nonwelded with obvious signs of vapor-phase alteration. In most places, but especially where the unit is thick, columnar joints are pervasive normal to compaction layering. Where the upper parts of the ashflow have been stripped or where the basal unwelded part is unusually thick and not protected from erosion by the resistant welded zone, erosion has sculptured scenic badland topography (fig. 28) characterized by numerous odd-shaped hoodoos (hence the name of the canyon for which the unit is named).

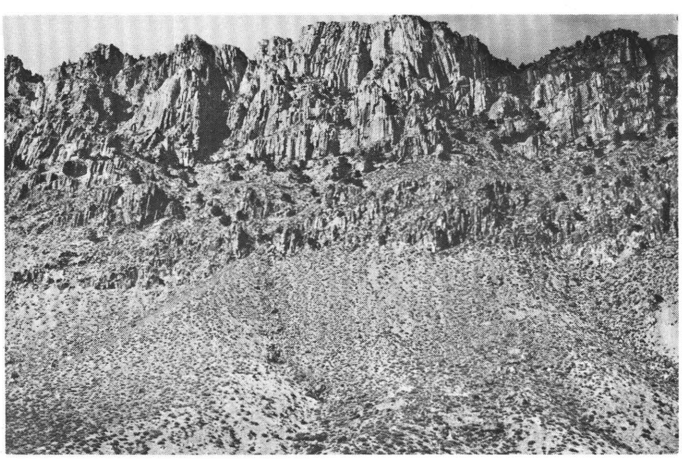


FIGURE 27.—Tuff of Hoodoo Canyon in Wildcat Canyon. The formation is about 400 feet (125 m) thick here.

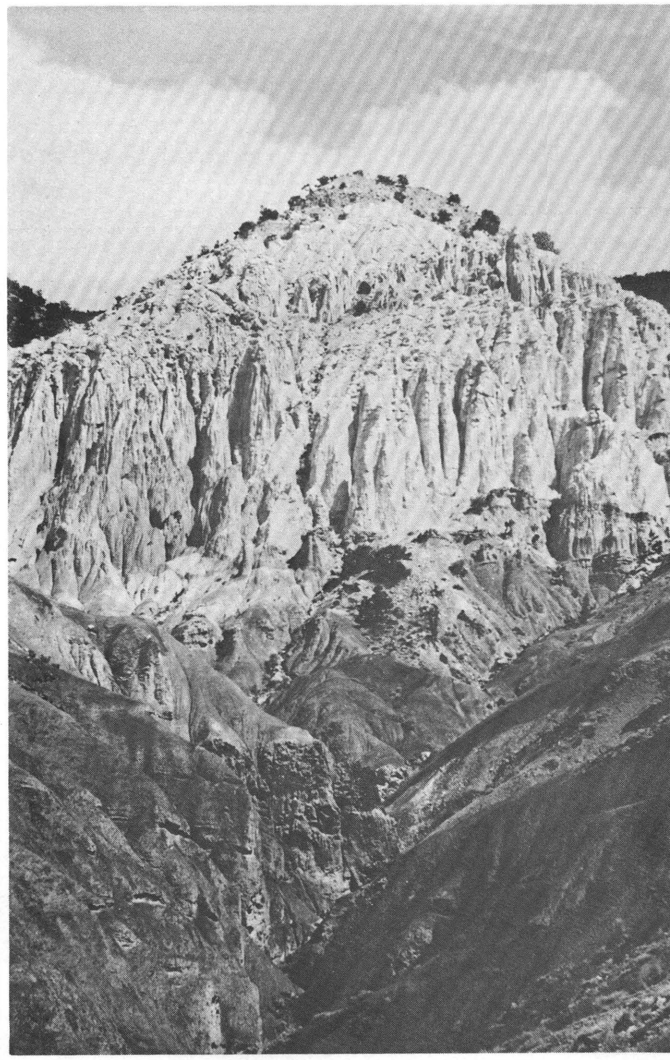


FIGURE 28.—Basal part of the tuff of Hoodoo Canyon in Hoodoo Canyon.

LITHOLOGY

The tuff of Hoodoo Canyon is gray with darker shades corresponding to increase in welding and density of the rock. Typically the nonwelded part of the tuff is nearly white, and the densely welded part almost black. Surfaces parallel to the eutaxitic fabric are nearly covered with oriented plates of biotite that make the rock look dark and accentuate the amount of biotite in it. Surfaces normal to this fabric are of lighter hues, and biotite appears much less significant. In places large elongate cavities as much as 3 inches (7.5 cm) long and parallel to the compaction surfaces emphasize the welded and compressed fabric of the rock. Locally abundant pumice also reflects the degree of compaction and welding and ranges from angular unoriented fragments to streamlined fiammi (fig. 29).

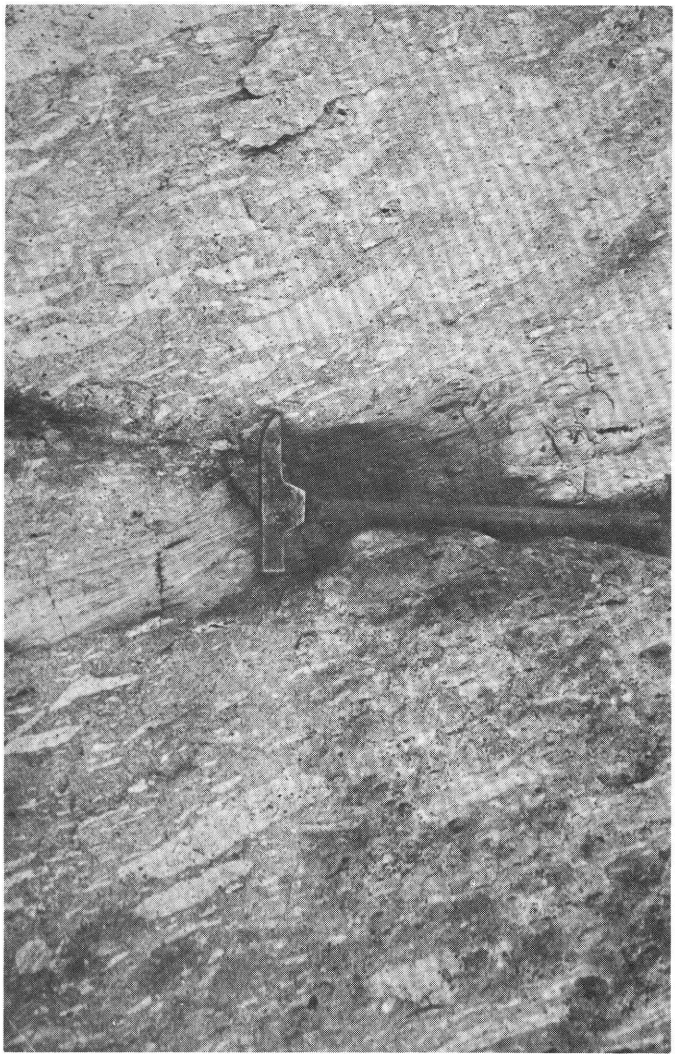


FIGURE 29.—Oriented and slightly flattened pumice in the basal part of the tuff of Hoodoo Canyon.

In the most densely welded parts of the tuff, the pumice is so flattened that it is not obvious except in thin section, and thus the rock appears homogeneous.

Phenocrysts compose about a quarter of the rock and consist of about 65 percent plagioclase (andesine), 15 percent biotite, 5 percent pyroxene, and 11 percent sanidine. A few quartz and hornblende phenocrysts were seen in some thin sections. The prevalence of plagioclase, the large percentage of mafic minerals, and the sparsity of quartz contrast markedly with the other tuffs in the area. The groundmass contains devitrified shards bent around phenocrysts and compressed and welded to various degrees. A few xenolithic fragments of dacitic volcanic types or chert are present locally. Vapor-phase crystallization is ubiquitous in the soft unwelded top of the flow.

