AN ABSTRACT OF THE THESIS OF

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Title: PALEOZOIC STRATIGRAPHY OF THE NORTHERN HOT CREEK RANGE, NYE COUNTY, NEVADA

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Abstract approved:

J. G. Johnson

Paleozoic dolomites, limestones, and quartz arenites, mostly of the shallow-shelf eastern assemblage, were mapped in the Morey Peak $7\frac{1}{2}$ minute quadrangle in the northern Hot Creek Range, Nevada. The Woodruff Formation, an allochthonous mudstone and chert unit of early Famennian age, lies in thrust contact with Middle Devonian dolomites, and was probably thrust from the west during the Antler Orogeny.

The seven oldest mapped units were deposited in shallow shelf to supratidal environments. The Lower Ordovician Antelope Valley Limestone consists of about 1000 feet of wackestones, packstones, and algal-nodule grainstones, in ascending order. The superjacent Eureka Quartzite, of probable Middle Ordovician age, is a quartz arenite about 200 feet thick, and was probably a beach-bar-dune deposit. The contact with the overlying dolomite of the Hanson Creek Formation, of probable Late Ordovician-Early Silurian age, is

transitional, and probably results from a slow deepening of marine waters. The Hanson Creek Formation is about 300 feet thick. Above the chert zone which marks the top of the formation is a coarse-grained, recrystallized dolomite which lacks primary features. An unnamed dolomite (at least 860 feet thick), probably equivalent to the upper Lone Mountain Dolomite, is the next youngest unit. It has a cross-bedded quartzose uppermost part, and may have been deposited supratidally. Progressively deeper depositional conditions are hypothesized for the overlying Kobeh and Bartine Members of the Lower Devonian McColley Canyon Formation. The Kobeh (480 feet thick) is a medium bedded dolomite of normal marine, shallow subtidal origin. The overlying Bartine is a very fossiliferous, argillaceous thin bedded wackestone to packstone about 90 feet thick. Conformably above the Bartine are two units, the Coils Creek Member of the McColley Canyon Formation (495 feet thick) and the Denay Limestone (795 feet thick), which were deposited at depths nearly exceeding those occupied by the deepest-dwelling shelly benthic fauna. Both are limestones with alternating platy and massive-weathering, generally thin beds. The Eifelian age Denay contains bioclastic packstone beds of probable debris-flow origin. It also contains an upper interval of beds with soft-sediment slumping features. This evidence of relatively steep slopes is unmatched in the pre-Mississippian rocks of the map area. The slopes were probably the result of westward progradation of shallow-water sediments, represented next higher in the column by stromatolite-bearing Middle Devonian dolomites (1700 feet thick). The superjacent Devils Gate Formation (at least 1000 feet thick) contains oncolites and oolites which are also indicative of very shallow sedimentation.

The map area falls well inside the regional boundaries of the Antelope Valley, Eureka, and Hanson Creek Formations. However, the units in the interval from the Unnamed Dolomite through the Middle Devonian dolomite are all close to their respective regional boundaries. The east side of the transition zone between shallow and deep environments apparently was located very close to the longitude of the Hot Creek Range during Early and Middle Devonian time. In this sense, the paleogeographic position of the map area is analogous to that of the well-studied Roberts Mountains, about 100 miles due north.

Paleozoic Stratigraphy of the Northern Hot Creek Range, Nye County, Nevada

by

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Harry Dodge of the U. S. G. S. suggested the project originally, and provided information from his own field studies. George Kendall assisted during the first few days in the map area, and introduced us to Nevada stratigraphy. Claudia DuBois handled all the preparation of the conodont samples.

I thank these paleontologists, who identified fossils from the map area: Gilbert Klapper, John L. Carter, Dick Flory, Ellis Yochelson, J. G. Johnson, and W. A. Oliver.

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PALEOZOIC STRATIGRAPHY OF THE NORTHERN HOT CREEK RANGE, NYE COUNTY, NEVADA

INTRODUCTION

Purpose

The thesis project was undertaken to determine what Paleozoic lithofacies are present at the longitude of the Hot Creek Range, and to relate this information to that of better-known areas in Eureka County and northern Nye County. Although reconnaissance maps indicate considerable exposure of Paleozoic rocks in the Hot Creek Range, little published information about this region is available. Such information is important for reconstructing sedimentologic regimes of the east side of the Paleozoic transition zone which separated deep water environments to the west from shallow water environments to the east. The relatively detailed map of part of the Hot Creek Range presented here provides an important reference for these reconstructions.

Location and accessibility

The map area is the northern two-thirds of the Morey Peak $7\frac{1}{2}$ minute quadrangle, located immediately north of Morey Peak in the Hot Creek Range, south-central Nevada. The quadrangle is located about 60 miles SSW of Eureka and 70 miles NE of Tonopah

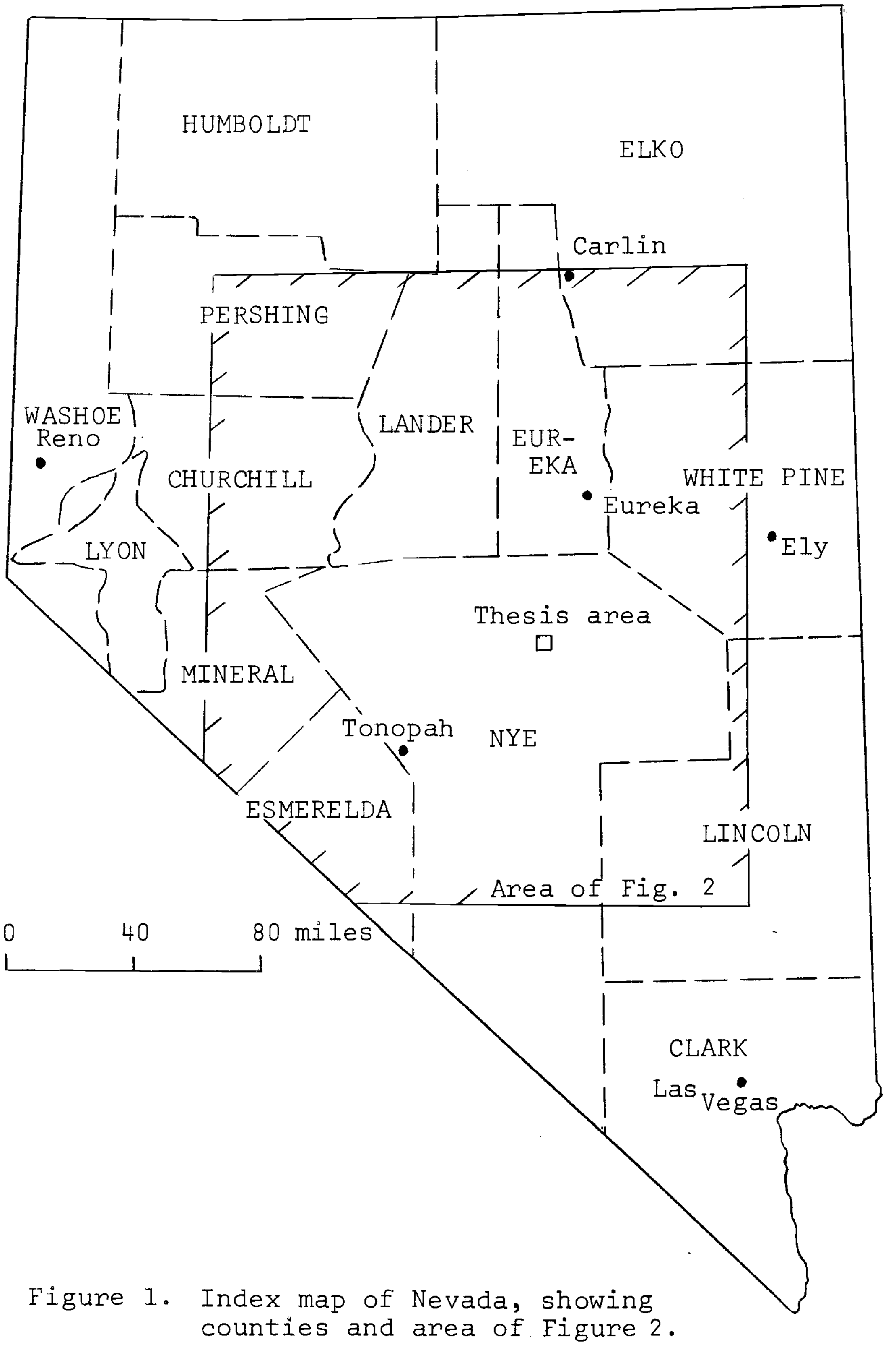

Figure 1. Index map of Nevada, showing counties and area of Figure 2.

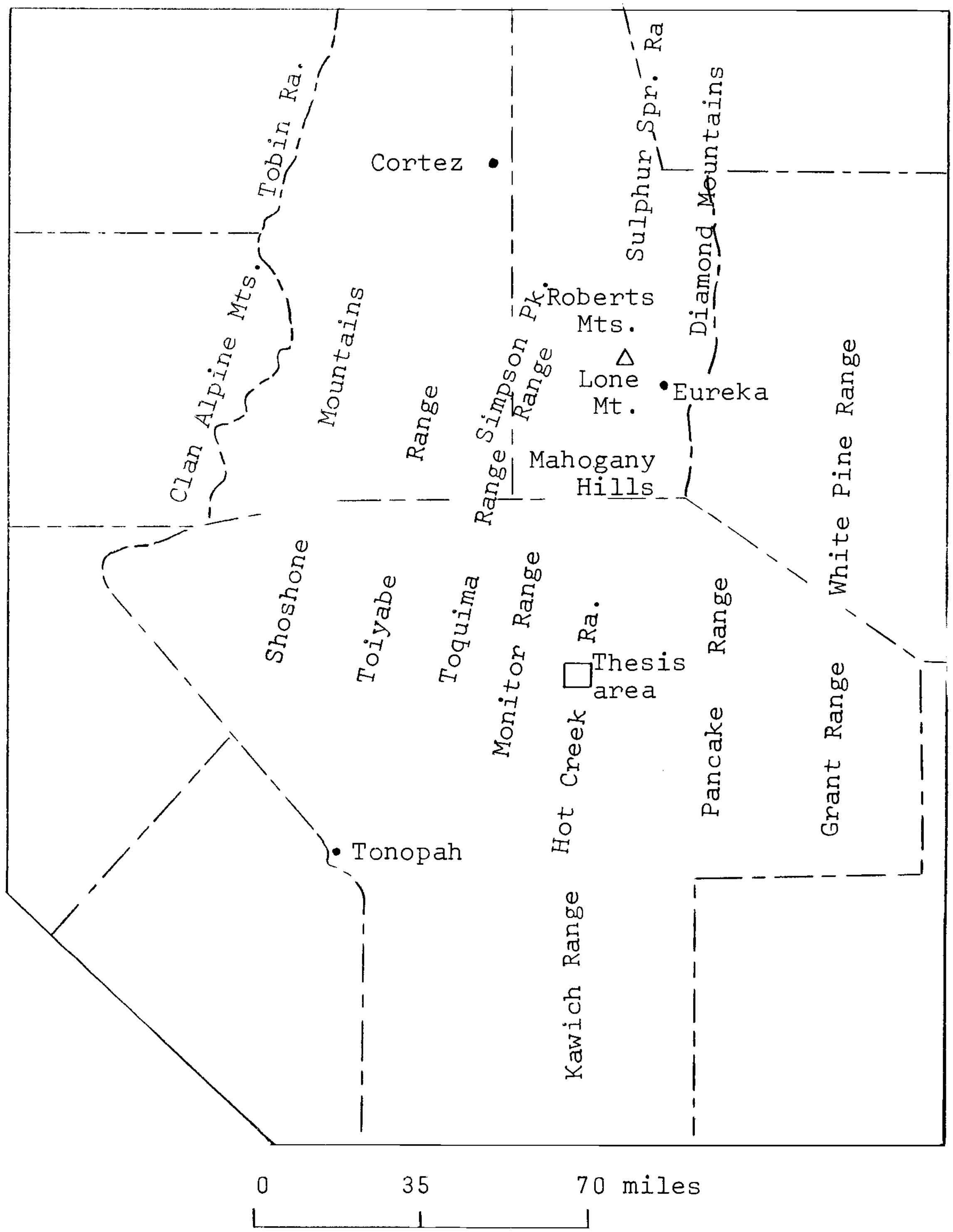

Figure 2. Index map of central Nevada, showing thesis area.

(Figures 1, 2). Easiest access is by dirt and gravel roads from Route 6, 22 miles south. Within the map area, dirt roads maintained once yearly by the Bureau of Land Management allow access to within 3 miles of most outcrops.

Previous work - vicinity of map area

Two previous studies in the map area have been made by U. S. Geological Survey geologists. Kelinhampl and Ziony (1967) did reconnaissance mapping on a 1:250,000 scale. Harry Dodge (of the Survey) has spent three field seasons in the Hot Creek Range. His work is not yet published. Nearby, the geology of the Moores Station Quadrangle (scale 1:48,000) was mapped by Ekren and others (1973) of the Survey. Lowell (1965) and Cook (1966) did thesis studies of Paleozoic rocks in the vicinity of Hot Creek Canyon, 10 miles south of the map area. Important work by Merriam (1963) 40 miles north in Antelope Valley, established the Paleozoic section for the vicinity.

Geologic setting

During early and middle Paleozoic time, the site of the present Great Basin was part of the north-trending Cordilleran Geosyncline, which extended from southern California to the Canadian Arctic (Roberts, 1972). In a plate tectonic context, this zone is believed to have been a series of marginal basins (between continent and offshore island arcs) (Churkin, 1974; Rogers and others, 1974; Johnson and Potter, 1975). Within this area in Nevada, three depositional belts are commonly recognized (Roberts and others, 1958). A north-trending zone, about 50 miles wide, known as the transitional belt, separates western assemblage rocks (deep-water limestones, cherts, volcanics, and fine clastics) from eastern assemblage rocks (shallow shelf dolomites, quartz arenites, and shelly limestones). The transitional belt is characterized by fine clastic and carbonate deposits, many of which have characteristics indicating deposition on a slope. Some of these slope deposits are believed by several workers to have been deposited on the continental slope (Winterer and Murphy, 1960; Smith and Ketner, 1975; Johnson and Potter, 1975). The transitional belt was located near the longitude of the Roberts Mountains (Eureka County) in Silurian-Devonian time; during Cambrian-Ordovician time, it was located some 25 to 75 miles farther west.

Roberts and others (1958) noted that the transitional zone oscillated east and west through time. Resultant facies patterns show corresponding interfingering relations on a large scale. In the map area, this is illustrated by relatively deep-water Denay Limestone rocks overlying and overlain by eastern assemblage shallow-water carbonate rocks.

In Late Devonian-Early Mississippian time, this pattern of sedimentation throughout the Great Basin was abruptly terminated by the topographic and sediment-contributing effects of the Antler Orogeny (Johnson, 1971; Brew, 1971; Poole, 1974). An elongate north-northeast trending highland rose in west-central Nevada, and was accompanied by folding and large scale eastward thrusting of western assemblage rocks. Largest displacement apparently occurred along the Roberts Mountains Thrust. Remnants of allochthonous plates exist today as far east as the Sulphur Springs and Pinyon Ranges, Before, after, and probably during the thrusting, coarse and fine grained clastic debris was shed eastward from the orogenic belt. These "flysch deposits" (Poole, 1974) were referred to by Roberts and others (1958) as the "overlap assemblage" because they were deposited over the long-existent (although oscillatory) boundaries between transitional, eastern, and western assemblage rocks.

The thesis area is located near the west edge of the zone where eastern assemblage rocks were deposited through much of the middle Paleozoic. The Antelope Valley Limestone, Eureka Quartzite, Hanson Creek Formation, Unnamed Dolomite, and the Kobeh and Bartine Members of the McColley Canyon Formation represent typical eastern assemblage facies shallow-water deposits. Coils Creek and Denay limestones probably represent an incursion eastward of the transitional belt. Upwards in the section, Middle Devonian dolomites and the Devils Gate Limestone are representative of shallow eastern facies environments. Mississippian limestones and fine terrigenous clastic rocks contain detrital chert, plagioclase, and volcanic rock fragments, probably derived from the Antler orogenic highland to the west, and thus record a fundamental change in sedimentologic regimes in the map area.

Figure 3 is a summary stratigraphic section for the map area.