CHEMISTRY

A whole-rock chemical analysis (table 5) used as the basis for the normative orthoclase, albite, and anorthite diagram (figure 30) indicates that the rock is significantly less silicic than the other ash-flow tuffs in the northern Toquima Range (table 5) and that it is a quarz latite. The normative orthoclase, albite, and anorthite content of three samples of biotite-rich tuff from localities a short distance north and northeast of the Toquima Range are plotted on figure 30 to show the similar petrochemistry of other biotite-rich welded tuffs about the same age as the tuff of Hoodoo Canyon.

AGE AND CORRELATION

Biotite from two samples of the tuff of Hoodoo Canyon yielded potassium-argon ages of 30.1±1.1 and 30.6±1.2 m.y. old (table 4). These radiometric dates suggest that the unit is of Oligocene age (Harland and others, 1964). The age fits well with other dated units in the stratigraphic sequence. The relatively long time gap (about 7 m.y.) between the tuff of Hoodoo Canyon and the overlying Bates Mountain Tuff is recorded in the field as the erosional unconformity between these units. During this time the tuff of Hoodoo Canyon was completely eroded from parts of the region and remains elsewhere as discontinuous pockets preserved beneath the Bates Mountain Tuff.

Welded tuff similar in composition to the tuff of Hoodoo Canyon and of about the same age is widespread in central and eastern Nevada and southwest-

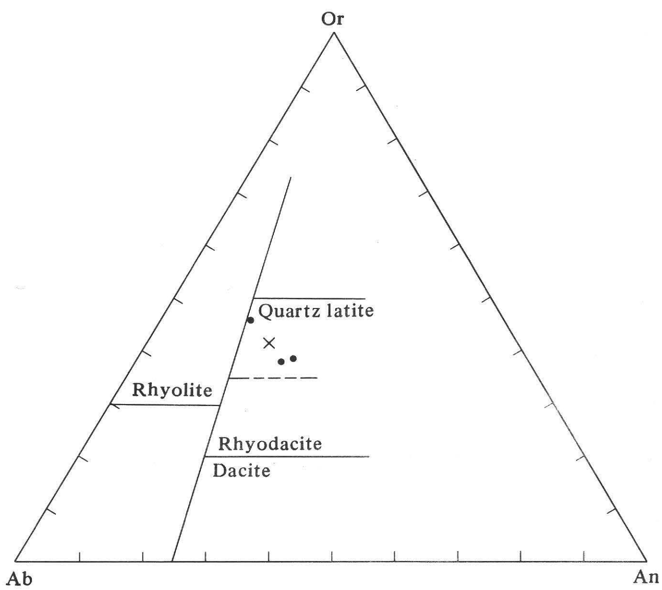


FIGURE 30.—Normative molecular albite, anorthite, and orthoclase. Normative classification after O'Conner (1965, fig. 3). X is tuff of Hoodoo Canyon; the dots are biotite-rich tuff from a few miles north and east of the Toquima Range.

ern Utah. The most widespread of these units in southwestern Utah and eastern Nevada is the Needles Range Formation (Mackin, 1960; Cook, 1965). This formation has been recognized on the basis of natural remanent magnetic direction (Grommé and others, 1972) as far west as Currant, Nev. (long. 115°30′) about 70 miles (110 km) southeast of the northern part of the Toquima Range. Other biotite-rich tuffs of the same age as the Needles Range Formation and the tuff of Hoodoo Canyon are in Eureka, Lander, Nye, and Churchill Counties; some may be parts of the tuff of Hoodoo Canyon, but others undoubtedly are not. Biotite-rich tuff in the Simpson Park Range and the Grimes Hills directly north of the Toquima Range has been informally called tuff of Bottle Summit (McKee, 1968a, b; Stewart and McKee, 1968) and most likely is the tuff of Hoodoo Canyon.

The wide distribution across about half of Nevada and part of Utah of a number of different biotite-rich ash-flow sheets can best be explained as having been caused by eruptions from different centers. Why they are so similar petrochemically and of about the same age leads to interesting speculation on their ultimate source.

NATURAL REMANENT MAGNETIZATION

Measurement of the natural remanent magnetic direction with a fluxgate magnetometer at many places indicates that the tuff of Hoodoo Canyon has reverse magnetization. In fact a reverse remanent magnetic direction was used as one criterion in identifying the formation. In the detailed study of remanent magnetic direction of ash flows in the Great Basin by Grommé, McKee, and Blake (1972), a variety of magnetic directions were found in biotite-rich tuffs of Oligocene age, precluding the possibility that all the tuffs belong to the same unit. Two of the samples measured by Grommé, McKee, and Blake (1972), samples N 1 and N 4, table 2 and fig. 6) were collected from the tuff of Hoodoo Canyon in the northern part of the Toquima Range.

BATES MOUNTAIN TUFF

The Bates Mountain Tuff forms flat uplands in the central part of the area from Corral Canyon south to Stoneberger Canyon. This extensive surface of low relief contrasts with the rugged canyon and ridge topography typical of the eroded pre-Tertiary rocks elsewhere in the region. Other exposures of Bates Mountain Tuff are on the west side of the range from Clipper Gap Canyon north to the Linka mine. Here, as in the central part of the area, Bates Mountain cooling units form a series of slopes and low cliffs (fig. 31). The thickest section of the formation in the area is directly north of Clipper Gap Canyon. It has been described by Sar-

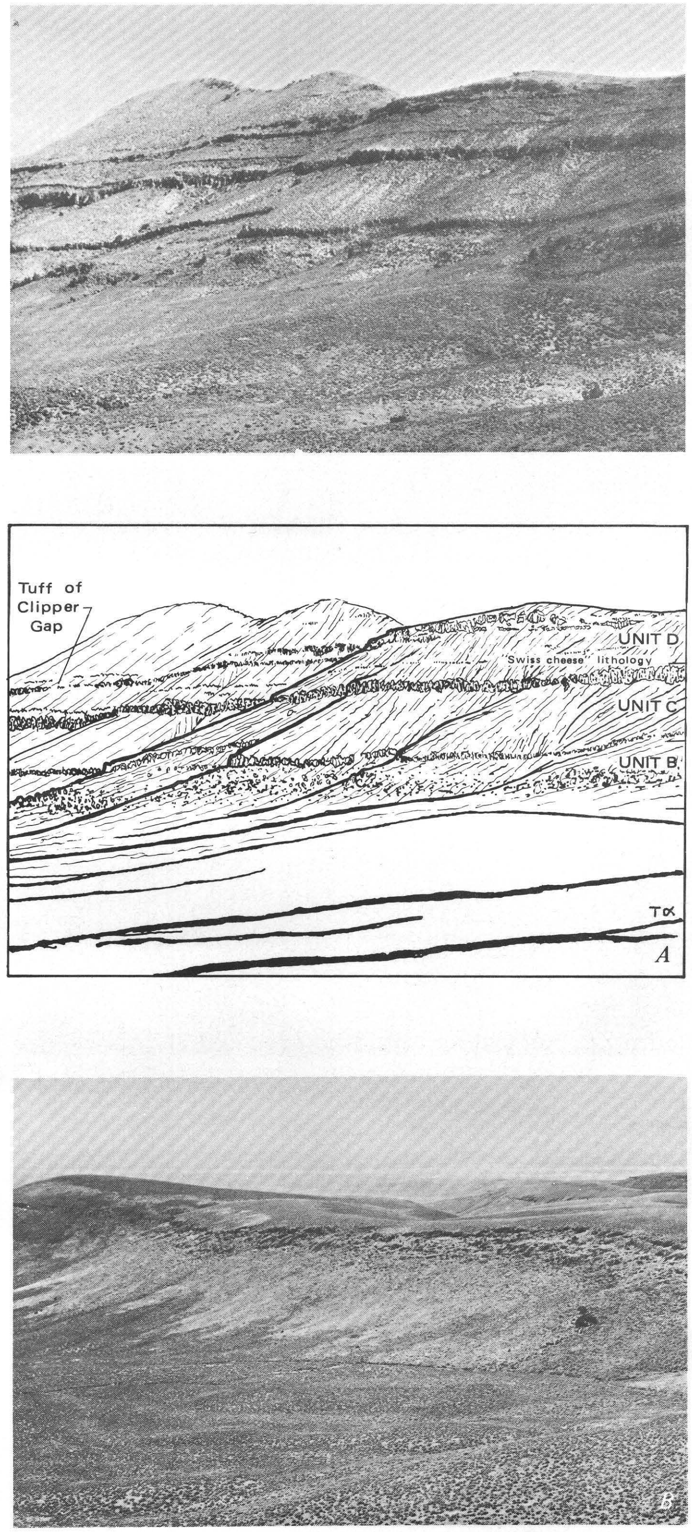


FIGURE 31.—Bates Mountain Tuff. *A*, At Clipper Gap Canyon. The lowest two cliffs are densely welded zones in unit C (a compound cooling unit), the upper cliff is the devitrified zone of cooling unit D. Rounded peaks are tuff of Clipper Gap. *B*, Bates Mountain Tuff in the low hills east of the Linka mine. These exposures are typical of the formation over large areas in central Nevada.

gent and McKee (1969), and Grommé, McKee, and Blake (1972). Here, three cooling units with a total thickness of slightly more than 600 feet (200 m) compose the formation. In other places northwest of here (about 15 miles or 25 km north of Austin) four cooling units are recognized, but in most parts of the Toquima Range south of Clipper Gap Canyon and to the east in the Monitor, Fish Creek, and Antelope Ranges, only one of these units is found.

This report uses the subdivisions of the Bates Mountain Tuff defined by Grommé, McKee, and Blake (1972, table 4), which recognizes four cooling units A through D. It differs from the earlier five-unit (cooling units 1 through 5) system of Sargent and McKee (1969). A comparison of the Grommé, McKee, and Blake (1972) and the Sargent and McKee (1969) subdivisions is shown on table 6 and in figure 32.

UNIT B

The oldest cooling unit of the Bates Mountain Tuff in the Toquima Range is unit B of Grommé, McKee, and Blake (1972, p. 1631 and table 4). At Clipper Gap Canyon the unit is about 160 feet (50 m) thick and consists of a nonwelded base, a middle densely welded zone, and an upper moderately to densely welded zone. The upper part erodes to a steep slope or discontinuous cliff. Vapor-phase alteration forms a lacework near the top of the unit; pumice cavities and other holes are lined with vapor-phase minerals.

Welded devitrified tuff of unit B is white to pale pink, contains only about 10 percent phenocrysts, and is punky to moderately hard. Thin sections reveal eutaxitic texture as would be expected of this welded shard-rich rock. Phenocrysts consist mostly of 50–60 percent sanidine and 30–40 percent plagioclase. Quartz, biotite, hornblende, unidentified altered mafic minerals, and opaque grains form only a few percent of the phenocrysts. A chemical analysis of a sample from the most densely welded part of this unit (table 5) and a plot of normative minerals (fig. 33) incidate that the rock is a rhyolite.

Measurements in the field with a fluxgate magnetometer corroborate the detailed measurements reported by Grommé, McKee, and Blake (1972) indicating that unit B has normal remanent magnetism. The study by Grommé and co-workers defines a direction of natural remanent magnetization at 39.1° inclination (downward) and 17.8° declination east of north for 14 cores drilled near Clipper Gap Canyon (Grommé and others, 1972, table 2, site B10). This direction is the same as that determined at sites from three other widely separated localities in central Nevada.

TABLE 6.—*Cooling units in the Bates Mountain Tuff as described by Grommé, McKee, and Blake (1972) and Sargent and McKee (1969)*

[This stratigraphic column was compiled from measured parts of the section north of Clipper Gap Canyon (shown in fig. 32; also see fig. 31*A*) and is described here because it is the thickest and most complete section of the formation in the region]

Unit (Grommé and others, 1972)		Unit Sargent and McKee, 1969)
Tuff of Clipper Gap		5
D		4
C	Compound cooling unit	3
		2
B		1
A		

UNIT C

Unit C is the thickest cooling unit and forms prominent cliffs about midway in the ash-flow sequence. The unit consists of two parts. The lower part is moderately soft, forms slopes, and grades upward into progressively more welded tuff that merges into densely welded tuff that forms a cliff about 30 feet (10 m) high; locally at the base of the cliff is a lenticular zone of black glass as much as 1 m thick. Above the cliff is a second slope-forming part that begins with nonwelded to weakly welded tuff. A partial cooling break, somewhere in this slope but not obvious in the field, separates the two parts of unit C. Above the cooling break the tuff again becomes increasingly more welded upward and merges into densely welded tuff that forms a second cliff about 50 feet (15 m) high. Above this cliff the cooling unit consists of vapor-phase mineralization that is nonwelded to partly tuff showing. The two densely welded cliff-forming intervals in unit C are so similar in general appearance that they are indistinguishable in the field. Detailed studies of petrography, chemistry, and remanent mangetism confirm this similarity and provide the basis for considering them as parts of one compound cooling unit.

Densely welded tuff from the lower cliff of unit C is pink to dull orange and knubbly weathering and contains about 7 percent phenocrysts. Eutaxitic texture is clearly seen in thin section, and large flattened pumice fragments scattered throughout the rock are obvious when viewed on the plane of compaction; at right angles to this plane they show as thin whispy fiammi or elliptical cavities. Pumice in various states of collapse is seen in the underlying soft partly to nonwelded tuff. Phenocrysts are sanidine and somewhat less quartz and plagioclase. A few crystals of pyroxene were noted in thin section. A chemical analysis of this rock (table 5) and its normative feldspar content (fig. 33) indicate that the rock is rhyolite. Densely welded tuff from the upper cliff in unit C is similar to the lower cliff described above, although modal counts of several thin sections indicate that it may contain relatively more plagioclase and less sanidine than the lower densely welded zone.