Methods

Eight weeks were spent in the field in the summer of 1974 by the writer and his assistant, C. S. Eldridge, an undergraduate at O. S. U. Fieldwork involved mapping Paleozoic units and measuring and describing sections. Samples were collected for lithologic and paleontologic study. Studies of lithology included preparation and analysis of thin sections; etched, stained, and oiled slabs; and acetate peels. Mechanical size analyses and determination of oxidizable organic content were undertaken for selected samples. X-ray diffraction analyses were performed on five carbonate samples to check results of carbonate staining procedures.

Large (30 lb) samples were collected in the field for the purpose of obtaining brachiopods and conodonts. Brachiopod samples were prepared as follows. Silicified brachiopods were etched free of enclosing carbonate rock by immersion in concentrated hydrochloric acid. Non-silicified brachiopods were freed by cracking the sample

AGE	ROCK UNIT	THICKNESS (feet)
Tertiary	Volcanics	several thousand
Triassic	Triassic rocks	200 (incomplete section)
Mississippian	Mississippian rocks	750 (incomplete section)
Devonian	Devils Gate Limestone	1500
	Woodruff Fm.	
	Middle Devonian Dolomite	1700^{+}
	Denay Limestone	795
	McColley Canyon Fm. — Coils Cr Mbr	495
	McColley Canyon Fm. — Bartine "	90
	McColley Canyon Fm. — Kobeh Mbr	480
Silurian?	Unnamed Dolomite	860 (incomplete section)
	Silurian? Dolomite	?
Ordovician	Hanson Creek Fm.	300
	Eureka Quartzite	200
	Antelope Valley Limestone	1000 (incomplete section)

Figure 3. Summary stratigraphic section for the Morey Peak Quadrangle.

using a hydraulic rock-splitter. Cleaned and sorted fossils were submitted to J. G. Johnson, Oregon State University, for identification and age determination. Two Mississippian brachiopod collections were identified by John L. Carter of the Carnegie Museum, Pittsburgh. Conodont samples were prepared by Claudia DuBois. After dissolution of the carbonate matrix in dilute formic acid, the insoluble residue was separated in heavy liquid and the conodonts were hand-picked from the heavy fraction. Gilbert Klapper, University of Iowa, made identifications and age determinations of the conodonts. A small number of tabulate coral samples (thin sections, peels, and slabs) were identified by Richard A. Flory, California State University at Chico.

Terminology

The sandstone classification of Gilbert (1955) and the carbonate classification of Dunham (1962) were used for hand sample and thin section description. The term "micrite" is retained from the earlier carbonate classification of Folk (1959) and is used as a part of some sample descriptions.

Bed thickness descriptions are based on the following scale: 1 to 4 inches is thin-bedded, 4 inches to $1\frac{1}{2}$ feet is medium-bedded, and more than $1\frac{1}{2}$ feet is thick-bedded.

Crystal size designation for recrystallized carbonates follows the scheme proposed by Folk (1965, p. 25): .004 to .016 mm is very finely crystalline, .016 to .062 mm finely crystalline, .062 to .25 mm medium crystalline, .25 to 1.0 mm coarsely crystalline, and 1.0 to 4.0 mm very coarsely crystalline.

Rock colors were determined by comparison with the Rock Color Chart published by the Geological Society of America (1963).

TOPOGRAPHY, GEOMORPHOLOGY, AND VEGETATION

The center of the map area is on the crest of the Hot Creek Range, just north of its highest point (Morey Peak, 10,246 feet elevation). The extreme northeast and northwest corners of the area are portions of Quaternary bajadas in Hot Creek Valley and Little Fish Lake Valley, respectively. Topography above the bajada levels is characterized by steep-walled NNE trending valleys with typical relief of about 900 feet between valley floor and ridge crest. Total relief in the area is 2985 feet, from a maximum of 9825 (Mahogany Peak) to a minimum of 6840 (Luther Waddles Wash). The two largest valleys, North Sixmile Canyon and Big Cow Canyon, are each several miles in length. Of the hundreds of smaller valleys, many are either parallel or perpendicular to the general NNE trend of the range.

Dominant among geologic controls of landforms are NNE trending normal faults along which the major valleys have been eroded. Other controls, having a lesser effect on relief but important in modifying the topography locally, include smaller faults not associated with NNE trend, and rock type exposed at the surface. In some places, every small canyon is the site of a normal fault (see the area near the base of the measured section ½ mile east of Big Cow Canyon). The importance of rock type as a topographic control may be summarized by the generalization that quartzite, chert, dolomite, and volcanic rock are the dominant ridge formers (the importance of the volcanics in this respect is also related to their relative youth). Generally, mudstones, limestones, and the basal Tertiary volcanic unit (a tuff) are less resistant to weathering than the quartzite, chert, and dolomites.

Flint (1957, p. 309) listed evidence of Pleistocene glaciation in the Toiyabe Range (about 35 miles west), but no evidence of glaciation was noted in the map area.

Vegetation is typical of that found on mountains throughout the Great Basin. Mountain Mahogany, pinyon pine, limber pine, sage, rabbit brush, and locally, aspen, predominate. No reliable relationship could be discerned between rock formations and vegetation types. Types of vegetation appear instead to be a function of altitude and availability of moisture. For example, sagebrush is confined to valley bottoms, and aspen groves exist only at seeps and springs on high slopes, regardless of bedrock type.

ANTELOPE VALLEY LIMESTONE

General statement

The oldest exposed unit in the map area is the Lower Ordovician Antelope Valley Limestone, which crops out beneath (west of) the Eureka Quartzite of Crescent Hogback in the western part of the map area. Merriam (1963) assigned formational rank to the Antelope Valley Limestone, and designated a type section on the west side of the Antelope Range (45 miles north of the Morey Peak Quadrangle). The formation there consists of medium-bedded, fine-grained, medium bluish gray limestone. Merriam measured 1000 feet of section at an incomplete exposure on Martin Ridge (Antelope Valley), and estimated that the formation in its entirety is more than 1200 feet thick. An informal subdivision of the formation was made by Merriam, using three fossil zones, which are (oldest to youngest): Orthidiella Zone, Palliseria Zone, and Anomalorthis Zone. These faunal zones correspond fairly well to lithologic divisions. At the type section, the lowermost 75 feet of the formation (Orthidiella Zone) is thin-bedded, argillaceous limestone. The Palliseria Zone, 650 feet thick at the type section, is thicker bedded, less argillaceous, and more resistant to weathering - locally forming small cliffs. This zone is characterized by dominance of gastropods, particularly Palliseria longwelli (Kirk) and subordinate large Maclurites. Additional genera of both gastropods and brachiopods are typically present, as are ovoidal algal nodules.

The Anomalorthis Zone is 350 feet thick at the type section, and characteristically is thinner-bedded than the unit below. It weathers to a mottled gray-brown, and is locally rich in small high-spired gastropods, brachiopods, stony bryozoa, and algal nodules.

Thickness

The best-exposed section of Antelope Valley Limestone in the map area is complicated by at least one normal fault ($\frac{1}{4}$ mile SE of Summit 8012). Thickness estimates are further hindered because the upper contact is obscured by large quartzite talus blocks from the overlying Eureka Quartzite. Nevertheless, the formation's thickness along the north side of Little Cow Canyon was roughly estimated to be 1000 feet.

Contacts

The lower contact of the Antelope Valley Limestone is a normal fault which juxtaposes much younger, Middle Devonian (Denay) rocks, and Antelope Valley beds. The upper contact, as noted above, is obscured by thick Eureka Quartzite talus. The Copenhagen Formation, a thin, fossil-rich carbonate unit which is stratigraphically above the Antelope Valley and below the Eureka Quartzite in the

Monitor Range, was not noted in the map area. Although its stratigraphic position is covered with talus, there is no evidence of its distinctive lithology in the local float. It is not present in the Antelope Range, 40 miles north along depositional strike (Merriam, 1963), and probably does not exist in the map area. However, it has been reported in Hot Creek Canyon (10 miles south) by Lowell (1965).

Middle unit

Rocks typical of the *Orthidiella* Zone are not exposed in the map area; they have been downdropped along a NNE trending normal fault which strikes approximately parallel to the strike of the Antelope Valley beds. The lowest exposed Antelope Valley rocks correspond lithologically to Merriam's middle unit (*Palliseria* Zone). Typically these rocks are well-bedded wackestones weathering grayish orange pink. Thickness of beds ranges from one inch to one foot; commonly several superimposed thick beds form small cliffy outcrops. This is especially true where the limestone is rich (up to 20%) in fine quartz sand grains. The tops of some sandy beds are extensively burrowed parallel to the bedding plane. The 3 to 5 mm thick burrows commonly radiate from a common point. Occasional platy-weathering lime mudstone beds are encountered, and these also are heavily burrowed as described above.

Fossils in this unit are diverse and locally abundant. Brachiopod valves, pelmatazoan fragments, and internal molds of gastropods are common throughout the unit. Particularly large numbers of gastropod molds were noted in platy-weathering beds on Summit 8012. One planispiral mold 12 cm in diameter was found. Petrographically, these beds are pellet wackestones. Micrite pellets of very fine sand size constitute about 15% of the rock, with small shell fragments, very fine sub-angular to angular quartz grains, and glauconite collectively constituting less than 5% of the rock. Minor limonite cement imparts the characteristic pale yellow brown rock color on weathered surfaces.

Upper unit

The upper unit of the Antelope Valley consists of approximately 400 feet of medium-crystalline dolomite and dolomitic limestone overlain by about 100 feet of bioclastic limestone. The dolomite and limestone of this unit weather recessively compared to the Eureka Quartzite above and the *Palliseria* Zone below.

The dolomite occurs in 2 to 6-inch beds. On outcrop, dark-colored ghosts of algal-encrusted shell fragments can be seen. In thin section, these ghosts are seen to consist of finely crystalline, dolomite and limonite cemented, cloudy rims around shell fragments. Medium-grained crystalline dolomite with about 2% fine-grained detrital quartz and plagioclase make up the remainder of the rock.

Comparison of this texture with that of the overlying bioclastic lime grainstones indicates that the origin of the dolomite beds was by replacement of the lime grainstones. Primary structures and textures have been almost completely obliterated by the dolomitization (see Figures 4 and 5).

The distinguishing feature of the grainstone beds of the uppermost Antelope Valley Limestone is the abundance of algal-encrusted shell debris. The encrustations vary from thin layers less than one millimeter thick on brachiopod fragments to ovoidal bodies 1.5 cm by 2 cm by 3.5 cm. Locally these encrustations constitute up to 30% of the rock. The coatings become thicker and more evenly concentric up-section. In thin section and on acetate peels, loosely intertwined cylindrical tubes of constant .03 mm diameter are seen to make up the concentric encrustations (see Figure 6). These tubes are the typical form of the blue-green alga Girvanella (J. H. Johnson, 1961, p. 198, 199). Each lamination is about .5 mm thick, and commonly the laminations enclose very fine-grained detrital quartz. The encrustations are stained yellowish brown by limonite.

Where the nodules are most abundant, they are in grain contact, or nearly so. The remainder of the rock is shell debris and sparry calcite cement, in roughly equal amounts. The shell debris is a grain-contact hash of fragments of pelmatozoans, stony bryozoa, brachiopod valves, and gastropod shells. Pelmatozoan fragments

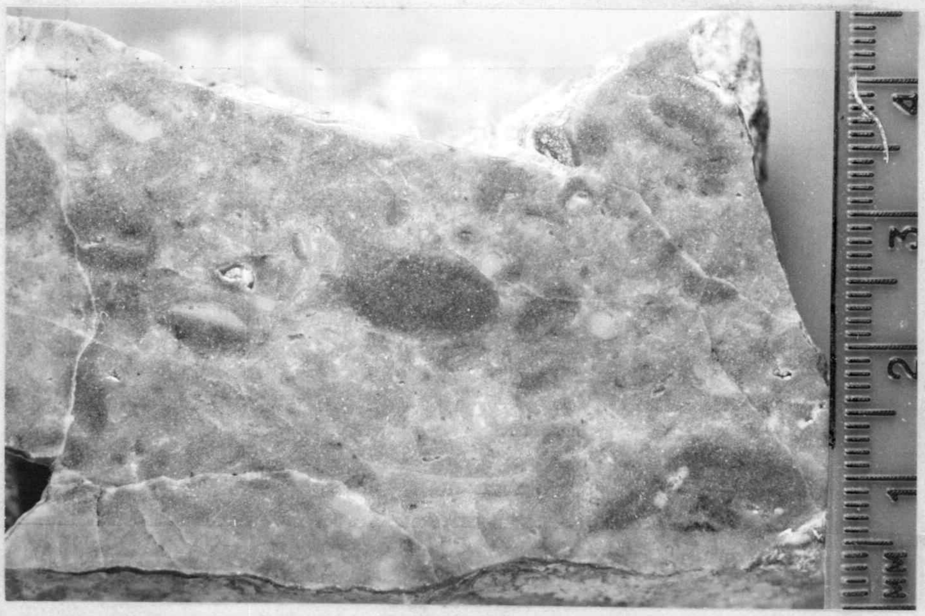

Figure 4. Faintly visible ghosts of features shown in Figure 5. Dolomite from the upper part of the Antelope Valley Limestone.

Figure 5. Ovoidal Girvanella nodules with shell fragments as nuclei. Note incomplete algal coating on some (lower right-hand corner of slab), whereas others have regular coatings of constant diameter. The matrix is shell hash, predominately echinoderm fragments. Note that the largest shell fragments are the nuclei of algal nodules. The rock is a bioclastic algal-nodule grainstone. Paper clip gives scale.

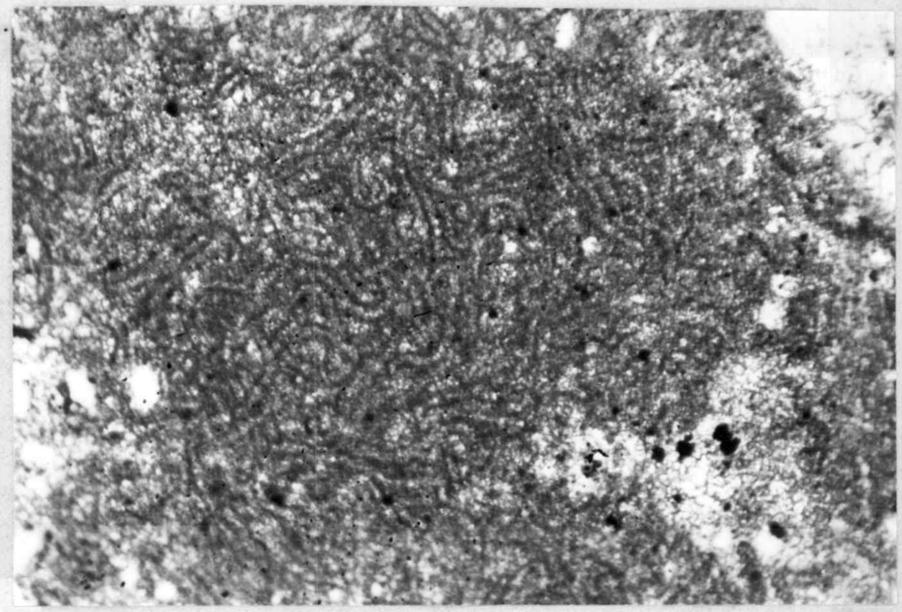

Figure 6. Photomicrograph (X 50) of Girvanella. Note characteristic constant-diameter loosely intertwined tubes. Diameter of each tube is about .03 mm.

with optical overgrowths of sparry calcite make up about 45% of the rock.

Depositional environment

The lithologies and fossils in the Antelope Valley Limestone suggest shallowing marine waters during deposition. Oldest rocks exposed (the middle unit of this report) are locally fossiliferous wackestones. The extensively burrowed bedding planes, the lack of winnowing, and the abundant pellets all suggest relatively low current energies; the normal marine brachiopod-pelmatozoan-gastropod fauna suggests open, relatively shallow marine shelf-type environments.

The upper unit, characterized by algal-nodule bioclastic grainstones (some of which are replaced by dolomite) was formed in very shallow water with much higher current energies. Bioclastic components consist of pelmatozoan, brachiopod, gastropod, and trilobite debris, suggesting a normal marine environment of relatively shallow depths. However, the algal nodules are the key to interpreting the depositional environment of the upper unit.

The shell fragments not encrusted by more than about .5 mm of algal material are significantly finer-grained (.25-2 mm) than the shell fragments forming the nuclei of the large algal nodules (2-12 mm). This suggests that thick algal encrustations strengthened certain shell fragments, protecting them from abrasion and breakage in what must have been a high-energy environment, as indicated by broken shells and lack of fine-grained material.

Further indication of relatively high-energy depositional environment comes from analysis of the orientation of the nodules. The long axes of most are sub-parallel to the bedding plane. However, close inspection of acetate peels and oiled slabs shows that many nodules grew by concentric addition of algal laminations while the shell fragment nucleus was in a concave-down position, then was flipped over before burial. This situation can be recognized because some shell fragments have well-developed algal encrustations on the convex surface and around the shell edge, but little encrustation on the middle concave portion. In this situation the convex side is inferred to have been the top, because light-dependent algae could not have grown beneath the shell.

The presence of many shell fragments evenly encrusted with concentric algal laminations necessitates rolling by current or wave action. This information, together with the dominant winnowed, grain-supported abraded shell-hash lithology, suggests a very shallow subtidal or low intertidal depositional environment. Similar conclusions in studies of algal nodules are reported by LaPorte (1967) and Logan and others (1964).

The fact that algal encrustations are thicker and more evenly concentric up-section suggests upward shoaling within the upper unit. Thus the Antelope Valley Formation in the map area records a probable shallowing upwards from below wave-base shelf depths (middle unit) to low intertidal or very shallow subtidal positions (top of the upper unit). This trend sets the stage for the deposition of the overlying Eureka Quartzite at or above mean sea level.

Regional perspective

The transitional belt was not near the longitude of the Hot Creek Range during Antelope Valley time. Bassler (1941) described a time-equivalent richly fossiliferous shelf limestone in the Toquima Range about 35 miles west of the map area. Webb (1956, 1958) showed that the Antelope Valley Limestone extends eastward into Utah.

The Ordovician transitional belt must have existed at least as far west as the Toquima Range. This supports the observation of Johnson and Potter (1975) that the position of the Silurian-Devonian transitional belt which occurs immediately west of the map area represents a significant eastward shift of the continental slope.

EUREKA QUARTZITE

General statement

The thick, resistant quartz-arenite composing the Eureka Quartzite crops out almost continuously above the Antelope Valley Limestone, and forms the caprock of Crescent Hogback, in the western part of the map area. The Eureka was named by Hague (1883, p. 253, 262), but the original type locality near Eureka, Nevada is severely deformed, and subsequent workers have regarded Lone Mountain as the "type" locality for the Eureka (Merriam, 1963, p. 29. This formation may have the largest areal distribution of any Great Basin Paleozoic unit. It is easily recognized from central Nevada (Monitor Range) to western Utah to the Inyo Mountains of southern California.

Lithology and thickness

In the map area, as at other Great Basin localities, the Eureka is quartz arenite composed almost solely of medium-fine to fine quartz grains cemented by quartz in the form of overgrowths and as pore fillings. The quartz sand is commonly well-sorted and is sub-angular to well-rounded. Quartz overgrowths commonly obscure original rounding and sorting of the quartz grains.

The extremely tight cementing by quartz overgrowths and pore fillings may be the cause of the lack of well-defined bedding on outcrop. Locally, 12 cm high sets of planar cross-laminations can be seen on weathered surfaces, but other evidence of bedding is lacking. The general appearance of the outcrops is massive light-brown weathering, blocky-fractured quartz arenite. Color of the fresh rock is white. Outcrops are all cliffy exposures 10 to 50 feet in height.

Exposures of cross-laminations were not adequate for paleocurrent determinations. The cross-laminations are best exposed on talus blocks west of the ridge summit, but these, of course, are not useful for obtaining attitudes.

No fossils were observed in the Eureka. The formation is about 200 feet thick.

Petrography

Thin sections show that the rock is more than 98% quartz grains and quartz overgrowths. The average grain diameter is 0. 3 mm. Those grains whose original shape could be discerned are sub-angular to well-rounded. About 15% of the quartz grains exhibit undulatory extinction, indicating that they have been strained. Some of the quartz grains have small tourmaline inclusions. Chert-cemented veins containing well-rounded quartz grains of sedimentary aspect are common and their origin is difficult to determine. Perhaps this lithology represents a stage that most of the rock has already passed through. A few very well-rounded, fine grains of brown tourmaline and zircon are present in each thin section.

Age

The age of the Eureka Quartzite has not been directly determined because of the nearly universal absence of fossils. The possible age range has been determined in Antelope Valley by Merriam (1963) by bracketing the Eureka between fossil zones of the Antelope Valley Limestone below and the Hanson Creek Formation above. On this basis, Merriam (p. 30) assigned a late Middle to early Late Ordovician age to the Eureka at Antelope Valley.

Contacts

The lower contact with the Antelope Valley Limestone is obscured because talus from the Eureka covers the upper part of the recessive-weathering Antelope Valley. No major differences between the attitudes of the two units were observed. The base of the Eureka elsewhere in Nevada overlies a significant disconformity. At Roberts Creek Mountain, 100 miles north of the map area, the Eureka lies upon the lower part of the Goodwin Limestone. Antelope Valley Limestone, Ninemile Formation, and part of the Goodwin Limestone are missing. At Cortez (120 miles NNW) the time gap may be much longer (Merriam, 1963, p. 30). At Bellevue Peak (Antelope Valley area) subaerial exposure is indicated by edgewise mud-breccia between Antelope Valley strata and the overlying Eureka (Merriam, p. 30).

The upper contact in the map area appears to be gradational with the overlying dolomite beds. Increasing dolomite content relative to quartz grains is reflected on outcrop by the recessive-weathering light gray beds on the east flank of Crescent Hogback. The contact was arbitrarily drawn to separate blocky, cliffy quartz arenite beds from subdued-weathering decreasingly quartzose dolomites up-section.

Depositional environment

Good sorting, well-rounded grains, and medium-scale planar cross-stratification of the Eureka Quartzite collectively suggest a beach-bar-dune environment. The general lack of well-defined bedding or lamination, described earlier, may be a primary feature, because several micro-environments of a beach regime might produce non-bedded strata. A pool receiving wind-blown sand is an example of one such micro-environment. Unfortunately, exposures were not adequate for a definitive study of cross-lamination types. Subaerial exposure, with wind-formed dunes, cannot be ruled out. Grain sizes of the sand fall within the wind-transport range, and medium scale, planar cross-stratification (sets 12 cm high; 25°

average dip) could have been wind-formed. The very broad geographical distribution of the Eureka (more than 10, 000 mi^2) suggests that the gross environment of this depositional regime was a broad, emergent, slightly sloping shelf, and that the formation is diachronous. Occasional eastward incursions of marine conditions are suggested by rare intercalated dolomite layers in the Inyo Mountains of California, and "an exceptional occurrence of coarals at Cortez, Nevada" (Merriam, 1963, p. 30).

Source

The well-sorted, well-rounded, quartz-tourmaline-zircon association is characteristic of supermature sandstones (Folk, 1968). The provenance of the Eureka remains unknown, but must have been a previously existing sandstone. The monomineralic nature of this formation is unmatched lower in the section in the Great Basin except for the very extensive transgressive quartz arenites of Late Precambrian to Early Cambrian age (Prospect Mountain Formation and equivalents). The quartz grains of the Eureka are coarser than the relatively rare quartz grains in the Antelope Valley Limestone. This profound lithologic change, from normal marine carbonates to pure quartz sands, has not been adequately explained in the literature. Webb (1956) demonstrated that the Pogonip Group of central Nevada (of which the Antelope Valley Limestone is the youngest formation) is time-correlative with shale and quartz-arenite rocks in western Utah. Webb interpreted the shale and quartz-arenite association as having formed in a beach-bar-lagoon regime. The terrigenous material of this regime apparently was limited to the area of western Utah, and did not interfere with carbonate sedimentation to the west during Chazyan (Pogonip) time (Webb, 1956, p. 21).

The blanket-like quartz sands of the Eureka quartzite represent significant westward extension of terrigenous sedimentation during Middle Ordovician time. Webb (p. 19) wrote that "the cratonal interior emerged, shedding regressive detritus into the western sea." He interpreted the upper portions of the Eureka as transgressive deposits of reworked "regressive detritus." His lithofacies maps (Figures 5, 6, 7, 8, 1956) clearly show westward progradation of Middle Ordovician quartzites from western Utah into eastern Nevada. Implicit in this interpretation is the existence of an eastern source for the quartz sands, and Webb (p. 32) speculated that the sources were basal Cambrian and latest Precambrian quartz sandstones of the Rocky Mountain area, plus emergent pre-Eureka Ordovician sands from western Utah (Swan Peak Quartzite).

This interpretation accounts for the super-maturity of the Eureka sands and their distribution with time. It also is compatible with interpretation of the depositional environment as a beach-bar-dune complex migrating across a broad, shallow, emergent shelf.

In addition, Webb's interpretation is compatible with the sedimentary sequence concept of Sloss (1963). A sequence is a very thick and widespread rock-stratigraphic unit, deposited during a specific transgressive-regressive cycle lasting tens of millions of years. The resultant sedimentary packet is bounded above and below by time-transgressive unconformities which widen (in the time sense) toward the interiors of continents. In this context, the Eureka sands represent the initial transgressive phase of the Tippecanoe Sequence (Sloss, 1963). The up-section shallowing suggested by features within the Antelope Valley Formation would thus be indicative of the waning of the preceding Sauk Sequence. However, Eureka sands would have had to be distributed over the emergent shelf prior to transgression or they would not have been available along the transgressive shoreline. For this reason, the hypothesis of Webb that the lower Eureka was regressive may be valid. Alternatively, during the low stand of sea level between the Sauk and Tippecanoe sequences, rivers may have brought the sand from sources to the east, and longshore currents accomplished the distribution of the sand along the eastward-migrating shoreline. In this case the entire Eureka could be transgressive.

Regional perspective

As previously noted, the Eureka extends more than 100 miles east, and about 25 miles west of the map area (McKee and Ross, 1969). Thus, the transition belt must have existed well to the west of the map area during Eureka time.

HANSON CREEK FORMATION

General features

Rocks here assigned to the Hanson Creek Formation crop out on the dip slope (east flank) of Crescent Hogback. The Hanson Creek formation in the map area is a very light gray, recessive weathering, medium crystalline dolomite containing microscopic bioclastic fragments.

Defined by Merriam (1940, p. 11) at a type section along Pete Hanson Creek in the Roberts Mountains, the Hanson Creek Formation and its equivalents crop out immediately above the Eureka Quartzite throughout its areal extent. In Utah and eastern Nevada, the Hanson Creek is completely dolomitic, but the lithology changes to limestone west of a line running approximately from Cortez to the Roberts Mountains, west of Lone Mountain, through the Mahogany Hills, and along Little Fish Lake Valley immediately west of the map area. Along this line the dominant lithology is dolomitic limestone.

Thickness at the type locality is 560 feet. At Lone Mountain, the dolomitic Hanson Creek strata are 318 feet thick. Exposures of the formation at Whiterock Canyon (Antelope Valley) are 350 feet thick (Merriam, 1963). Formation thicknesses are measured from the top of the uppermost vitreous quartzite beds of the Eureka to the base of the "marker chert" beds of the basal Roberts Mountain

Limestone above (Merriam, 1940, p. 10). Lower contact with the Eureka is actually transitional, involving an increasing proportion of carbonate material relative to quartz grains over an interval of about 15 feet (Merriam, 1963, 1940; Webb, 1956, 1958; Kirk, 1933).

Age of the Hanson Creek was regarded by Merriam (1940) to be Middle and Late Ordovician. The upper age limit was changed by Mullens and Poole (1972, p. 2) to Early Silurian on the basis of conodonts identified from the upper part of the formation in Eureka County.

Contacts

The Hanson Creek Formation was defined to include all rocks stratigraphically above the Eureka and below the basal chert of the Roberts Mountain and Lone Mountain Formations. In the map area, the lower contact is easily mapped because the top of the Eureka is topographically and lithologically distinctive. The exact location of the upper contact is conjectural because neither the Lone Mountain nor the Roberts Mountain Formations were recognized (see the next unit of this report). However, a very distinctive chert unit about 300 feet above the base of the Hanson Creek is interpreted as the basal chert of the Lone Mountain Dolomite. The chert unit consists of 10 to 20 feet of resistant, pale yellowish brown to very light gray chert in 1 to 2 inch beds with occasional intercalated yellowish gray dolomite lenses of similar thickness. The chert unit is best exposed on a small ridge 1500 feet WNW of Summit 7845, south of Big Cow Canyon, in the NW 1/9 of the map area. South of this locality the chert zone is not seen in outcrop, but is believed to occupy the base of the prominent arcuate-plan valley approximately 1500 feet east of the crest of Crescent Hogback along its entire length. The ground in this valley bottom is littered with small pieces of gray chert identical to those from the outcrops described above. The chert does not crop out in the lower ridges east of this valley, except at one place just east of the Big Cow Canyon road. This outcrop is on the upthrown side of a normal fault, the trace of which runs north-south through Big Cow Canyon.

Lithology and outcrop characteristics

In the map area, the Hanson Creek dolomite is poorly exposed, cropping out on the dip slope of Crescent Hogback. The rock is medium grained dolomite, indistinctly bedded, with bedding-plane partings at approximately 2-inch intervals. The weathered surface is yellowish gray and the fresh surface is light gray to light olive gray. Occasional thin chert lenses were observed on outcrop. No fossils were seen in the field.

In thin section, however, bioclastic elements in the rock are apparent. Small plate-like objects (less than 1 mm in diameter) with reticulate networks of isotropic material locally make up 10% of the rock. The objects are probably echinoderm fragments or red algae fragments replaced by an isotropic substance (phosphatic ?). Fragments of bryozoa make up 5% of some thin sections. The balance of the framework consists of well-sorted, very well-rounded to subangular quartz grains in the very fine sand size range. This quartz component decreases up-section, from more than 50% near the basal contact to 10% or less about 60 feet higher.

The dolomite occurs as somewhat cloudy intergrown crystals .2 to .3 mm in diameter. Porosity of 1%, due to dissolution of 1 to 2 mm-long shell fragments, is observed in some thin sections.

The chert which marks the upper contact of the Hanson Creek dolomite consists of microcrystalline to cryptocrystalline quartz, with a discontinuous limonite coating which imparts a yellow tinge to the weathered surface. In thin section, the chert is mottled due to irregular variations in crystal size. A trace of microcrystalline carbonate in the thin sections studied suggests that the chert formed by replacement of fine-grained limestone or dolomite. No fossils were observed in thin sections or on outcrop.

A unit very similar to the Hanson Creek as described here was mapped by Quinlivan and others (1974) about 20 miles to the east in the Portuguese Mountain Quadrangle. They mapped the unit as the Ely Springs Dolomite, which is the equivalent of the Hanson Creek in southeastern Nevada. Quinlivan and others describe an inconspicuous oolite zone near the top of the formation, as well as *Girvanella*-like forms in the sandy dolomite near the base. Total thickness in the Portuguese Mountain Quadrangle is about 310 feet, nearly identical to the thickness in the map area.

Depositional environment

Few features of environmental significance were observed in the Hanson Creek Formation. Early deposition was undoubtedly very shallow, as indicated by the transition from Eureka lithology upwards to pure dolomite. The dolomite has been recrystallized, as evidenced by straight crystal boundaries and the general interlocking crystal mosaic. The presence of bryozoan fragments suggests relatively shallow normal marine environments.

Regional perspective

The Hanson Creek Formation, and its equivalents, the Ely Springs and Fish Haven Dolomites, are present over a very large area from the Toiyabe Range eastward into Utah. Johnson and Potter (1975) noted that the transition belt must have been west of the Toiyabe Range, which is about 50 miles west of the map area (see Figure 1).

ROCKS OF SILURIAN? AGE

No rocks in the map area have been positively identified from the stratigraphic interval between the chert overlying the Hanson Creek Formation and the Unnamed Dolomite of probable middle Lower Devonian age. In the central Great Basin, this interval is occupied in most places by either the Lone Mountain Dolomite or the Roberts Mountain Limestone, or both (Winterer and Murphy, 1960; Matti and others, 1974; Johnson, 1974, 1965; Berry and Boucot, 1970; many other workers).

Rocks up-section (unless there are large unrecognized structural displacements) from the chert zone in the vicinity of Big Cow Canyon may be equivalent to the Lone Mountain Dolomite, but they are entirely recrystallized dolomites, lacking fossils and primary structures. For these reasons, no formation name has been assigned to these rocks in this report. They are mapped as Silurian? dolomites.