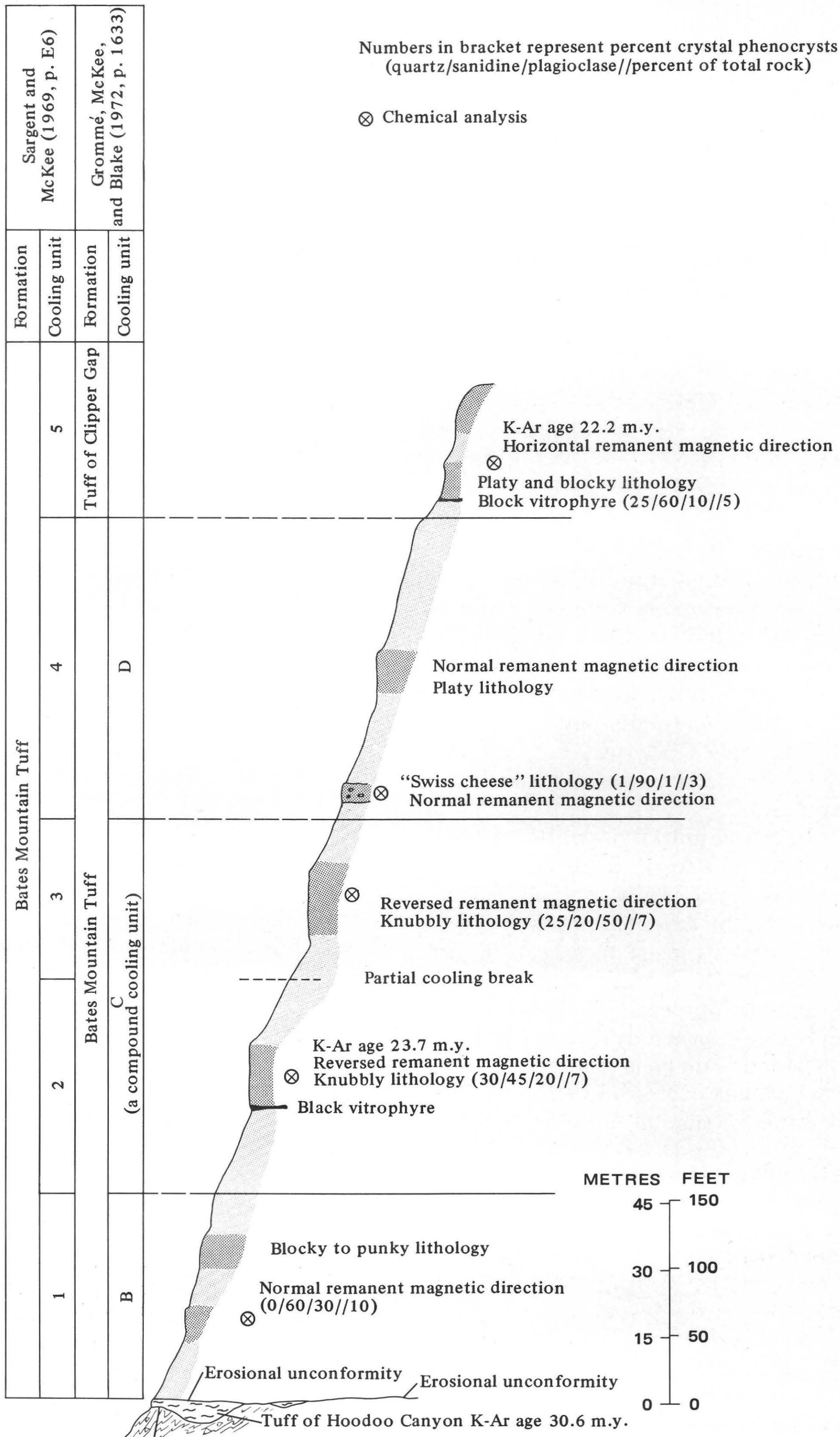


FIGURE 32.—Composite of measured sections of the Bates Mountain Tuff at Clipper Gap Canyon. Subdivisions according to Grommé, McKee, and Blake's (1972) revision of Sargent and McKee's (1969) classification included.

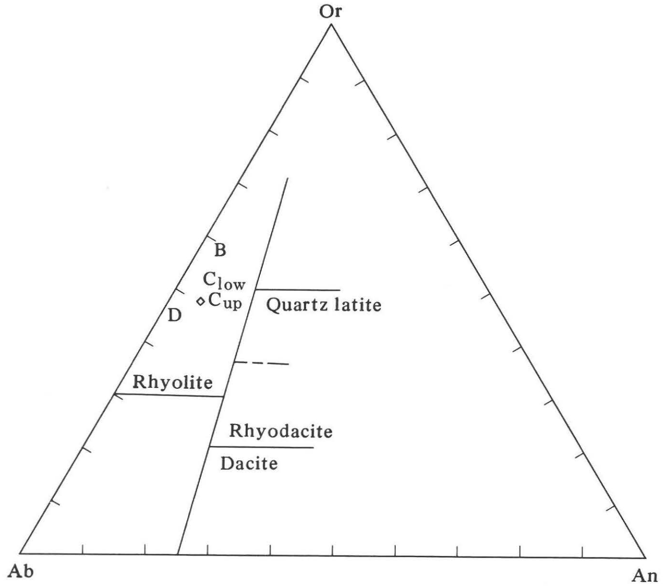


FIGURE 33.—Normative albite, anorthite, and orthoclase plot of the Bates Mountain Tuff and the tuff of Clipper Gap. Normative classification after O'Conner (1965, fig. 3). Points B, C_{low}, C_{up}, and D are cooling units of the Bates Mountain Tuff; ◇ is the tuff of Clipper Gap.

The direction of natural remanent magnetization of both densely welded zones in unit C is almost identical and is reversed (southern hemisphere). Ten oriented drill samples from each unit, collected near Clipper Gap Canyon, were demagnetized, and their natural remanent magnetic direction was determined by Grommé, McKee, and Blake (1972). The direction they reported is −41.4° inclination, 202.2° declination, and −50.4° inclination, 202.9° declination for the lower and upper cliffs, respectively (table 2, B11, B12). This direction is the same as that determined at a locality about 50 miles (80 km) to the northwest on a unit of Bates Mountain Tuff, which is considered to be the same ash-flow cooling unit.

Sanidine separated from a sample of tuff from the lower densely welded zone yielded a potassium-argon age of 23.7 m.y. (table 4). This age is the same as the average age of seven samples of Bates Mountain Tuff (including this sample from Clipper Gap Canyon) reported by Grommé, McKee, and Blake (1972, table 7). The 22.2 m.y. age of the tuff of Clipper Gap serves a minimum age of the Bates Mountain Tuff. Unit C, which is about midway in the section of cooling units of Bates Mountain Tuff, probably gives a good average age for the formation as a whole. Units below and above are probably not more than about 1 m.y. older or younger, respectively.

UNIT D

Unit D is the youngest cooling unit of the Bates Mountain Tuff. At Clipper Gap Canyon it is a little more than 200 feet (60 m) thick and contains two densely welded zones that erode to form cliffs. The lower of these has the swiss-cheese lithology illustrated in figure 34 that distinguishes unit D from all other welded tuffs in the region. The upper, densely welded zone is similar-appearing pink crystal-poor tuff but without the gas cavities of the swiss-cheese zone. The two densely welded zones merge along strike in about 1 mile, suggesting that they are local compaction and welding phenomena. The lower and upper parts of the cooling unit are nonwelded to weakly welded, the upper part showing alteration and vapor-phase mineralization.

Densely welded tuff from unit D is pink to purplish red, very hard, and almost devoid of crystal phenocrysts. Of the phenocrysts, sanidine is the most abundant (about 3 percent of the rock), less than 1 percent is quartz and plagioclase, and about 1 percent is unidentified opaque oxides or altered mafic minerals. A few flakes of biotite can be found in most outcrops. In thin section the tuff shows oriented and compressed shards around gas cavities that are slightly elliptical with their long axes alined in the eutaxitic fabric formed by the welded shards. Less welded parts of the tuff are soft to punky, and the gas cavities are poorly developed or absent. A chemical analysis (table 5) and normative feldspar content plotted on a ternary diagram (fig. 33, point D) indicate that the rock is a rhyolite.