On outcrop the very light gray weathering dolomite is fairly resistant, forming ridges up to 400 feet high. The dolomite is extensively disrupted by limonite-silica veinlets, which form brown-stained networks on weathered surfaces. Pervasive brecciation of these rocks contributes to obliteration of original bedding. Further destruction of primary structures probably was caused by recrystallization. Rare, indistinct surfaces which may have been bedding plane partings were observed about 1000 feet north of Summit 8479, $\frac{1}{2}$ mile west of the Big Cow Canyon road. Vugs 1 to 5 mm in diameter are very characteristic of this unit, and occur throughout its extent.

In thin section, the recrystallized nature of the unit is evident. Planar crystal boundaries predominate in the coarse crystalline dolomite mosaic. The only elements of undoubted detrital origin are a few (less than 1%) fine quartz grains. No fossils were seen in thin section.

UNNAMED DOLOMITE

General statement

A fine-grained, medium thick bedded, very light gray dolomite with a quartz-rich, cross-bedded upper unit crops out in several places in the map area. The lithology of this unfossiliferous unit is much like that of the Sevy Dolomite as described by Osmond (1954, 1962). However, more recent stratigraphic studies in Eureka County and vicinity (Kendall, 1975; Johnson, unpublished chart, 1974) show that the Sevy and the Kobeh Member of the McColley Canyon Formation (upper Lower Devonian) have the same regional unconformity as their basal contact. In other words, the lower Sevy and the Kobeh Member are lateral equivalents. The Kobeh Member is recognized with confidence in the map area on the basis of lithologic and fossil correlation. Large brachiopod collections from above the cross-bedded sandy unit have yielded faunas of fossil zones characteristic of the Kobeh Member (Murphy and Gronberg, 1970, p. 128).

Thus, a lithologically distinctive dolomite and quartzose dolomite unit occurs below the Kobeh in the map area. In other Great Basin localities this stratigraphic position is occupied by the Lone Mountain Dolomite (Kendall, 1975, Murphy and Gronberg, 1970, Nolan and others, 1956). The Lone Mountain-McColley Canyon contact is considered to be a disconformity or a slight angular unconformity at some localities.

The rock below the Kobeh in the map area does not resemble the Lone Mountain Dolomite. The Lone Mountain (at the type section at Lone Mountain and at most other localities) is medium to coarse-grained dolomite, massive to blocky weathering, has poorly defined bedding, and weathers light gray. Generally the formation forms resistant, cliffy exposures. There is no mention in the extensive literature on the Lone Mountain of a sand zone at its upper contact. Cross-bedded quartzose dolomites a few tens of feet thick in the basal McColley Canyon interval have been reported in the Sulphur Springs and Pinyon Ranges (Carlisle and others, 1957). A similar occurrence at the base of the equivalent Beacon Peak Dolomite near Eureka was described by Nolan and others (1956).

In the map area, the dolomite in the Lone Mountain position below the Kobeh is well bedded ($1\frac{1}{2}$ to 3 foot beds), very finely crystalline to finely crystalline, and resistant (2 to 3 foot benches), but not cliff-forming. Because it resembles the Sevy, but is below the Sevy interval as it is recognized in Eureka County, and because it does not resemble the Lone Mountain Dolomite, this dolomite is not assigned a formation name, and is referred to in this report as the Unnamed Dolomite. An identical unit has been mapped in nearby quadrangles by U. S. G. S. geologists (Ekren and others, 1973; Quinlivan and others, 1974).

Thickness

The lower contact of the Unnamed Dolomite is not exposed. Below the measured section on Devils Cave Ridge, the lower part of the unit is covered. Considering the top of the unit as the top of the cross-bedded sand interval, the minimum thickness exposed on Devils Cave Ridge is about 860 feet. Similar incomplete sections can be seen 2500 feet NNE of Summit 8782 in the central 1/9 of the map area.

Quartzose upper unit

The quartz-rich interval at the top of the Unnamed Dolomite was studied in detail. Figure 7 shows that the quartz-rich interval is distinct from the lithologies above and below. In the field, beds with more than a 5% content of quartz grains crop out as small cliffs. The quartz grains can easily be seen on weathered surfaces. Because of this, the cross-beds in this interval are easily seen. The cross-beds occur as planar sets 6 inches to 1 foot high and 1 to 4 feet in length along the dip slope. Apparent dips vary from 15 to 20 degrees; true dips are probably about 20 to 25 degrees.

Thin sections show that the quartz grains are rarely in grain contact. A mosaic of very fine grained recrystallized dolomite dominates the lithology. A sample from the 40-foot level in the Devils

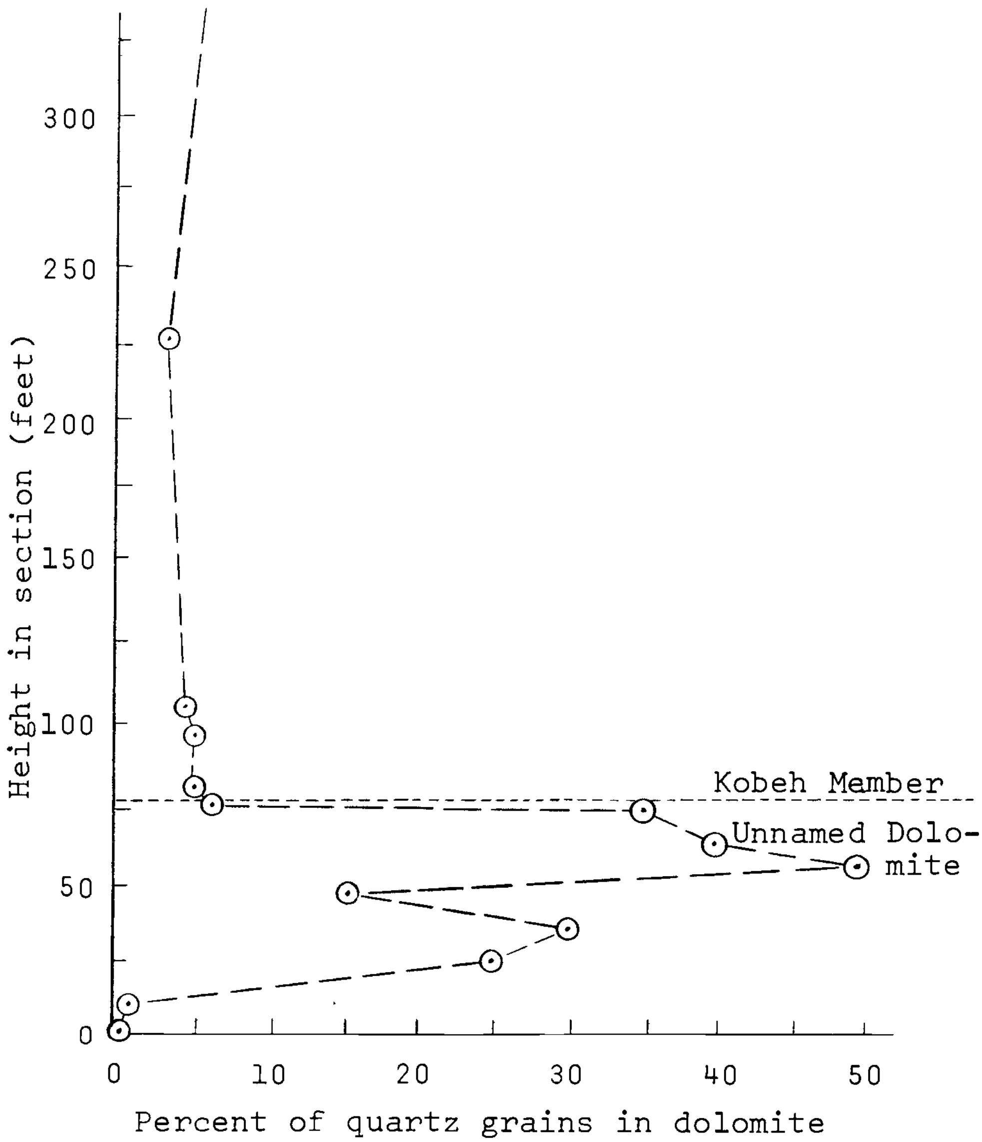

Figure 7. Percentage of quartz grains in dolomite, estimated from thin sections (data points), versus height in the Devils Cave Ridge measured section.

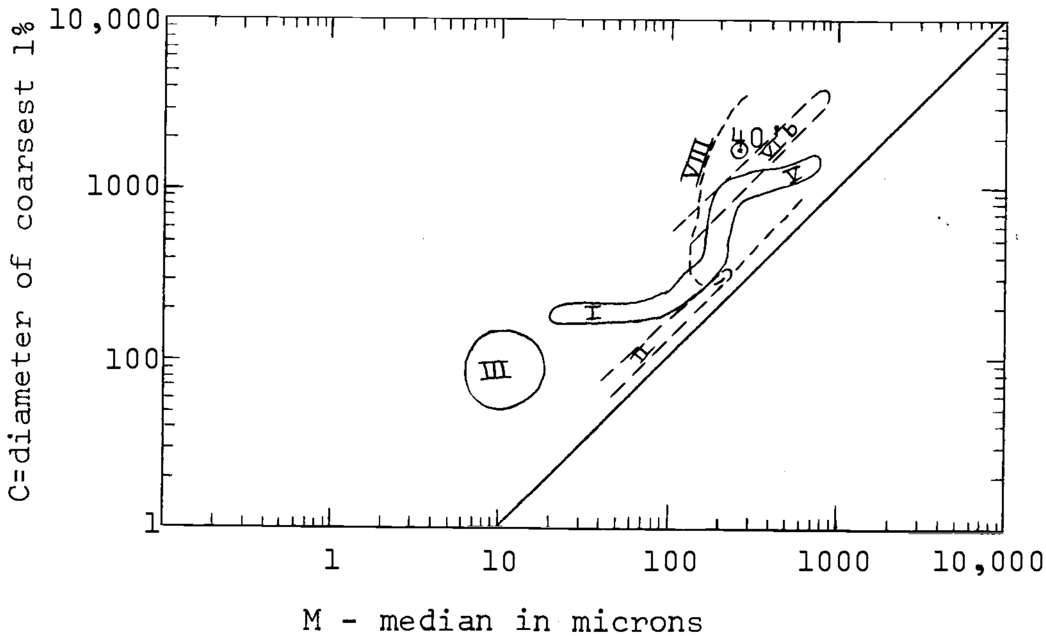

Figure 8. C-M diagram (after Passega, 1957), showing position of sample from the 40-foot level in the Devils Cave Ridge measured section. C= diameter of grain (in microns) such that 1% of the sample is coarser. M=median grain size.

I,IV, V: rivers, tractive currents
II, VIa,VIb: turbidity currents
III: quiet water deposits
VIII: beaches

Cave Ridge section was disaggregated in HCl, and a size analysis of the quartz grains was performed by standard sieve techniques. The quartz grains are sub-angular to well-rounded, moderately sorted (S_o=.76) (Folk and Ward, 1957), and the grain size distribution is nearly symmetrical (S_k=.022). Grains finer than 2ϕ (1/4 mm) are single particles, whereas coarser grains consist of quartz-cemented aggregates of up to 15 fine quartz grains. The clumped nature of these aggregates in thin sections indicates that the aggregates are fragments of a pre-existing quartz arenite. This interpretation is further suggested by the supermature heavy mineral suite (brown tourmaline and zircon) seen in thin section. Recycling of the sands of the Eureka Quartzite was suggested by Osmond (1954, 1962) as the source of a similar sand interval in the Sevy Dolomite. Such an interpretation seems equally valid for the quartz sand of the un-named dolomite. Johnson (1971) and Matti and others (1974) presented compelling arguments that the post-Lone Mountain, pre-Kobeh time interval was one of very low stand of sea level. The resulting erosion of lands to the east could plausibly rework exposed Eureka or Swan Peak Quartzite.

Depositional environment

Approximately 20% of the quartz grains from the 40-foot level in the Devils Cave Ridge measured section are of coarse sand size.

This probably rules out wind as the transport agent, because wind cannot move particles larger than $\frac{1}{2}$ mm (A. R. Niem, pers. comm.). The quartz grain-size parameters were plotted on a C-M diagram (Passega, 1957). This diagram delimits environments based on the median grain size and the size of grain with 1% of the sample coarser (see Figure 8). The sample from the 40 foot level (midway through the sandy unit) plots in area VII, which is typical of beaches.

The lack of fossils in the fine grained dolomite below the sandy unit suggests restricted or supratidal environments. The fine grained dolomite may have been primary dolomite forming in the supratidal environment in a manner analogous to that of the Persian Gulf sabkhas (Illing and others, 1965). Possible birdseye structures seen in thin section support this interpretation (Shinn, 1968) for the lower unit of the un-named dolomite.

More extensive grain-size analysis was not attempted on rocks of the quartz sand interval, because the insoluble portion constitutes only 45% or less of the whole rock. To interpret mechanical size analyses, we must assume that whatever carbonate particles were originally present were hydraulically equivalent to the quartz grains. If this assumption is made, then the cross-beds, the grain-size data, and the rounding of some grains suggest a beach or bar environment. This interpretation suggests a relative rise in sea level up-section, especially when the presence of normal marine faunas immediately above the sandy unit is considered. The progression of environments up-section may have been supratidal, intertidal (beach), and subtidal, based on the observations made above.

Regional perspective

J. G. Johnson (pers. comm., 1975) stated that the lower Kobeh Member of the McColley Canyon Formation in the map area is time-equivalent to the upper member of the Rabbit Hill Formation of the Monitor and Simpson Park Ranges. The basis for this correlation is the ?Pre-Trematospira fauna collected immediately above the Unnamed Dolomite (see Plates 2, 3). On the basis of stratigraphic position, the Unnamed Dolomite is probably equivalent to the lower Rabbit Hill Formation, and possibly the underlying Windmill Limestone in the Monitor Range. Both of these formations consist largely of thin-bedded laminated lime mudstones and wackestones with intercalated grain-flow and turbidite beds. Matti and others (1975) have interpreted the depositional environment as quiet-water basinal, with allochthonous carbonate sand sheets derived from shallow environments to the east. The Unnamed Dolomite of this report, plus the overlying Kobeh Member, were deposited in those shallow environments. The close proximity of basinal environments to the west represents an important eastward shift of the transitional facies, as noted by Johnson and Potter (1975) for the underlying Silurian rocks of central Nevada.

McCOLLEY CANYON FORMATION

General statement

Conformably overlying the un-named dolomite unit are Lower Devonian dolomites and limestones of the McColley Canyon Formation. These rocks form the lower part of a stratigraphic interval originally referred to by King (1878) as the "Nevada" and by Hague (1883) as the Nevada Limestone. Merriam (1940) restricted the use of the Nevada Formation to rocks above the Lone Mountain Dolomite and below the upper limit of the *Stringocephalus* Zone (Middle Devonian). Carlisle and others (1957) proposed a 3-fold subdivision of the Nevada Formation based on work in the Sulphur Springs and Pinyon Ranges. They recognized, in ascending order, the McColley Canyon Member, the Union Mountain Member, and the Telegraph Canyon Member. Johnson (1965) recognized a faunal and lithologic break at the top of the McColley Canyon beds in the Simpson Park Range and in the northern Roberts Mountains, and designated the McColley Canyon as a new formation. The former Nevada Formation became the Nevada Group and has been treated as such by many workers (Murphy and Gronberg, 1970; Matti and others, 1975; Kendall, 1975; Niebuhr, 1974; Flory, 1974). However, U. S. G. S. geologists continue to map the Nevada as a formation, and do not recognize mappable subdivisions. Commonly U. S. G. S. maps record the "Nevada

Formation" and recognize informal members in columnar sections which are not differentiated on the maps (for example: Quinlivan and others, 1974; Ekren and others, 1973; Merriam, 1973).

Murphy and Gronberg (1970) recognized three members (Kobeh, Bartine, and Coils Creek in ascending order) of the McColley Canyon Formation at Lone Mountain. On the basis of faunal correlation and lithologic similarity, these three members are also recognized in the map area. The McColley Canyon Formation is best exposed north of Summit 8782 (east of Big Cow Canyon). An incomplete section is exposed on the west flank of Devils Cave Ridge. Sections were measured at each of these localities (see Plates 1, 2, and 3), and large brachiopod collections were made at several heights in the sections.

Kobeh Member

Conformably above the quartz-sand zone of the un-named dolomite, medium-thick bedded, locally fossiliferous dolomite of the Kobeh Member crops out. These dolomite beds differ from those of the un-named dolomite in that the Kobeh beds are locally fetid, fossiliferous, well-laminated, and darker colored. Their quartz content is less than 5% (see Figure 7). Beds in the Kobeh Member vary in thickness from 6 inches to 4 feet. The unit is less resistant than the quartzose zone below, but nonetheless forms locally cliffy topography. Stylolites and chert lenses become more numerous near the top of the member. Bedding planes typically exhibit minor scour (3 inches to 1 foot relief). Color varies from light gray and yellowish gray to brownish gray on the weathered surface; fresh surfaces are medium dark gray, yellowish gray, and brownish gray. The alternation of light colored beds with darker, fetid beds in part of the Devils Cave Ridge section is noteworthy, and indicative of fluctuating depositional conditions.

In thin section, rocks of the Kobeh Member are seen to consist of 80% or more dolomite crystals forming an interlocking mosaic. Size of the crystals varies from .05 to .3 mm. The remainder of the rock consists of varying amounts of fine quartz grains, fossil fragments (brachiopod, crinoid, and coral debris) and dolomicrite pellets. The pellets are generally well-sorted and fine sand sized. Although some are ovoidal, most are irregular in shape. At 225 feet in the Devils Cave Ridge section they compose 30% of the rock. More commonly the pellets form 4 to 15% of the rock. It is not known if these bodies are fecal pellets or intraclasts, although fecal origin is suggested by the size and good sorting.

Fossils in the Kobeh Member are generally silicified or dolomitized and are concentrated in lenses, suggesting deposition or winnowing by currents. Brachiopod valves and crinoid fragments are dominant, although small horn corals are locally abundant, Brachiopod valves (Kobehana Zone and Costispirifer subzone) are commonly abundant enough to form grain-supported "shell-banks" a foot or more in thickness (see Figure 9). Although disarticulated, the individual valves are commonly intact, indicating that although currents were strong enough to concentrate and disarticulate the valves, depositional conditions were not sufficiently vigorous to break the shells.

The above evidence, plus the existence of very fine laminations parallel to bedding in many parts of the member, suggests deposition below wave base. However, deposition was probably relatively shallow, as evidenced by thick master bedding and presence of abundant shelly fauna including corals and large brachiopods. This interpretation is in accord with the relative position of the Acrospiriferid-Leptocoeliid Biofacies as described by Johnson (1974, p. 817). This position is a relatively shallow "inner belt across the marine benthic environment" (p. 817).

The Kobeh Member is 355 feet thick on Devils Cave Ridge and 480 feet thick in the Summit 8782 section. Profuse stylolites in the Kobeh of Devils Cave Ridge indicate that intrastratal solution may have decreased the thickness of this section significantly (see Figure 10).

Figure 9. Grain-supported silicified valves of brachiopods from the Costispirifer Subzone of the Kobeh Member of the McColley Canyon Formation. Sample is partially etched with hydrochloric acid. The rock is medium-fine grained dolomite with abundant small crinoid fragments. Dime gives scale.

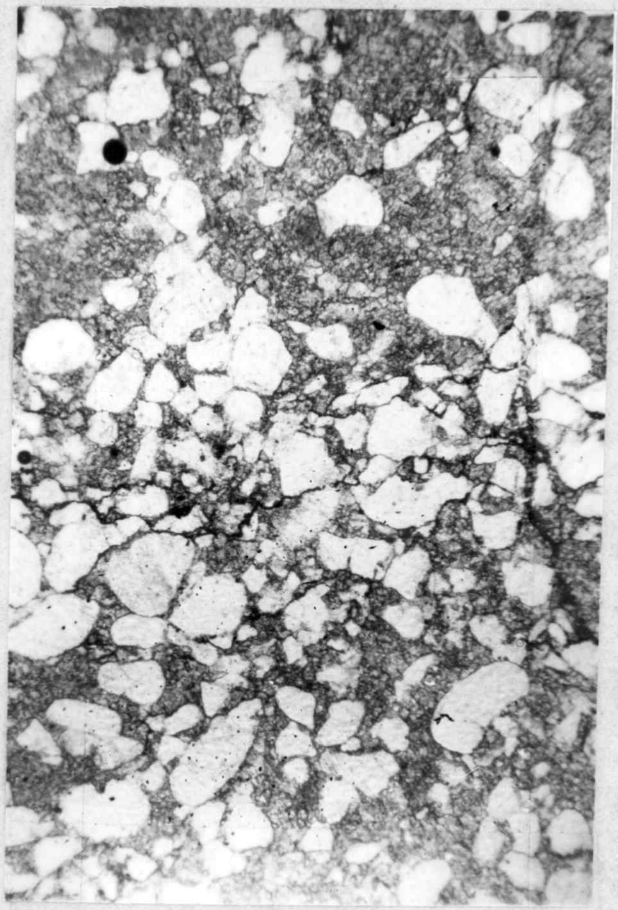

Figure 10. Photomicrograph of quartz grains clustered along a stylolite in the Kobeh Member. Quartz grains average .25 mm in diameter.

Bartine Member

Very fossiliferous beds of the Bartine Member conformably overlie the Kobeh Member dolomite. The transition from relatively barren dolomites to fossil-rich limestone occurs over a 2 to 3 foot interval.

The Bartine is a distinctive unit in the field. It weathers recessively, forming gentle slopes with few outcrops. Weathered colors vary from yellowish gray to olive gray and medium gray. Soils derived from the Bartine are a distinctive yellowish-tan. Beds of non-constant 2 to 12 inch thickness are commonly separated by argillaceous, flaky-weathering interbeds. The rock is rich in megafossils, including brachiopods, tabulate and horn corals, crinoid, byrozoan, trilobite, and nautiloid fragments, gastropods, pelecypods, and ostracodes. Preservation is generally by recrystallized calcite or molds. Brachiopods are typically articulated. Whole gastropods are common.

Most thicker beds are bioclastic packstones. The thinner beds are generally lime mudstones. Many beds are crudely graded, from bioclastic packstone upwards to mudstone.

In thin section, Bartine limestones are commonly rich in micrite pellets, which are well-sorted, generally ovoidal, and about .15 mm in long diameter. These pellets commonly make up 50% of both mudstone and packstone beds. Sparry calcite inside fossil chambers makes up 5% of some thin sections. This sparry filling is not geopetal, as it does not occupy the "tops" of different fossil chambers. Instead, it seems to occupy spaces sheltered from micrite infilling by currents.

The abundance of shelly fossils of several phyla, the irregular bedding contacts, and the presence of many corals collectively suggest a shallow marine environment. The intact fossils and the lack of winnowing suggest that deposition was below wave base and probably not in an area with constant currents. However, crudely graded beds and aligned long axes of fossils (see Figure 11) suggest the action of intermittent currents. Niebuhr (1974, p. 117) stated that the Bartine communities in Eureka County and vicinity were probably confined to Boucot's Benthic Assemblage 3 (Boucot, 1975), which spans depths in the range between 40 and 200 feet, J. G. Johnson stated that a wider range of depths is now known for the Bartine communities (pers. comm., 1975).

The Bartine Member is about 90 feet thick in the Summitt 8782 measured section. On Devils Cave Ridge, the section is incomplete, but at least 50 feet of Bartine is present. The _Eurekaspirifer pinyonensis_ Zone assignment, based on two brachiopod collections, is indicative of Emsian age.

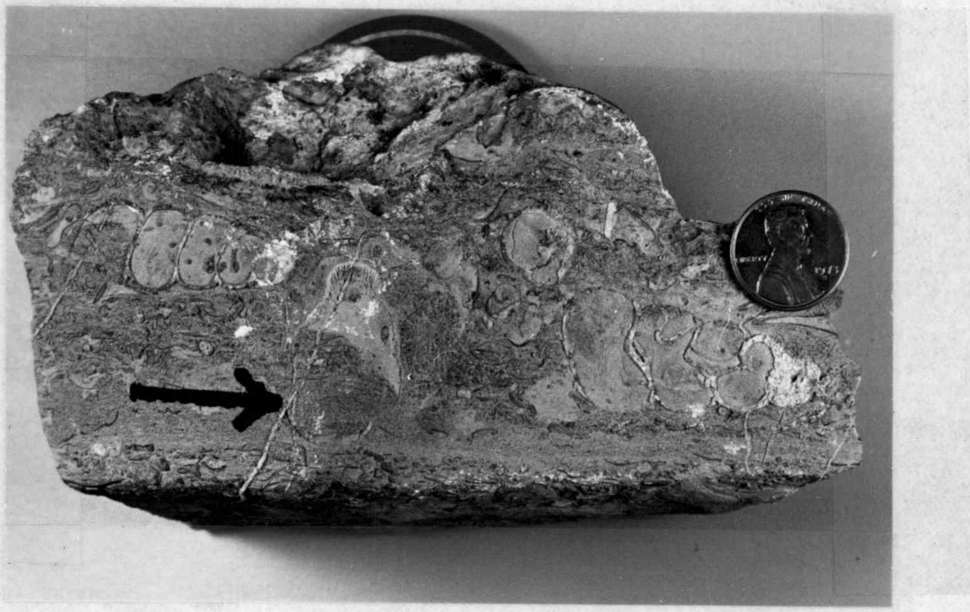

Figure 11. Photograph of elongate gastropods whose long axes are parallel to one another, indicating paleocurrents in the direction in the direction indicated by the arrow. Bartine Member of the McColley Canyon Formation. Penny gives scale.

Coils Creek Member

The yellowish-tan weathering limestones of the Bartine Member are conformably overlain by lime wackestone and packstone beds of the less recessive, light gray weathering Coils Creek Member. The transition upward to markedly less fossiliferous, more quartz-rich beds occurs over the distance of a few feet.

Coils Creek lithology is dominated by varying proportions of quartz and micrite, with locally very high concentrations of thin-shelled tentaculites. In addition, platy-weathering argillaceous interbeds separate the more massive, thicker beds (see Plate 3). On outcrop, the massive beds, which weather into blocks, vary in thickness from 2 inches to $1\frac{1}{2}$ feet. The platy-weathering interbeds are generally about half the thickness of the adjacent massive beds. Without magnification, the rocks of the Coils Creek Member appear to be uniform-textured lime mudstones. However, under a hand lens or in thin section, many of the beds are seen to be rich in very fine quartz grains (7-50% quartz grains). Sorting of the quartz grains ranges from moderate to good by visual estimate. The grains are sub-rounded to angular. In several thin sections, rocks judged in the field to be wackestones were seen to be quartz-tentaculite packstones (grain contact). Micrite is present in all samples, however, indicating that currents were not strong enough to cause winnowing.

Some current activity is evidenced by microlaminations of quartz grains alternating with tentaculite layers. Also, a few low-angle cross-laminations are present midway through the unit.

Tentaculites are present throughout the unit, composing up to 30% of the rock locally (see Figure 12). The tentaculites of the Summit 8782 measured section were studied by Eldridge (1974) (see Appendix). Abundance of tentaculites increases upward through the Coils Creek Member. The only other megafossils found were uncommon, very small brachiopods (collections EP-105, EP-105.5) and hundreds of siliceous sponge spicules from formic acid residues of sample EP-105. Elongate isotropic brown fragments seen in thin sections are pieces of inarticulate brachiopods (David Perry, pers. comm.).

Organic matter can be seen in thin section, and frequently is confined to the inside of tentaculite shells. When broken, rocks of the Coils Creek Member commonly emit a fetid odor, although this is not as pronounced as in the overlying Denay Limestone.

In the upper 50 feet of the Coils Creek Member, extensive burrowing in and through the bedding planes at low angles is common. The burrows are symmetrically ramifying cylindrical tubes 2 to 4 mm in diameter (see Figure 13). The tubes do not anastomose, and are of constant diameter. This burrow pattern is referred to Ichnogenus <u>Chondrites</u> (Hantzschel, 1962, p. 190). Simpson (1957, p. 475)

Figure 12. Negative print of an acetate peel showing grain-contact tentaculites from the upper Coils Creek Member. Black bar is 5 mm long.

Figure 13. Ichnogenus Chondrites: probable worm burrows, mostly in the bedding plane, from the upper Coils Creek Member of the McColley Canyon Formation. The burrows have piped fine sand into the micrite from the superjacent bed. Small dark pointed objects are thin-shelled tentaculites.

suggested that this type of pattern formed by a sipunculoid worm feeding "by means of its extensible proboscis." Seilacher (1964, p. 311) noted that this type of burrow pattern occurs over a large range of depths, the deepest of which is bathyal.

Much smaller burrows, .5 mm in diameter, and parallel to bedding, may be seen in thin section.

The extreme paucity of benthic fossils, the smallness of the few brachiopods present, the abundance of tentaculites, which are thought to be pelagic (Bouček, 1964, p. 37), and the relatively fine grain sizes present all suggest a relatively deep depositional environment - certainly deeper than that of the underlying Bartine Member with its shelly fauna.

The lack of a shelly benthic fauna cannot be attributed to a thixotropic substrate, because study of the walls of Chondrites burrows shows no evidence of collapse, which would suggest an unstable (thixotropic) substrate (Rhoads, 1970).

The alternation of massive and platy-weathering beds reflects variation in the influx of quartz grains. In the absence of any evidence indicating strong currents, it is concluded that intermittent gentle bottom currents or, possibly, near-bottom density currents, deposited the fine quartz sand in the massive beds. This would explain the general uniformity of texture (broken only by a few small cross-laminations) of the "massive" beds.

The upper contact of the Coils Creek Member was not described by Murphy (1973, unpubl. manuscript) indicates that in the Roberts Mountains, contrasting lithologies and fossil assemblages suggest a disconformity between the Coils Creek and the overlying Denay Limestone. This supports original observations made by Johnson (1962). The fauna of the Coils Creek in the Roberts Mountains, as reported by Murphy, includes several taxa (calcareous algae, favositids, *Alveolites*, horn corals) suggestive of relatively shallow water (photic zone). However, these taxa are not found in the Coils Creek of the Hot Creek Range; only tentaculites, a few conodonts, rare small brachiopods, and sponge spicules are present. Murphy described the lower 3 to 5 feet of the Denay as silty, medium dark gray weathering limestone with detrital quartz grains. Such an interval was found at 1070-1080 feet in the Summit 8782 section. The 1070 horizon is taken as the base of the Denay Formation, based on Murphy's lithologic criteria and on comparison of conodont zones with the conodont zones in the type section of the McColley Canyon Formation at Lone Mountain (Klapper and D. B. Johnson, 1975). Conodont collections at 1041 feet (29 feet below the contact) and 1320 feet (250 feet above the contact) in the Summit 8782 section yielded *Polygnathus laticostatus* and *P. costatus costatus* respectively. *P. laticostatus* occurs in the lower Coils Creek Member at Lone Mountain; *P. costatus costatus* occurs immediately above the Coils Creek as defined at Lone Mountain (Klapper and Johnson, 1975).

Thus, the upper contact of the Coils Creek in the map area is in about the same time-stratigraphic position as at Lone Mountain.

The Coils Creek Member on Summit 8782 is 495 feet thick.

Regional perspective

As previously noted, in early Kobeh time, an important facies boundary existed between shallow environments represented by dolomite (Hot Creek Range) and deeper environments where limestone was deposited (Monitor Range). During Bartine and Coils Creek time, limestone was deposited as far east as easternmost Eureka County (Kendall, 1975). To the south, the easternmost occurrence of McColley Canyon limestones is probably a few miles east of the map area. The McColley Canyon interval is occupied by dolomite in the Portuguese Mountain Quadrangle, 15 miles east of the map area (Quinlivan and others, 1975). The McColley Canyon Formation exists as far west as the Cortez area on the Eureka-Lander County boundary (D. B. Johnson, 1972). Farther west, time-equivalent rocks are probably siliceous clastics, although documented ages are few, because of the lack of fossils (J. G. Jonnson, pers. comm., 1975).

DENAY LIMESTONE

General statement

The Denay Limestone is best exposed in the complete Summit 8782 section, and on the west flank of the Summit 8782 ridge. There are two exposures in the northeast part of the map area, but these are incomplete sections largely disrupted by faulting.

The Denay was established by Johnson (1966) and defined as the limestone unit occupying the stratigraphic interval above the McColley Canyon Formation and below the Devils Gate Formation in the Roberts Mountains and the northern Simpson Park Range. This corresponds to the lower middle Nevada Formation (Merriam, 1940, 1963). Details of the lithology and upper and lower contacts have not been published, although Murphy (1973, unpublished manuscr.) has furnished these details based on field work in the Roberts Mountains and the Simpson Park Range.