Eleven oriented drill samples collected from unit D at Clipper Gap Canyon were demagnetized, and their direction of natural remanent magnetization was determined by Grommé, McKee, and Blake (1972, table 2, site B–13). The direction is normal, with 57.1° inclination and 330.2° declination, and is about the same as that measured on samples from seven other widely

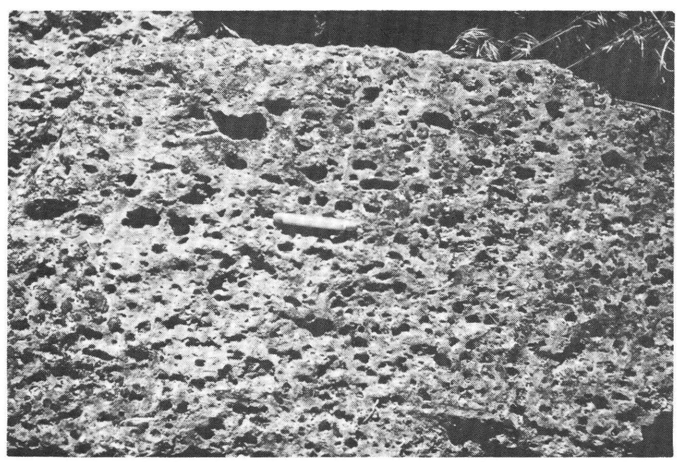


FIGURE 34.—Densely welded tuff of unit D, Bates Mountain Tuff; note swiss-cheese lithology.

spaced localities in central Nevada (Grommé and others, 1972, fig. 7).

The four cooling units of Bates Mountain Tuff, three of which are at Clipper Gap Canyon, alternate from a normal to reversed remanent magnetic direction starting with the basal unit A, which is normal. Because of this alternating sequence of reversals, determination of the direction of any three cooling units in a series offers a unique means of correlation and of determining which units are represented. The youngest ash-flow sheet in the region, the uppermost unit in the Clipper Gap Canyon section (see "Tuff of Clipper Gap" p. 145), can be distinguished from the Bates Mountain units by its unique horizontal natural remanent magnetic direction.

TUFF OF CLIPPER GAP

The youngest Tertiary unit in the northern part of the Toquima Range is an ash-flow sheet informally called by Grommé, McKee, and Blake (1972, p. 1633) the "tuff of Clipper Gap." This ashflow is recognized in an area of over 6,000 km^2 in central Nevada and has an estimated volume of approximately 190 km^3 (Grommé and others, 1972, table 1). The northwesternmost outcrops of the formation are in the Toquima Range. In this range and areas to the east, it lies on top of the Bates Mountain Tuff, which it greatly resembles, and because of this similarity, it was included in the Bates Mountain Tuff by Sargent and McKee in 1969. When the total distribution of the Bates Mountain Tuff was compiled (Grommé and others, 1972), it was apparent that the cooling unit they termed the tuff of Clipper Gap cropped out over a different region than the underlying units and probably erupted from a different center. The age and direction of remanent magnetism of the tuff of Clipper Gap provide additional reasons for separating it from the Bates Mountain Tuff.

In the northern part of the Toquima Range, the tuff of Clipper Gap forms the nearly flat surface north of Wildcat Peak and the top of the ridge north of Clipper Gap Canyon (figure 31*A*). At the latter locality it is about 100 feet (32 m) thick and is composed of nonwelded, slightly welded, and densely welded tuff. At other places it is less than about 70 feet (23 m) thick and is mostly densely welded.

The tuff of Clipper Gap, as exposed at Clipper Gap Canyon, consists of a basal nonwelded zone that progressively becomes more intensely welded upward into a zone of dense, hard, welded tuff about 40 feet (13 m) thick. Most of the densely welded tuff in this zone is devitrified, but a black glassy lens occurs discontinuously at its base. Above this zone the tuff is moderately to slightly welded. It is pink to light red and contains about 5 percent phenocrysts. Sanidine is the most common phenocryst but also present are quartz, plagioclase, and biotite. The tuff is most like unit D of the underlying Bates Mountain Tuff, but it lacks the gas cavities and swiss-cheese texture characteristic of that unit. Typically the tuff splits into platy slabs that have strongly developed eutaxitic fabric. In places the rock has a fabric resembling flow banding common in rhyolite lavas, indicating that some flowage occurred in the hottest, most compressed parts of this ash flow.

On the basis of modal analysis from thin sections of the rock (fig. 32) and a chemical analysis (table 5), the tuff of Clipper Gap is classified as rhyolite.

Sanidine from a sample of the tuff of Clipper Gap collected at Clipper Gap Canyon yields a potassium-argon age of 22.2 m.y. (table 4). A sample from near Eureka, Nev. has a potassium-argon age of 22.1±0.7 m.y. (Grommé and others, 1972, table 7). On the basis of these dates, the unit is considered to be of early Miocene age. The actual age difference between the tuff of Clipper Gap and the underlying Bates Mountain Tuff dated at about 23 m.y. cannot be determined from the potassium-argon data because of analytical uncertainty (±0.7 m.y.), but the dates suggest that the tuff of Clipper Gap may be about 1 m.y. younger than unit D of the Bates Mountain Tuff. It is significantly younger than other Bates Mountain units.

Directions of natural remanent magnetization at seven widely spaced sites in the tuff of Clipper Gap are distinctive. They are nearly horizontal, with inclinations ranging from 8.3° to −8.1° and a declination of about 158° (average of 6) east of north (Grommé and others, 1972). No other volcanic rock in the region has a similar remanent magnetic direction, and as such, it is a useful means of identifying the tuff of Clipper Gap.

QUATERNARY SYSTEM

Deposits of Quaternary age in the northern part of the Toquima Range include older alluvial fans on the margins of the range, younger alluvial fans, dune deposits, and playa deposits in the valleys east and west of the range, tufa at hot springs, gravels in modern streams, and colluvium over bedrock. The transition between many of these deposits is subtle, and the contact is most easily discerned from aerial photographs. Since this study was primarily concerned with bedrock geology, Quaternary deposits were mapped only where bedrock beneath them was not completely obscured.

Fan deposits.—Two generations of unconsolidated to weakly consolidated fan deposits are recognized and mapped. The older deposits are distinguished on the basis that they are being eroded today, whereas the younger fan deposits are actively forming at the range front. Some of the deeply dissected older fan deposits may be as old as Tertiary if their incompatibility with today's topography and amount of dissection is any measure of age; others are only slightly out of adjustment with the present cycle of erosion and deposition and merge with the modern fan deposits. As no fossil or radiometric evidence is available, all the fan deposits are classified as Quaternary.

Colluvium and talus.—A discontinuous veneer of loose rock and soil covers bedrock in the range, especially in the forested area near Petes Summit and on the flat top of the range near the headwaters of Stoneberger Canyon. This cover is seldom thick or persistent enough to obscure the bedrock beneath it, and it consists of debris from the underlying bedrock. In a few places in the rugged terrain around Wildcat Peak, Mill Canyon, and June Canyon, talus forms great piles in the valley bottoms and sheets streaming down the slopes. Many of these piles are shown on the map (pl. 1), but in most places bedrock contacts can be projected beneath them with little difficulty.

Alluvium.—About half the area included on the geologic map (pl. 1) is alluvium of various types. This material includes sand, gravel, and boulders of alluvial fans that spread outward from the range into Big Smoky Valley on the west and Monitor Valley on the east. These fans merge valleyward with gravel, sand, and silt deposited by the stream systems of both valleys. A short distance west of the western edge of the study area in Big Smoky Valley, alluvial deposits include silt and salts of a large playa. (Salt is abundant enough in these deposits to have made it economically valuable as a flux in silver smelters at Austin during the latter part of the last century.) Few of the alluvial deposits are cemented except locally by a small amount of caliche or by clay and mud. The surface is generally soft, but the material is remarkably coherent at a few cm depth.

Dune deposits.—In Big Smoky Valley the remains of ancient shore levels are clearly seen on aerial photographs. Some of these strand levels are now the sites of small dunes formed from sand accumulated by the prevailing north winds that blow the length of Big Smoky Valley and across the playa. The dunes are largely anchored by vegetation and seem to have reached an equilibrium in growth or reduction.