Age

The lower contact of the Denay in the map area is just below an occurrence of *Polygnathus costatus costatus* (early Eifelian). The upper contact in other parts of Nevada may be as young as Late Devonian (Johnson, pers. comm., 1975). The age of the upper contact in the field area is not known. A brachiopod collection

(EP-107) from 20 feet below the upper contact was described by Jonnson (1974, written comm.) as "Middle Devonian or Frasnian." Middle Devonian is probably correct, as the dolomite unit above the Denay yielded brachiopod collections, one of which Johnson characterized as "probably about Castanea Zone." This collection (EP-900) was not from the measured section, but indicates that Castanea Zone time (mid Middle Devonian) is represented by dolomites rather than by limestones. Thus the uppermost Denay is older than Castanea Zone.

Lithology and outcrop characteristics

Murphy (1973) subdivided the Denay into three parts based on percentage of coarse-grained limestone beds. Such a subdivision is applicable in the map area, although the top of the formation is older in the Hot Creek Range than in the Roberts Mountains, where it extends above the Hippocastanea Zone (uppermost Givetian) (Murphy, unpublished measured section Willow Creek II).

The lower part (see Plate 2: 1070 feet to 1280 feet) consists of quartz-rich micrite, changing up-section into alternating blocky-weathering packstone and platy weathering wackestone beds. Thin-shelled tentaculites are locally very abundant, and occasional crinoid and brachiopod fragments are found. This part weathers recessively, and is best exposed along the ridge of Summit 8782. The lower contact, as described in the section on the upper contact of the

Coils Creek Member, is taken as the base of the finely cross-laminated quartz-sand rich horizon (see Figure 14) described by Murphy (1973).

The middle part (1280 feet to 1590 feet) consists of 310 feet of limestone characterized by abundant medium to thick graded beds of coarse bioclastic packstone. In the lower part of the unit, platy weathering argillaceous interbeds are present, but these disappear between 1325 feet and 1465 feet. All beds are tentaculite-rich in this part (Eldridge, 1974). Crinoid fragments, including common 2-hole ossicles which are indicative of the Emsian-Eifelian interval in Nevada (Johnson and Lane, 1969) are extremely abundant in the packstones (see Figure 15), and are often in grain contact. Other less common fossils are fragments of brachiopods, trilobites, and solitary horn corals and Hexagonaria. Fish scales and plates were found in the conodont residues of EP-105. 7, from the 1320 foot level. Above 1325 feet, micrite intraclasts up to 3 by 5 inches in size are common in the graded crinoidal packstone beds.

The upper part (1590 feet to 1865 feet) consists of markedly deformed beds of thinly laminated, pelletal lime mudstones and wackestones. The deformation is clearly due to penecontemporaneous slumping, as shown by the following field characteristics:

1. Heavily contorted beds commonly occur immediately adjacent to relatively undisturbed beds which have attitudes similar to

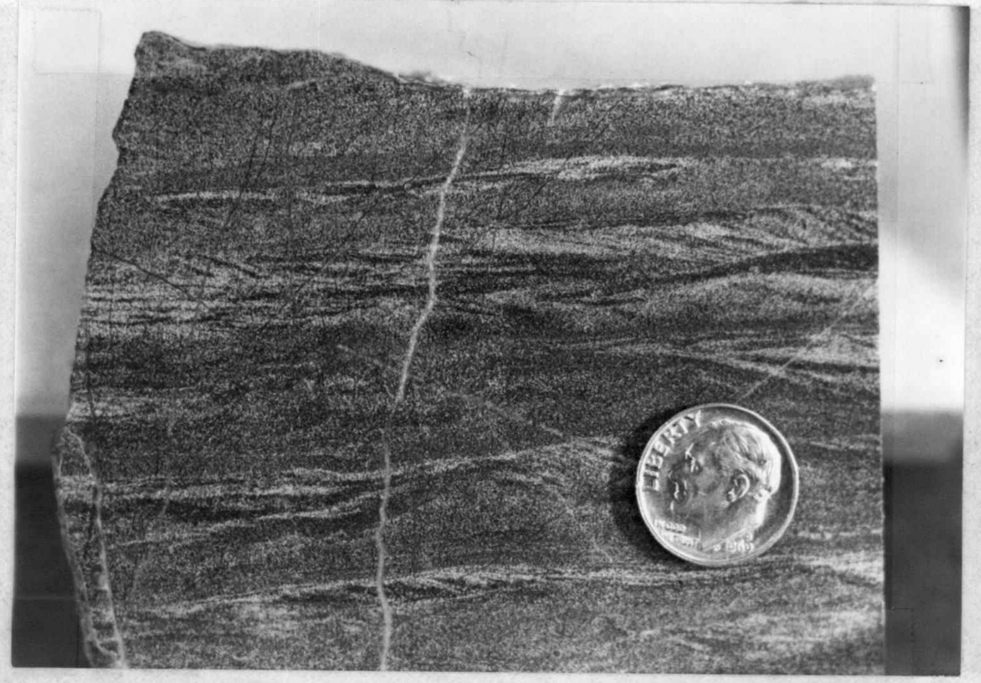

Figure 14. Basal Denay Limestone: quartzose, tentaculite-rich, pelletal neomorphic grainstone. Note the sets of very fine cross-laminations. The framework grains are very well sorted. Sample is from the 1070 foot level in the Summit 8782 measured section.

Figure 15. Crinoidal packstone from the middle part of the Denay Limestone. Note the abundant 2-hole crinoid ossicles.

those in the lower undisturbed part of the section.

2. Intricate contortion of thinly laminated beds exhibits no fracture lines, veins, or other indicators of brittle or tectonic deformation.

3. Existence of relatively large intraclasts within the beds immediately lower in the section indicates that there already was a significant depositional slope. The proximity of a slope was also necessary for the deposition of the graded, coarse bioclastic beds of the middle unit.

Upper contact

The upper contact of the Denay Limestone in the map area is one of abrupt transition to brownish-gray medium bedded dolomite. In the Roberts Mountains, where the top of the Denay is younger, the superadjacent formation is the Devils Gate (Murphy, 1973; Johnson, 1971, 1966). In the map area, however, the entire Denay is probably of Eifelian age.

Depositional environment

The lower part of the Denay is lithologically nearly identical to the underlying Coils Creek Member, and probably originated under the same conditions. The environment was probably a relatively quiet-water outer shelf or upper slope regime, at depths greater than those inhabited by most shelly benthic faunas. Intermittent gentle bottom currents probably deposited the fine-grained non-platy beds. The similarities between features illustrated in Figure 14 and features of contourites (Bouma, 1973) is striking. These features include: good to excellent sorting, small sets of fine cross-laminations, alignment of elongate microfossils (tentaculites and phosphatic brachiopod fragments in the Denay), and fine sand to silt-sized particles to the exclusion of any coarse material. Contourites are deposits created by reworking of bottom sediments by thermo-haline currents flowing parallel to bathymetric contours (Bouma, 1973). Such deposits commonly occur on continental slopes (Bouma, 1973).

The middle part of the Denay contains abundant evidence of deposition on or at the base of a slope. Fossils (including horn corals, tabulate corals, and Hexagonaria) indicative of relatively shallow water occur in coarse packstone beds which are crudely graded. The lack of Bouma sequences and the large size of some intraclasts suggests the movement of this material downslope by mass movements such as submarine debris flows. The lack of benthic fossils in the fissile interbeds suggests that the postulated debris flows came to rest in relatively deep water.

The upper part of the Denay, characterized by widespread penecontemporaneous deformation, must have been deposited on a slope, or a slope must have come into existence at the depositional site before lithification. Murphy (1973) has interpreted coarse bioclastic externally derived limestone beds ("allodapic beds") overlain by slumped beds in the Denay of the Roberts Mountains as indicative of slopes steepened by westward progradation of shallow-water sediments. In addition, Matti, Finney, and Murphy (1975) explained a similar sequence in the Lower Devonian Rabbit Hill Limestone of the Monitor Range by using the same model.

The model probably applies to the Denay of the Hot Creek Range as well. The Hot Creek Range and Roberts Mountains exposures of the Denay are the easternmost known. Thus, time-equivalent shallow-water beds must have existed within a few miles to the east. An extremely fossiliferous exposure dominated by crinoid fragments and encrusting stromatoporoids and dated as probable Eifelian on the basis of tabulate corals (see J-32 in Appendix) is located in the eastern part of the map area, 1 mile ENE of North Sixmile Canyon Spring. This Denay-equivalent exposure, which contains two-hole crinoid ossicles, horn corals, tabulates, and small brachiopods, may have been deposited in much shallower water than the Denay outcrops three miles west. It is plausible that this facies prograded over the Denay (the superjacent unit is a crinoid-rich dolomite of shallow-water origin - see the next section).

Regional perspective

As noted above, the easternmost occurrence ofthe Denay is along a north-south line immediately east of the Roberts Mountains and the thesis area. To the east of the line are time-equivalent dolomites of shallow-water origin (Johnson, 1971). To the west, the Denay probably has volcanic-clastic equivalents west of central Lander County (Murphy, 1973). Hence, the Denay occupies a north-trending belt about 25 to 35 miles wide in Eureka and Lander Counties. The Denay is probably present in the Monitor Range west of the study area, but no occurrences have yet been described.

MIDDLE DEVONIAN DOLOMITES

General statement

The stratigraphic position above the Denay Limestone in the Hot Creek Range is occupied by a medium to thick-bedded dolomite unit. Numerous normal faults cut the dolomite unit, and make the study of its original thickness, vertical stratigraphic characteristics, and degree of structural disruption very difficult. For these reasons, the unit has not been assigned to a particular formation, although it is probably equivalent to the upper Telegraph Canyon Member of the Nevada Formation in the Sulphur Springs Range.

Lithology

The rocks of this unit are all fine to medium-crystalline dolomites. They weather into 2 to 4 foot benches, form steep slopes, and commonly are exposed along ridge crests. There are two general modes of coloration. Dolomites of the lower part of the unit are grayish orange to pale yellow gray on weathered surfaces; fresh surface colors are brownish gray, light brownish gray, and medium dark gray. However, higher in the section, a distinctive alternation of lighter-weathering (very pale orange to light gray) and darker-weathering (medium gray to dark brownish gray) beds is common. Such alternating dolomite beds are common elsewhere in the Great

Basin in this part of the section. They occur in the Simonson Formation (Osmond, 1954) and its westward equivalents (Johnson, 1971, Figure 5), the Sentinel Mountain, Bay State, and Telegraph Canyon Members of the Nevada Formation near Eureka (Nolan and others, 1956, p. 44; Carlisle and others, 1957).

The lower part of the dolomite unit is characterized by an abundance of fine dolomitic crinoidal debris, which locally makes up 5 to 30% of the rock. Other fossils locally present (all silicified) include bryozoan fragments, horn corals, brachiopod valves, Thamnopora, other larger branching tabulate corals, and Syringopora. Occasional stromatolites are found. They are of the discrete vertically stacked hemispheroidal type, with variable basal radius (Logan and others, 1964) (See Figure 16).

Other important features of the lower part include faint cross-laminations in the dolomite beds, scour contacts between beds, and lensoid shape of some beds. The cross-laminations are locally abundant, and occur in 7 to 12-inch high sets in 2 to 3 foot thick master beds. Apparent inclinations of the cross-laminations (only one plane was visible in the outcrops) varies from 15° to 25°. Scour contacts, with relief varying from 3/4 to 12 inches, are common, and in some places cause beds to pinch out laterally within a horizontal distance of 10 to 20 feet. Faint to distinct laminations, spaced a few millimeters apart, parallel the bedding in most outcrops.

Figure 16. Club-like stromatolite head (type SH-V of Logan and others, 1964), from near the top of the lower part of the Middle Devonian dolomite unit. This stromatolite and several others in the same 1-foot bed were oriented perpendicular to the bedding.

The upper unit, characterized by alternating light and dark beds, is best exposed along the crest of Devils Cave Ridge east of North Sixmile Canyon Spring, and on east-facing slopes on the west side of North Sixmile Canyon. Light-colored beds are commonly gradational upward into dark beds, whereas the dark beds typically have sharp contacts with overlying light beds. The scoured nature of this contact is apparent in places where laminations of the dark beds are truncated by the contact. The relief on such contacts ranges up to 1 foot. Both light and dark beds vary from 1 to 10 feet in thickness. Dark and light beds may be of similar thicknesses locally, or the light beds may be much thicker. Near Summit 9612 on Devils Cave Ridge, the light beds are 10 to 12 feet thick, while dark beds have a thickness of only 3 feet. The darker beds emit a fetid odor when broken.

The darker beds are never cross-laminated, while the lighter beds occasionally show cross-laminations like those of the lower part of this dolomite unit. Stacked-hemispheroid type stromatolites occur sporadically, and only in the light beds. Dolomitized brachiopods and Thamnopora are occasionally found in the dark beds only.

In thin section, the darker beds of the upper part resemble the beds of the lower part of this unit. The dolomite crystals are .1 to .5 mm in diameter and have well-defined interlocking boundaries. The dolomite is somewhat clouded by tiny, disseminated flakes of organic material, which is present both at grain boundaries and inside dolomite crystals. Light-colored rocks are commonly coarser-grained, with patches of coarse euhedral dolomite crystals present in each thin section. Porosity is noticeable in both upper and lower parts of the unit, and ranges from 1% to 4% in thin sections. The porosity is generally in the form of irregularly shaped vugs 1 to 2 mm across.

Age, correlation, and thickness

The dolomite of this unit contains *Warrenella kirki* and *Stringocephalus* Zone brachiopods (Harry Dodge, pers. comm., 1975). Brachiopod collections EP-700 and EP-900 were obtained from the lower part of the unit. *W. kirki* and *Stringocephalus* Zone brachiopods establish the unit as late Middle Devonian in age. A complete section was not found in the map area. The lack of distinctive marker beds makes the estimation of sense and amount of movement along faults tenuous.

The problems introduced by very numerous normal faults make any estimate of the thickness of this unit tentative. Nonetheless, assuming relatively minor disruption of original stratigraphic relationships, a rough estimate of 1,700 feet has been made. This estimate is based on sections exposed south (up-section) from the top of the Summit 8782 measured section. Corroboration of this as a minimum estimate is afforded by the very thick but incomplete exposures on the slopes west of North Sixmile Canyon Spring. Carlisle and others (1957) noted that the Telegraph Canyon Member is more than 2000 feet thick at its type section in the Sulphur Springs Range.

A tentative correlation with the upper part of the Telegraph Canyon Member of the Nevada Formation is made here, on the basis of fossil ages and lithologies. The Telegraph Canyon Member was defined by Carlisle and others (1957) as equivalent to the Bay State, Woodpecker, and Sentinel Mountain Members of the Nevada (Nolan and others, 1956) in the Eureka district.

Depositional Environments

The association of horn corals, several tabulate genera, crinoids, occasional brachiopods, and cross-beds almost a foot high suggests a shallow, normal marine environment for the lower part of the dolomite unit. However, club-shaped stromatolites were also found (higher in the lower part), and these strongly suggest intertidal environments. The stromatolites are of type SH-V (discrete vertically stacked hemispheroids, variable basal radius - see Figure 16) (Logan and others, 1964). Type SH-V stromatolites are found today in intertidal areas at Shark Bay, Western Australia, "in headlands and locations where sea waves are moderate" (Logan and others,

p. 80). Several stromatolites of this type were found in beds in the upper part of the lower unit, at an elevation of about 8520 feet, 1.9 miles N40W from Summit 9638, west of North Sixmile Canyon Spring.

The environmental framework for the lower part of the unit appears to have been shallow subtidal conditions followed in time by possible intertidal conditions.

The upper part of the Middle Devonian dolomite unit is characterized by alternating dark, fetid, occasionally fossil-bearing beds and light, cross-bedded beds barren of fossils except for occasional stromatolites. The fluctuating environmental conditions which gave rise to the color alternation also caused scouring of the tops of dark beds. However, the fact that the lighter beds grade upwards into dark beds suggests that the light-bed environment actively shifted its boundaries, while the dark-bed environment was a frequent successor, gradually becoming dominant again after incursions of the light-bed environment.

The cross-laminations, stromatolites, and lack of both shelly fossils and fetid organic material suggest a high intertidal, and possibly occasional supratidal environment for the light beds. The dark beds, with occasional brachiopod and coral fossils (poorly preserved), high content of organic material, and lack of cross-laminations, probably were deposited in a protected subtidal environment. The very close spatial association with beds of probable intertidal or supratidal origin suggests that the environment of the dark beds was very shallow; the lack of evidence of strong currents suggests that this environment was lagoonal, and was located behind a barrier which was the site of the light-bed environment. Slight changes in sea level, or changes in current, wind, or wave regimes, or combinations of these variables, may have caused incursion of the barrier facies over the lagoonal facies, resulting in alternation of light and dark beds.

Regional perspective

The shallow-water features of this unit are typical of eastern assemblage rocks. The tentative correlation with the upper Telegraph Canyon Formation places this unit as the easternmost extension of very widespread, shallow-water Middle Devonian dolomites which extend eastward into central Utah (Osmond, 1954; Johnson, 1971).

DEVILS GATE FORMATION

General statement

Conformably overlying the light and dark alternating Middle Devonian dolomites is a thick, distinctive cliff-forming limestone unit, the Devils Gate Formation. The formation was established by Merriam (1940) as the upper division of the thick limestone unit referred to by Hague (1883) as the Nevada limestone. As originally defined by Merriam, the Devils Gate was composed of rocks between the top of the faunal zone containing *Stringocephalus* and the level of disappearance of the *Cyrtospirifer* ("*Spirifer disjunctus*") fauna. Nolan and others (1956) subsequently discovered that the Devils Gate, in the sense originally intended by Merriam, could be mapped on the basis of lithology alone.

The Devils Gate is widely recognized in the eastern Great Basin because it commonly forms cliffs and contains caves. In contrast to the older, recessive-weathering, thinner-bedded limestones generally confined to the transitional belt during the lower and middle Paleozoic, the commonly thick-bedded, resistant Devils Gate (together with the Guilmette Formation) extends from the longitude of the Hot Creek-Antelope-Fish Creek Range eastward to central Utah, over subjacent dolomites (Johnson, 1971, Fig. 5).

Rock description

The Devils Gate Formation in the map area is mostly thin to thick bedded micritic limestone which looks uniformly thick-bedded to massive from a distance. It occupies a roughly two square mile area east of North Sixmile Canyon, centered on Section 30, T. 10 N., R. 51 E. Upon close examination, the Devils Gate is seen to have some thin ($\frac{1}{2}$ to 1 inch) as well as thick ($1\frac{1}{2}$ to 12 foot) beds. Generally the thick and medium bedded portions are confined to the lower part of the formation, and weather into cliffs with occasional caves, whereas the thinner bedded portions occur in the upper parts of the formation, forming gentler backslopes with fewer caves. The caves throughout the formation are solution chambers large enough to walk into, extending inward 10 to 30 feet. Very extensive solution cave development occurs just east of Summit 8661 of Devils Cave Ridge. There, interconnected chambers with three north-facing portals honeycomb the ridge crest (see Figure 17). The portals are visible from elevated points miles away, and presumably are the "devils caves" for which the ridge is named.

The entire formation is heavily fractured and calcite-veined. Light olive gray is the dominant color, on both fresh and weathered surfaces. Weathered surfaces are covered with small (1 to 2 mm) solution pits, which impart a rough texture to outcrop surfaces. In

Figure 17. Caves in the Devils Gate Formation.

the northwest $\frac{1}{4}$ of Section 19, faintly cross-laminated quartz-rich beds of medium thickness were observed about midway through the formation.

A series of oolitic pelletal grainstone beds were found midway up in the lower cliffy part of the formation. These pellets and oolites are not visible in hand sample, but are seen in thin section to be 0.3 to 0.6 mm ovoidal to spheroidal micrite bodies, many of which have thin concentric oolitic coatings. Some of the pellets are too large to be classed as pellets as the term is commonly used (Blatt and others, 1972, p. 421), and may instead be rounded micrite intraclasts. Some of the oolites have ostracode valves as nuclei. Oolites compose about 20% of the rock in these beds. This oolitic interval may be the same one referred to by Nolan and others (1956, p. 50) as the interval separating their lower (Meister) member from their upper (Hayes Canyon) member. Such a subdivision is not warranted in the Hot Creek Range because the beds containing oolites cannot be recognized in the field.

Contacts and thickness

The lower contact of the Devils Gate Formation was observed in only one place - near the base of the Devils Cave Ridge cliffs in North Sixmile Canyon, $\frac{1}{4}$ mile west of the NW corner of Section 30. Here, the underlying medium-grained Middle Devonian dolomite is topographically recessive compared to the Devils Gate. The contact is transitional over about a 30 foot interval, with one-foot thick dolomite beds alternating with thinner limestone beds.

The nature of the upper contact in the map area is not known. Contacts with younger beds (Kinderhookian age shelly limestones) to the east are normal faults.

The generally uniform lithology of the formation makes interpretation of amount of disruption by faults difficult. However, thickness of the formation is at least 1000 feet, and may be as much as 2000 feet (Harry Dodge, pers. comm., 1975).

Age

No fossil collections were made in the Devils Gate, but this formation has been dated nearby along depositional strike in the Antelope Valley area (Merriam, 1963). Merriam (p. 54) states that the Devils Gate ranges from probable late Middle Devonian to late Devonian in age.

Depositional environment

Harry Dodge (pers. comm., 1975) reports finding "algal biscuits" in the Devils Gate in the map area. These are oncolites similar to those found elsewhere in the Devils Gate, and are indicative of very shallow environments. Logan and others (1964) and LaPorte

(1967) showed that oncolites probably form in very shallow subtidal to low intertidal areas. The very shallow setting suggested by the presence of oncolites is corroborated by the presence of oolites mentioned previously. Oolites today appear to form only in very shallow water - less than 12 feet in the Bahamas, where detailed studies have been done (Purdy, 1961; Newell and others, 1960).

Because oolitic and oncolite-bearing beds are not present throughout the Devils Gate Formation, it cannot be said that the entire formation was deposited at very shallow depths. However, at least parts of the Devils Gate are indicative of a very shallow subtidal depositional environment.

WOODRUFF FORMATION

General statement

A rock unit of unknown thickness, made up of mudstone, chert, porcellanite, and minor quartz arenite, is found at many places in the map area in thrust contact with the underlying Middle Devonian dolomites and the Devils Gate Formation. The unit has been dated (using conodonts) by the U. S. G. S. (Harry Dodge, 1975, personal communication) as early Famennian. A rock unit very similar in lithology, in part correlative and similarly in thrust-fault contact with underlying units has been described in the Carlin-Pinyon Range area by Smith and Ketner (1968, 1975). They named the unit the Woodruff Formation, and characterized it as predominantly siliceous mudstone and chert, with minor shale, siltstone, and dolomitic siltstone. The Woodruff in the type area weathers dark gray, black and tan. The mudstone is generally in beds less than ½ inch thick, and weathers on outcrop to tabular, non-fissile chips. The chert occurs in 1 to 4 inch beds, and in thin section is seen to contain argillaceous and organic material, spheres of radiolarian tests, and detrital quartz grains in amounts of less than 1%. Age of the Woodruff in the Carlin area ranges from Early Devonian to mid-Famennian, and is based on conodonts, goniatites, and graptolites. Other fossils found included tentaculites and abundant "claw-like rami of Angustidontus."

Because rocks in the thesis area are very similar lithologically, are time-equivalent to the upper Woodruff, and also occur in thrust contact with lower units, these rocks are tentatively assigned to the Woodruff Formation.

Rock description

There are two dominant lithologies in Woodruff Formation rocks in the map area - tightly to gently folded, brecciated, thin-bedded chert, and contorted, very thin-bedded dolomitic silty mudstones. The chert occurs beneath the mudstone, but frequently crops out because the mudstone is extremely friable and is easily eroded.

The resistant chert, which occurs in 1 to 3 inch beds, caps major topographic highs (Summits 9611, 9638, and two unlabelled summits west of North Sixmile Canyon). The chert is medium dark gray on fresh surfaces, and yellowish gray where weathered. Limbs of folds in the chert typically are 1 to 3 feet long, and commonly are overturned to the east. Excellent exposures of such structures can be seen on the minor summit $\frac{1}{4}$ mile NNW of Summit 8505 (west of North Sixmile Canyon). In thin section, the chert is micro-to-cryptocrystalline, contains abundant clay and organic material, and up to 1% detrital quartz sand grains. Needle-like outlines 1 to 3 mm in length and filled with microcrystalline quartz, suggest the former existence of sponge spicules in the chert. Much of the chert is brecciated, both on a large (1 inch clasts) and small (1 mm clasts) scale.

The mudstone portions of the Woodruff are never found in natural outcrops because of their extremely low resistance to weathering. Two artificial outcrop areas are afforded by shallow road cuts and prospect pits. These occur in the NE$\frac{1}{4}$ of Section 19, T. 10 N., R. 51 E., and one mile north of North Sixmile Canyon Spring. In these outcrops, the extreme contortion of the very thin mudstone beds is obvious. These beds are brown, red, black, and tan weathering. Fresh surfaces are impossible to obtain. The rock is friable, breaking very readily into 1 cm by 1 cm pieces, but generally is not fissile. It is composed of about 45% clay material, 10% very fine quartz grains, about 10% by weight oxidizable organic material, and traces of glauconite, muscovite, and siliceous sponge spicules. About 35% of the rock is fine crystalline dolomite cement. Dolomite rhombs less than 0. 2 mm long are common in thin section. Parallel microlaminations are the result of alternation of quartz-rich and clay-rich laminations. Elongate objects parallel to this microlamination include muscovite flakes, sponge spicules, and elongated quartz grains.

The poor resistance of the mudstone makes the mapping of its contacts difficult, due to mantling by talus. Another problem with this part of the Woodruff Formations is its resemblance to younger (Mississippian) autochthonous formations in this part of the Great Basin. However, the general association with folded and brecciated chert, along with rare outcrops showing contortion in the mudstone, suggests that it is allochthonous. In addition, it directly overlies Middle Devonian rocks with demonstrable thrust contacts, as discussed later in this report.

An interesting sidelight in the study of the mudstone unit is the local profusion of "eurypterid claw" fossils in the mudstone in Section 19, T. 10 N., R. 51 E. The writer did not observe these in the field, but Harry Dodge of the U. S. G. S. states that they are abundant, and have been identified by Jean Berdan as the "rami of an arthropod," probably Angustodontus. The same diagnosis was made of similar fossils from the type Woodruff Formation (Smith and Ketner, 1968, 1975).

Other lithologies in the Woodruff include grayish green weathering porcellanite (which occurs with the chert) and occasional beds of poorly sorted quartz arenite. The quartz arenite is chert-cemented. The framework grains consist of medium to very fine grained quartz sand. The coarser grains are well-rounded. This lithology occurs associated with the mudstones, but its exact relationship to the mudstone beds could not be determined because of poor exposures of the mudstone.

Thickness and contacts

Because of the structural disruption of the Woodruff Formation, no reliable determination of original thickness can be made. Present (deformed) thickness ranges up to approximately 350 feet, of which about 50 feet is chert, and 300 feet mudstone. Thickness is probably underestimated because the extremely friable mudstone is very easily eroded.

The lower contact is rarely seen, but where visible it is a sharp thrust surface between chert and dolomite. Rocks above and below the contact are brecciated and the dolomite breccias commonly are veined with silica networks. The Woodruff appears to overlie parts of the Devils Gate Limestone, but thrust contact with the Devils Gate have not been observed.

Previous interpretations

The unit here called the Woodruff Formation was mapped by Kleinhampl and Ziony (1967) as Cockalorum Wash Formation. This arthropod rami-bearing unit was defined by Merriam (1973) 30 miles north of the map area. It is Middle Devonian in age, however, and thus cannot be correlated with the Late Devonian strata of the map area.

Harry Dodge (pers. comm.) believes the unit in the map area to be a parautocthonous equivalent of the Pilot Shale. The time-correlation with the Pilot is valid, based on conodont dates. However, I believe that structural evidence and stratigraphic position of the unit necessitate the interpretation of emplacement as a thrust sheet from the west. This hypothesis will be further discussed in a later section on the structure of the map area.

MISSISSIPPIAN ROCKS

General statement

Rocks of Mississippian age crop out in two faulted and fault-bounded exposures in the map area - the NW corner of the NW 1/9, and the E $\frac{1}{2}$ of the EC 1/9. For reasons discussed bleow, these rocks are not assigned to formations, although probable equivalents elsewhere are cited. The rocks are described from oldest to youngest.

Kinderhookian-Osagian

An extensive exposure of pale brown to grayish yellow weathering limestone, commonly in graded beds, occurs in the NW 1/9 of the map area. The exposed section is 300 to 400 feet thick. This limestone, which is pale to dark yellowish brown on fresh surfaces, contains abundant evidence of deposition by currents, some of which were probably turbidity currents. The evidence includes scoured basal contacts (5 inches of relief), $\frac{1}{2}$ to 2 foot thick graded beds, some of which contain 2 to 4 inch intraclasts, and mixtures of conodonts (lower Famennian through Osagian-Kinderhook boundary) from a single 8 inch thick bed.

The graded beds (see Figure 18) generally contain Bouma divisions A-B-E (Bouma, 1962). The graded (A of Bouma sequence)

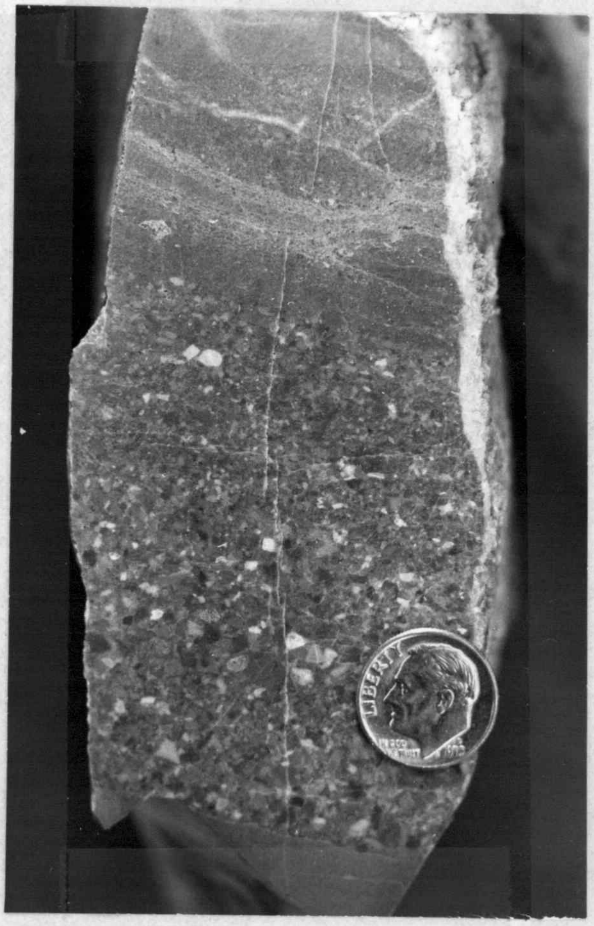

Figure 18. Graded bed composed of mudstone clasts, pelletal aggregates, crinoid fragments, forams, and other bioclastic fragments. The bed was probably deposited by a turbidity current, and contains Bouma divisions A, B, and E. C and D divisions may also be present. Division E of the lower (preceding) bed is visible at the bottom. Sample is from the Kinderhook-Osage exposure in the NW corner of the map area.

portions of beds typically contain lime mudstone intraclasts, planispiral forams(?) crinoid, ostracode, brachiopod, and echinoid spine fragments, and quartz sand grains. Some of the mudstone clasts are pelletal aggregates. The brachiopod fragments appear to be coated grains, which suggests possible algal encrustation before reworking. Size of the constituents of the graded portions of beds ranges from 5 mm to .5 mm.

The plane-parallel laminations (B of Bouma sequence) of the beds consist of pellets, ostracode valves, forams, and fine crinoidal debris. These constituents average 0.5 mm in diameter and appear to be well-sorted. Both A and B divisions are cemented by microspar (probably recrystallized micrite).

Division E of the Bouma sequence was not observed in thin section. It is a fissile, medium gray weathering, argillaceous silty limestone. Many D-E sequences (parallel laminations overlain by interturbidite "shale") are present throughout the section, suggesting that distal turbidites are interbedded with more proximal turbidites. Such occurrences are best explained as having been deposited by shifting lobes of a submarine fan.