Tufa.—Around Spencer Hot Springs and at two other places, one abut 0.5 mile (0.8 km) northwest and the other about 1 mile (1.6 km) east of the present springs, small aprons of tufa mark sites of present or former hot springs. These hot-spring deposits are predominately calcareous, although analysis by gamma spectrometry for selected radioactive elements reveals 1.51 percent potassium, 11.54 parts per million uranium (apparent), and 10.92 parts per million thorium (Wollenberg, 1974, table II). A chemical analysis of the spring water from Spencer Hot Springs is given below.

[Concentrations are given in milligrams per litre; L.M. Willey, analyst. Calcite is the major mineral precipitated from this spring]

SiO_2	77	HCO_3	671.54
Ca	43	CO_3	.23
Mg	9.4	SO_4	51
Sr	1.8	Cl	22
Na	200	F	4.8
K	36	I	.02
As	.031	PO_4	.01
		B	2.8

STRUCTURAL GEOLOGY

The northern part of the Toquima Range, like other ranges in the central Great Basin, contains both low- and high-angle faults (pl. 1). It is a mosaic of fault blocks such that stratigraphic units can be traced along strike without offset in most places for only a few hundred metres or less. Folding is visible on a small scale (folds with an amplitude of a few metres, see fig. 5) but is not typical of the tectonic style of the area. The faults are of two very different types and ages: (1) those that formed during the Antler orogeny of Devonian and Mississippian age (Roberts Mountains thrust and related thrusts) and (2) Basin and Range faults of late Cenozoic age that cut all rocks in the area and are responsible for the present topography. Structural features related to collapse and possible resurgence of the Northumberland caldera are found in the southwestern part of the area.

FIELD CRITERIA FOR FAULTS

The most obvious evidence of faulting is truncation of bedding. Sheared and brecciated rocks mark some of the faults but do not indicate the amount of offset. Indeed many of the larger faults have no gouge or breccia associated with them. Alinement of dikes, zones of alteration, or springs in some places indicate a fault but tell little about its size.

More subtle and usually more significant geologic evidence for faulting lies in fossil evidence and stratigraphic interpretations. Older rocks overlying younger, stratigraphic sequences lacking regionally persistent units, and the juxtaposition of contrasting facies of the same age are features of this category. Recognition of the existence, magnitude, and regional implications of the Roberts Mountains thrust, for example, stems from such data. The fault itself generally is not marked by a particularly well developed zone of shearing or alteration. Whereas the Roberts Mountains thrust has telescoped rocks of widely different lithologies and is readily recognized, other regional thrusts such as the Mill Canyon and June Canyon thrusts have brought similar sequences of carbonate rocks together. Only when the details of the units are compared is it obvious that thrusting with transport of 10 miles (16 km) or more is required to explain the facies patterns.

ROBERTS MOUNTAINS THRUST

The most widespread and regionally important tectonic feature in the area is the Roberts Mountains thrust. First documented by Merriam and Anderson (1942) in the area around Roberts Mountains about 50 miles (80 km) northeast of the Toquima Range and subsequently mapped at many places in central Nevada (see Roberts and others, 1958; Gilluly and Gates, 1965, for regional discussion), it is recognized in the northern part of the Toquima Range by the same evidence as elsewhere—the juxtaposition of sharply contrasting facies of rocks of about the same age. The facies involved are siliceous and carbonate types; the siliceous type is allochthonous.

The siliceous (allochthonous) rocks in the Toquima Range are assigned to the Vinini Formation and comprise shale, chert, quartzite, and a little dark limestone. Ordovician graptolites are common in places in these rocks. The carbonate rocks are mostly limestone ranging in age from Ordovician into the Devonian and contain a shelly fauna with some graptolites. In general the carbonate units of the autochthon are thicker bedded and of lighter hues than the siliceous rocks. The dark siliceous rocks of the allochthon are considerably more deformed than those of the autochthon (see fig. 5). This deformation is so strong and pervasive in the southern part of the area that no attempt was made to map the structural details. Beds can be traced for only a few metres before they are offset or sheared out, and generally they cannot be matched with beds in nearby fault blocks. Strike and dip of beds record the jumbled nature of these rocks because in areas of only a few square metres nearly any attitude can be measured.

Rocks along the Roberts Mountains thrust are not notably more deformed than those in the chaotic overlying thrust plates except in one area north of Petes Canyon where a 20-foot-thick (6 cm) zone of crushed and recemented chert and quartzite marks the sole of the thrust (fig. 35). This exposure of fault gouge extends for about one-half mile (0.8 km) before it is obscured by alluvium.

DISTRIBUTION

The Roberts Mountains thrust outlines two windows of carbonate rocks in the northern and southern part of the map area (fig. 3). These are named the Petes Canyon and Ikes Canyon windows, and the thrust around them is an easily mapped feature because of the marked contrast in rock type. In addition, two small

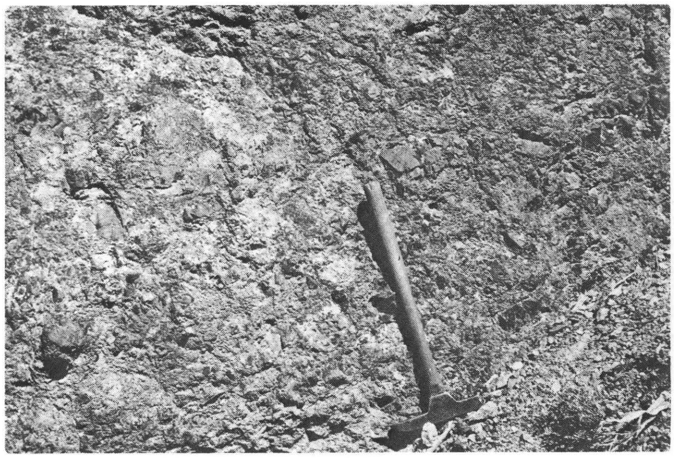


FIGURE 35.—Fault breccia that forms the sole of the Roberts Mountains thrust north of Petes Canyon. This fault gouge can be traced along the fault for about one-half mile (0.8 km). It consists of fragments of crushed chert and quartzite.

klippen of upper plate rock, remanents of the overthrust sheet, are exposed in the Ikes Canyon window. A total of about 10 miles (16 km) of the fault is shown on plate 1.

ATTITUDE

The attitude of the fault varies from horizontal to vertical but is mostly rather flat. The variation is probably due to a combination of the original nonplanar form of the fault and various amounts of later tilting. At many places it is cut and offset by later faults (as in Mill Canyon), but it retains continuity elsewhere for remarkably long distances considering its antiquity (see "Age of the Roberts Mountains Thrust in the Toquima Range") and the complexity of the regional structure. The fact that this fault can still be traced as an unbroken subhorizontal feature in places for several miles argues strongly that little tectonism affected the region since its formation.

IMBRICATE THRUSTS IN THE ALLOCHTHON

At least 10 imbricate thrust faults in the upper plate of the Roberts Mountains thrust offset the Vinini Formation in the area between Petes Canyon and Corral Canyon in the central part of the area. These faults could be traced because the stratigraphy in this part of the thrust plate is distinctive enough to establish mappable units that can be traced throughout the 50 km^2 of exposure, hence the naming by Kay and Crawford (1964) of the Clipper Canyon Group consisting of their Charcoal Canyon, Petes Summit, Sams Springs, and Joes Canyon Formations for these strata; however, this classification is not adopted here because of the limited extent of these formations. The structural pattern within this series of recognizable units is that of imbricate northwest-dipping thrust slices offset by some steeply dipping faults normal to the surface trace of the thrusts (see pl. 1 and cross section C–C′). There are at least 10 imbricate plates, all of which can be traced almost continuously for about 2 miles (3.2 km). The fact that these imbrications are so little disturbed suggests, as does the continuity and attitude of the Roberts Mountains thrust at the base of the allochthon, that there was little deformation in this area after emplacement on the thrust. The imbricate structure probably developed in a relatively competent part of the allochthon during its emplacement. In other parts where thin-bedded chert is the major lithologic units, shearing, disharmonic folding, and variously oriented faults developed, creating chaotic structure.

REGIONAL THRUSTS WITHIN THE AUTOCHTHON

In the lower plate of the Roberts Mountains thrust in the Ikes Canyon window, three sequences of carbonate rocks are separated by two thrusts (see fig. 3). The carbonate rocks in these sequences are interpreted to have been deposited in areas about 35 miles (56 km) apart and to have been juxtaposed by movement on these thrust faults (see pl. 2). These different sedimentary sequences and the thrusts between them were first recognized by Kay (1960) and more fully described and shown on a geologic map by Kay and Crawford (1964). The rocks in the different thrust plates were named the August Canyon, Mill Canyon, and June Canyon sequences, and the thrust faults between sequences the Mill Canyon and June Canyon thrusts (Kay, 1960; Kay and Crawford, 1964). This nomenclature is retained here.

The bottom sequence of strata, the August Canyon sequence, may or may not rest on a thrust, since no basal contact is exposed. As discussed in the section "August Canyon Sequence," these rocks are considered to be autochthonous.

MILL CANYON THRUST

Truncating the youngest formation in the August Canyon sequence, the Masket Shale of Kay and Crawford (1964), is the Mill Canyon thrust on which limestone of the Antelope Valley and younger formations have ridden over rocks of the August Canyon sequence. The thrust can be traced for about 5 km with only minor offset and is cut by a younger steeply dipping fault at its south end in Mill Canyon. The Mill Canyon thrust dips between 15° and 20° to the northwest and is generally subparallel to the strike of the underlying and overlying strata. The upper plate of this thrust contains the Mill Canyon stratigraphic sequence that is noticeably different from the August Canyon stratigraphic sequence in the lower plate. Stratigraphic comparisons and regional facies patterns of these two rock sequences suggest that transport on the Mill Canyon thrust was on the order of 20 miles (30 km). Thrusting was most likely from the northwest to southeast.

JUNE CANYON THRUST

Rocks of the June Canyon stratigraphic sequence overlie the Mill Canyon sequence on the June Canyon thrust. This fault is generally parallel to the Mill Canyon thrust and has about the same dip (15°–20° NW). It can be traced for about 5 miles (8 km) along strike and includes one klippe about one-quarter mile (0.4 km) in diameter. Along most of its length the sole is in the Antelope Valley Limestone, which rides over the Roberts Mountains Formation, but in a few places the fault cuts across other formations. Reconstruction of facies patterns between rocks in the upper plate (June Canyon stratigraphic sequence) which differ from those in the lower plate (Mill Canyon sequence) suggests a tectonic transport of about 15 miles (24 km) between plates. A total of about 55 km of telescoping from the west to the east is thus postulated along the two thrusts (Mill Canyon and June Canyon) separating the three plates in the Ikes Canyon window (see pl. 2).

Evidence bearing on the age of the thrusts within the autochthon is available for the June Canyon thrust. This fault is cut by the Roberts Mountains thrust and overlapped by allochthonous siliceous rocks of the Vinini Formation near the head of Mill Canyon. At another place about one-half mile (0.8 km) east of the head of the north tributary of Mill Canyon, a small (about 100 m in diameter) klippe of Vinini lies across the June Canyon fault. These two localities demonstrate the relative older age of this thrust with regard to the regionally more extensive Roberts Mountains thrust. Two other klippen of Vinini that lie on rocks of the June Canyon stratigraphic sequence on the ridge south of the upper reaches of Ikes Canyon and the klippe east of north Mill Canyon that is mostly on the Mill Canyon stratigraphic sequence also demonstrate that the allochthonous rocks above the Roberts Mountains thrust rode across the region after emplacement of the June and Mill Canyon thrusts.

AGE OF THE ROBERTS MOUNTAINS THRUST IN THE TOQUIMA RANGE

The Vinini Formation of the allochthon, which moved on the Roberts Mountains fault, is overlain unconformably by conglomeratic limestone of the Pennsylvanian Wildcat Peak Formation on the ridge south of Mill Canyon and at Wildcat Peak. Beds of the underlying Vinini are greatly deformed, whereas the overlying Wildcat Peak Formation is mostly undeformed. Erosion of the underlying Vinini and incorporation of Vinini chert fragments in the basal part of the Wildcat Peak Formation also mark the unconformity.

At Mill Canyon, therefore, faulting responsible for the transport of siliceous rocks (Vinini) onto carbonate rocks exposed in the Ikes Canyon window must have been pre-Wildcat Peak or pre-Early Pennsylvanian (see p. 79). The youngest rock involved in the thrust movements is unnamed dark limestone that crops out one-quarter mile (0.4 km) south of the mouth of Ikes Canyon and is Middle Devonian in age. Thus in the Mill Canyon–Ikes Canyon area, the time of emplacement of the siliceous western and transitional facies rocks on the Roberts Mountains thrust is between Middle Devonian and Early Pennsylvanian.

The age of this regional dislocation in the Mill Canyon area is in general agreement with that postulated by Roberts, Hotz, Gilluly, and Ferguson (1958, p. 2850). They concluded that movement took place generally between latest Devonian and Early Pennsylvanian and that it probably occurred in several distinct pulses taking place at different times and places within a broad region of central Nevada.

SUMMARY OF THRUST FAULTING

Major tectonism in the northern part of the Toquima Range is first recorded by thrusting of carbonate strata from west to east probably 10–35 miles (16–55 km). The result of this thrusting is the stacked series of thrust plates of carbonate rocks in the Ikes Canyon–Mill Canyon area. The youngest strata in any of these plates are Middle Devonian limestone, and this age serves as the older limit for the time of thrusting.

The upper plate of the regionally extensive Roberts Mountains thrust moved siliceous rocks eastward as much as 70 miles (110 km) to the present Toquima Range. This thrust plate rode over the stacked carbonate plates in the Ikes Canyon window and is the culmination of thrusting. It seems most likely that all the thrusting is related to the same tectonic episode, the Antler orogeny, with the most westerly and farthest traveled plate being the last to reach the regions at the longitude of the present Toquima Range. Thrusting had ended by Early Pennsylvanian time, for coarse clastic rocks of this age lap unconformably on the older rocks thrust into the area and across the thrusts. Erosion before the Early Pennsylvanian deposition stripped the upper plate of the Roberts Mountains thrust and exposed the lower (carbonate) plate in the Ikes Canyon area, suggesting that the thrusting had ended before the Early Pennsylvanian. In other places in central Nevada thrusting considered part of the Antler orogeny can be bracketed within slightly different limits, but as Gilluly and Gates (1965, p. 97) concluded "It is perhaps not surprising that a structure so large as the Roberts thrust should have been active at different times in different segments ***."

BASIN AND RANGE FAULTS

Basin and range faults that bound the Toquima Range are mostly buried by alluvium, and only at a few places are fault or faultline scarps visible along the range front. Evidence of basin and range faulting is mostly physiographic, as it is across most of the province, and a general view of the entire Toquima Range and bounding valleys gives the necessary perspective to evaluate this late Cenozoic structure. The vertical offset across the bounding faults of the Toquima Range is greater than the 2,000 feet (600 m) of topographic relief between the flat-lying tuffs at the top of the range and valley floors beneath which these same tuffs are buried. Because the entire range seems to be a rather discrete tectonic block, albeit cut by many small high-angle faults, it is most likely that the true basin and range faults responsible for its elevation are single large faults, several kilometres or more long and with several hundred metres of vertical offset. These faults are buried under alluvium 1 km or more from the present range front, which has retreated along a scarp in a manner typical of desert erosion.

MINERAL DEPOSITS

In the area of this report there are only four mines with recorded mineral production. Three of these are adjacent tungsten properties; the fourth is a small turquoise working.

TURQUOISE

The turquoise mine referred to as the Indian Blue 064
(Morrissey, 1968, p. 24) is about three miles (4.8 km) south of Petes Summit on the ridge south of Trail Canyon (SE¼ sec. 16, T. 15 N., R. 40 E., unsurveyed). The workings at the time of this writing consist of a pit a few metres across and about 6 m deep. Deeper workings are at the bottom of the pit as holes or shafts of unknown depth but probably less than 5 m. The turquoise occurs as small botryoidal nuggets coated with white carbonate material that makes them difficult to recognize as turquoise. Broken nuggets exhibit high-quality exceptionally blue and hard turquoise. Somewhat less than $10,000 worth of turquoise is reported to have been produced here (Morrissey, 1968).

The host rock is shale of the Vinini Formation, and the turquoise occurs on bedding planes. There is no obvious structural control or distribution pattern of the mineral; also, the area of the mine is not exceptionally altered, and there are no conspicuous intrusive bodies nearby. A large area of similar shaly Vinini Formation crops out in surrounding areas and should offer good ground for future turquoise prospecting.

TUNGSTEN

In the area near Spencer Hot Springs is a tungsten district comprising several once-productive mines and a number of minor prospect pits, trenches, and shafts. Within the Spencer Hot Springs district (unsurveyed T. 17 N., R. 46 E.), most of the tungsten production was derived from the Linka pit and shaft, the Conquest pit and incline, and the Hillside incline. Scheelite, associated with a rather continuous tactite zone, was discovered here in 1941, and the first ore was shipped to Battle Mountain, Nev., in 1943. Production continued intermittently until 1957. Over 4,000 tons of 0.98-percent WO_3 ore was shipped from the Linka from 1951 through 1956; approximately 60,000 tons of 0.4-percent WO_3 ore was milled at the Linka mill.

About 5 km^2 of bedrock almost completely surrounded by alluvium in Big Smoky Valley is exposed in the area of the Spencer Hot Springs district. Rocks related to the mineral district are limestone, probably the Ordovician Antelope Valley Limestone that is intruded and weakly metamorphosed by a granitic body of Middle Jurassic age (about 160 m.y. by potassium-argon dating). Noncarbonate rocks of the Ordovician Vinini Formation in thrust contact (Roberts Mountains thrust) with the limestone occur in this island of bedrock as does the Miocene Bates Mountain Tuff, but these formations are unrelated to mineralization or its distribution. The limestone is recrystallized as far as several hundred metres from the surface trace of the granite contact and contains a mineral assemblage typical of the albite-epidote-hornfels metamorphic facies. Bedding in the limestone, however, is still discernible. At the Linka pit, near the intrusive contact within the tactite zone, medium- to coarse-grained grossularite-vesuvianite-calcite-quartz bands are intercalated with epidote-ferroactinolite-tremolite series bands; scheelite, pyrite, molybdenite, and chalcopyrite(?) are confined to the garnet-bearing strata or pods. In general the metasomatic mineral banding is controlled by the original bulk composition of the sedimentary layering within the argillaceous carbonate sequence and is therefore parallel to bedding surfaces.

A second variety of mineralization, characterized by a scheelite-bismutite ($Bi_2O_2CO_3$)-molybdenite(?) assemblage, has formed in at least one place, and its distribution pattern does not seem related to the granite-limestone contact. This pattern probably reflects some late fracture system that breached the granite-carbonate contact and provided a conduit along which epithermal solutions could migrate, react with wallrock, and deposit their contained metals.

The exposure of mineral-bearing rock in the Spencer Hot Springs district is confined to a small area surrounded by alluvium or covered by a thin veneer of younger welded tuff. Scheelite occurs along much of the exposed intrusive contact, and exploration by shallow drilling has been carried out near the mines in an attempt to define the attitude, at depth, of these scheelite-bearing zones or to trace the ore bodies along strike. Some of the drill holes are spaced between the mines and along the presumed granite-limestone contact, but most of the adjacent areas along the contact, now masked by a thin veneer of alluvium, have not been explored in great detail by drilling or geophysical methods. The broader area around the mining district, also covered by only a thin layer of alluvium or welded tuff, has not been explored. Except in a very general way, the outline of the buried intrusive body and the distribution of limestone, the potential host for tactite deposits, are unknown.

GOLD

Although no gold has been mined from the region in the northern Toquima Range described here, the area has certain stratigraphic and structural characteristics that are similar to gold-producing districts elsewhere in central Nevada. Because of this, it warrants more than cursory examination for this metal. In particular the stratigraphic setting is like that of the large disseminated deposits at Carlin and Cortez about 100 and 50 miles (160 and 80 km) respectively, to the north. Carbonaceous limestone of the Roberts Mountains Formation, including the upper part of the formation, which is the gold-bearing lithologic unit at Carlin and Cortez, is present in the northern part of the Toquima Range in windows in the Roberts Mountains thrust—a structural setting that is also found at Carlin and Cortez.

Only a few miles south of the area described here, gold was mined in Northumberland Canyon from carbonate rocks similar to some of those in the Ikes Canyon window. These mines in limestone are clustered around a small granitic stock of Jurassic age (154 m.y. by potassium-argon; Silberman and McKee, 1971), and mineralization most likely is related to this body. There are no comparable intrusive bodies at the surface in contact with carbonaceous carbonate rocks in the region north of Northumberland covered by this report, but there are many small diorite and felsite dikes that might have introduced gold into a favorable host.

If the presence of intrusive rock is not significant in the introduction or concentration of gold in disseminated carbonate rock deposits but rather if the carbonaceous and silty limestone is the controlling factor, the Roberts Mountains Formation in the Petes Canyon window and this formation as well as some parts of the Antelope Valley Limestone in the Ikes Canyon window may contain disseminated gold of commercial concentration and quantity.

A few samples of carbonaceous limestone from the Petes Canyon and Ikes Canyon windows were analyzed by semiquantitative spectrographic methods to ascertain gold as well as other element content. None of these samples contained gold at the level of detection, although most contained silver in trace amounts. Mercury and arsenic, sometimes used as guides to heavy metal concentrations, showed no enrichment above levels considered background for the region (concentrations in other rock types in the region). Samples selected from near dikes or with visible alteration, however, yield traces of gold, silver ranging from 0.5 to 10 ppm, and a general increase in iron, manganese, and barium. Copper, zinc, and silver values were high near small prospect workings in the region, reflecting the presence of copper mineralization (usually pyrite or malachite) that served as a guide for the prospector. These local concentrations of obvious mineralization, however, are probably too small to be of potential commercial value.

BARIUM

Less than 10 miles (16 km) south of the area of this report in both East and West Northumberland Canyons, barite deposits of commercial size and grade are presently being mined. The barite of these bedded deposits was only recently discovered (Shawe and others, 1967, 1969) because of its general similar appearance to limestone common in the region. The bedded barite is of Ordovician age, represents rocks deposited in deep water (Shawe and others, 1969), and is part of the eugeosynclinal facies of the western assemblage that has been thrust eastward onto carbonate rocks of the same age. Rocks of the western assemblage, the upper plate of the Roberts Mountains thrust, are widespread in the area of this report and have never been explored in detail for barite. Most of the western facies strata examined during this study are bedded chert, shale, and some limestone. No barite beds were noted, but it is possible that some exist. The proximity to the Northumberland deposits and the presence of deep-water strata of the same age as those at Northumberland suggest that a more comprehensive study aimed at barite discovery might be productive in the northern part of the Toquima Range.

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