The very fossiliferous nature of these beds becomes apparent only in peel or thin section study. Most of the bioclastic material, except for a few crinoid fragments, is invisible in hand specimens.

The age of these beds was determined from a conodont collection (EP-400) taken about midway up the exposed section. Conodonts from an 8 inch bed range in age from lower Famennian to the Osage-Kinderhook boundary (Klapper, written communication, 1975). Considering sedimentological evidence, it seems likely that the older conodonts are in the intraclasts, and that the time of deposition coincides with that of the youngest conodont elements (earliest Osage - latest Kinderhook time).

A similar limestone turbidite unit of Kinderhook age was described by Ketner (1979) from Elko County, Nevada. Meager evidence from bedding surface features suggested transport from the east, possibly by turbidity currents triggered by "early pulses of the Antler orogenic episode" (Ketner, 1970, p. 20). No transport direction evidence was obtained from the rocks in the map area, but an isopach map of the Great Basin Lower Mississippian (Poole, 1974, p. 75) shows a very narrow, deep, north-northeast trending basin passing through or very near the map area. Thus the postulated turbidity currents may have come from the west or the east, or may have flowed along the basin axis. The western source for sediments, beginning in latest Devonian time at about this longitude in Nevada, was the rising Antler Orogenic Highland (Brew, 1971; Johnson, 1971).

This exposure is bounded by faults and by Tertiary and Quaternary cover, except for some vari-colored shales in small exposures to the northwest. These are mapped as Mississippian rocks, and may be Pilot Shale equivalents. Because of the faults bounding the Mississippian rocks to the east and south, relationships with the underlying rocks are unknown.

Osagian-Upper Tournaisian

Exposures in the east half of the EC 1/9 of the map area contain rocks of Osagian-Upper Tournaisian age. The age determination was made by Carter (written communication, 1974) on the basis of a brachiopod collection (EP-800) from 1000 feet SSE of the NE corner, Section 20, T. 10 N., R. 51 E. Rocks in these outcrops contain evidence of high-energy current or wave conditions at the site of deposition. The rocks are coarse crinoidal bioclastic grainstones, and contain a significant proportion of rounded radiolarian chert pebbles (see Figure 19). The beds are 6 to 18 inches thick and are occasionally crudely graded, with chert, flat limestone intraclasts, and large biotics making up the coarser fraction.

Biotic constituents include fragments of crinoids, large brachiopods, horn corals, tabulate corals, brachiopod and echinoid spines, bryozoa, and foraminifera. Other framework constituents include small rounded pebbles of chert, and rare quartz arenite and volcanic rock fragments. Less than 10% of the rock is sparry calcite pore-filling cement. Some samples are cemented by chert and chalcedony,

Figure 19. Coarse crinoidal bioclastic grainstone containing rounded pebbles of radiolarian chert. Osagian-Upper Tournaisian brachiopods were found at the same location.

apparently replacing sparry calcite.

In thin section, most of the chert pebbles, which range from 2 mm to 5 cm in diameter, are seen to contain ghosts of radiolarian spheres and spines (see Figure 20).

The area of the exposure described above is considerably faulted, and as a result, this lithologic type has fault contacts, and the character of its base is not known. However, observations to the south a few hundred yards suggest that a shale-siltstone-sandstone unit overlies the crinoidal bioclastic limestone of Osagian-Upper Tournaisian age.

The age of the limestone suggests correlation with the Joana Limestone (Poole, 1974). The Joana is a carbonate shelf unit deposited between the subsiding Antler Foreland foreland trough to the west and the stable craton to the east, according to Poole's reconstruction (1974).

Rounding of the chert pebbles must have occurred in fluvial or littoral environments, and supports other evidence (grain-supported fabric, lack of micrite, big-shelled brachiopods, flat intraclasts, thick beds) indicating very near-shore, shallow, high-energy original environment of deposition. Crudely graded beds suggest re-working, perhaps by submarine debris flows. No current-direction features were seen on outcrop, but it is plausible that the chert pebbles were derived from former deep-water cherts uplifted in the Antler

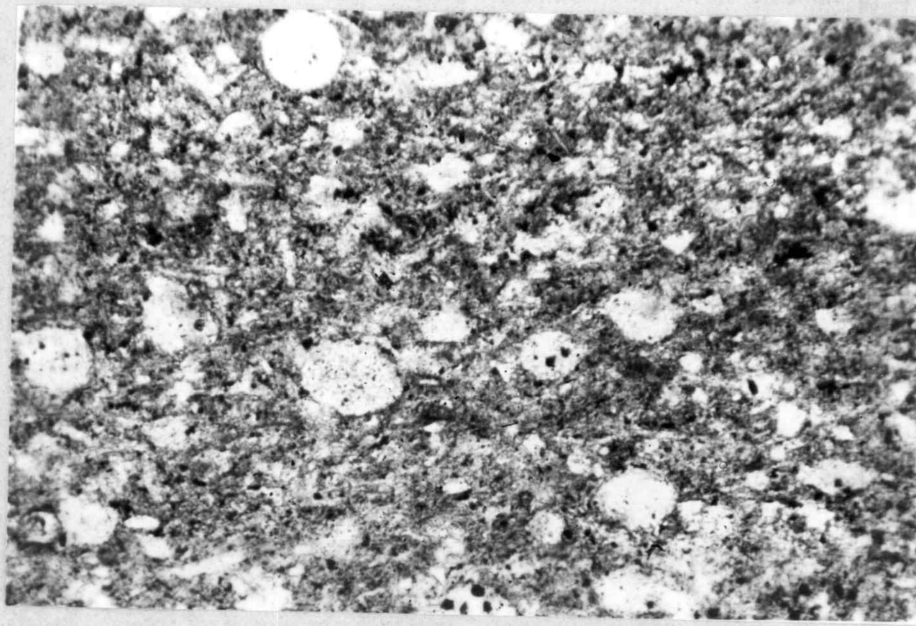

Figure 20. Photomicrograph (X 60) of radiolarian chert from sample pictured in Figure 19. The rounded, light-colored spots are recrystallized radiolaria, which average .15 mm in diameter.

orogenic highland to the west. The Joana Limestone in Eureka County contains chert pebbles from that source (Brew, 1971), and no eastern source is known for radiolarian cherts.

Visean

A distinctively fossiliferous medium gray weathering limestone forms a low knoll (Summit 7399) at the extreme east side of the map area. Brachiopods collected from this limestone (about halfway up the 200 to 300 foot section) are of Visean age (Carter, written communication, 1975; brachiopod collection EP-500). The knoll is fault-bounded, and its stratigraphic relationship to the shale, mudstone, and siltstone unit to the west is unclear. The Visean limestone occurs in 8 inch to 3 foot beds, and is very rich in large brachiopods, crinoid columnals (some up to 7/8 inch in diameter) and bryozoa (see Figure 21). It contains lesser amounts of forams, horn corals, echinoid spines, and pellets. Roughly 30% of the rock is sparry calcite, as pore-filling cement and overgrowths on crinoid fragments. Small chert pebbles are present but rare.

Correlation of this limestone unit with similar units of the same age in the Great Basin is difficult. The lack of Upper Mississippian limestones is apparent on correlation charts such as that of Poole (1974, Fig. 2). The only known occurrences of shelly limestones of this age in Nevada are thin limestone intervals in the Eleana

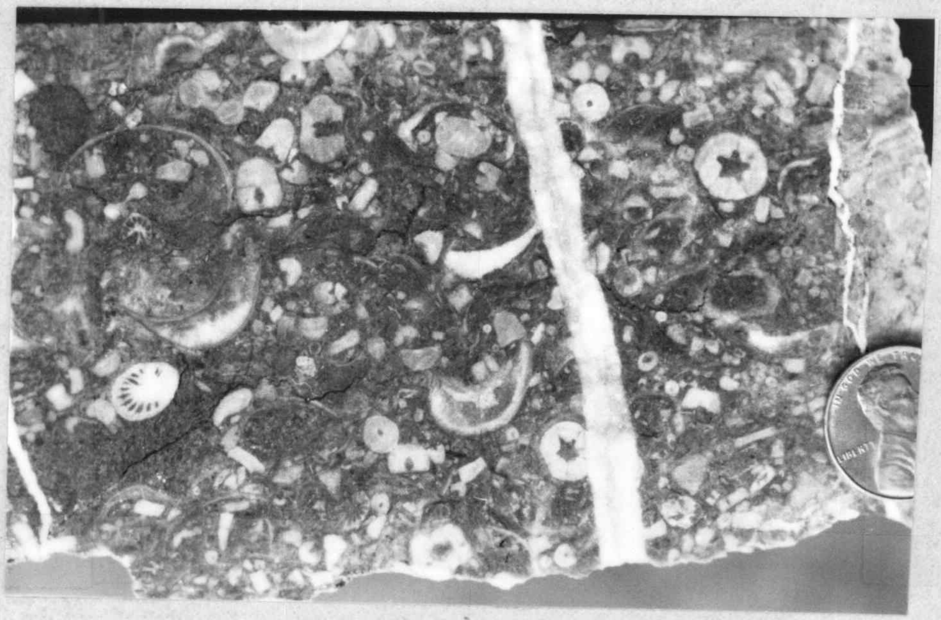

Figure 21. Bioclastic grainstone of Visean age, from fossil locality EP-500. Crinoid fragments, articulated and disarticulated brachiopods, horn corals, bryozoa, and forams make up most of the shelly component. In thin section, pellets, small chert granules, and echinoid spines can be seen.

Formation (Poole and others, 1961) of the Nevada Test Site vicinity north of Las Vegas.

The Visean limestone must have accumulated in shallow normal marine waters, judging by its rich shelly fauna. The fairly abundant pellets suggest a fairly quiet environment (Newell and others, 1960), which means the depositional environment of this limestone was probably below wave base.

The depositional environment of the limestone was certainly different from that of the adjacent mudstone, siltstone, and shale unit. Exposures of these clastic rocks, in which no fossils were found, cover the gently east-sloping area in the vicinity of the boundary between Sections 20 and 29, T. 10 N., R. 51 E. These rocks are lithologically equivalent to the Chainman Shale (see Brew, 1971 for description of the Chainman) and are mapped as Mississippian rocks. Gray, black, and yellowish brown fissile mudstone and friable siltstones dominate the lithology of this clastic unit. Occasional 8 to 12 inch thick graded limestone beds are interbedded with the mudstones. Sizes of the clasts in the graded beds range from 1 cm to silt-sized. Uncommon sandstone beds are also present, interbedded with the mudstones. In thin section, these sandstones are seen to be arkosic arenites with a significant component of altered volcanic rock fragments and detrital chert. Sorting is fair, and the grains are angular to well-rounded. Such lithologies are typical of the clastic rocks shed eastward from the Antler orogenic highland during Late Mississippian time in Eureka County and vicinity (Brew, 1971; Poole, 1974).

TRIASSIC ROCKS

Rock Description

Rocks firmly dated on the basis of conodonts as Griesbachian (early Early Triassic) crop out in the southwest corner of the north-central 1/9 of the map area. Mesozoic rocks are beyond the scope of this study, but lower Triassic marine rocks are a relative rarity, and fossils in the Griesbachian rocks in the map area will be of interest to those studying the massive Permo-Triassic extinctions. Therefore a short description of the Griesbachian rocks is included.

Most of the small exposure consists of fissile tan, yellow, and red calcareous, clay-rich siltstones. Interbedded with the fissile rocks are thicker beds of bioclastic packstones and calcite-cemented lithic arenites.

The packstones occur in rare 6 to 8 inch beds. One of these was sampled for conodonts (EP-200 - see Appendix). This bed contains very abundant shell debris, in addition to very numerous conodonts. Less than 1% of the bed consists of fine-grained quartz and chert. Slabs cut parallel to the bedding reveal very abundant whole ammonoids each about ½ to 1 inch in diameter. A concentration of about 15 of these shells per square foot of bedding plane is typical. Very numerous flat-based, coiled silicified tubes 2 to 4 mm long, resembling tiny gastropods, were seen on outcrop and in formic acid residues. The tubes were identified as Spirorbis by Ellis Yochelson (letter to J. G. Johnson dated May 30, 1975). Spirorbis is a serpulid worm with extremely cosmopolitan and long time-range distribution. In thin sections and peels, Spirorbis tubes are seen attached to thin pelecypod valves.

The lithic arenite beds commonly are 1 to 2 inches thick, and weather grayish-red to grayish-orange. In thin section, about 20% of the rock is seen to be altered rock fragments, which may have been volcanic. The rest of the rock is typically 15% angular medium grained quartz, 5% rounded medium grained chert fragments, 5% plagioclase, and less than 1% muscovite and sponge spicules. About 50% of the rock is calcite cement, with minor limonite and occasional small dolomite rhombs.

The Triassic exposure is areally very limted (about 1/8 mi^2). It dips gently northeast, and rests unconformably on rocks of probable Silurian age. The homoclinal attitude and undisturbed nature of the bedding suggest that these rocks were not part of a thrust plate. Time-correlative rocks in western Nevada were involved in eastward thrusting during Early Triassic time (Silberling, 1973).

Correlation

Rocks of Griesbachian age are rare in the Great Basin. Most studies show the Lower Triassic as a hiatus. Parts of the Moenkopi

Formation of the Colorado Plateau are time-correlative with the Hot Creek Range rocks. However, most of the Moenkopi is non-marine; the marine portions in southern Nevada and Utah have not yielded conodonts (H. J. Bissell, written comm., 1975). Time-correlative rocks west of the Moenkopi exposures are largely marine limestones or volcanics (Silberling and Roberts, 1962), although known exposures are very few. These exposures include the Tobin Formation which crops out in the New Pass and Tobin Ranges south of Winnemucca. The Tobin is a fine-grained, calcareous, terrigenous clastic rock with limestone interbeds containing "marine molluscs" (Silberling and Roberts (1962).

The Koipato Group, a "largely non-marine rhyolite and andesitic volcanic and clastic succession" (Silberling, 1973) is at least partly Early Triassic in age (p. 352, Fig. 3). The Koipato is exposed in the Winnemucca area.

Closer to the study area, the Candelaria Formation of Upper Permian-Lower Triassic age is exposed in the Toquima Range and near Tonopah. "Shallow water marine calcareous and fine-grained clastic rocks make up most of the section" according to Silberling (1973, p. 355), who hypothesized that the axis of an Early Triassic depositional basin trended NNE and passed about 50 miles east of the map area (p. 347, Fig. 1).

The Triassic limestones in the map area are marine, as evidenced by conodonts and ammonoids. Clastic-dominated deposition of the bulk of the Hot Creek Range Triassic rocks may indicate relative nearness of source areas, which would be to the west, according to Silberling's (1973, Fig. 1) reconstruction.

The Triassic exposure in the map area is probably equivalent to the Candelaria Formation of the southern Toquima Range, on the basis of lithologic similarities and time-equivalence.

Other lithologically similar exposures occurring beneath the volcanics in the northern part of the map area are tentatively mapped as Triassic. No fossil age determinations were made of these exposures.

STRUCTURE

Structure in the map area is typical of the structure at other localities at this longitude in the Great Basin. Post mid-Famennian thrust faulting brought fine-grained detrital rocks eastward over autochthounous carbonate rocks. Large-scale Tertiary normal faulting resulted in typical Basin and Range fault-block topography. In south-central Nevada, the ranges trend north or north-northeast, and summit ridges are generally 2500 to 5000 feet above adjacent valley floors.

Thrusting

The Woodruff Formation is shown on the map (Plate 1) as an allochthonous unit overlying the upper Middle Devonian dolomite and tentatively overlying the Devils Gate Limestone. Evidence that the Woodruff is allochthonous is as follows:

1. At two locations the base of the Woodruff is seen to overlie a shear zone characterized by silicified networks in the underlying dolomites. Both the Woodruff chert and the underlying dolomite are fractured and brecciated. These locations are just east of the top of Summit 9638, and $\frac{1}{4}$ mile SW of Summit 8200. The contact surface is locally sharp and appears to be roughly parallel to underlying bedding in a given outcrop.

2. The Woodruff rocks are extensively deformed - the chert into gentle, open folds and tighter, overturned folds; the mudstone into chaotic sheared masses of diverse orientations.
3. The Woodruff rocks are of an entirely different facies (siliceous clastics) than autochthonous carbonate rocks which bracket them stratigraphically.
4. Regional relations indicate that widespread eastward-directed thrusting occurred in latest Devonian-earliest Mississippian time in central Nevada (Johnson, 1975). These thrusts, of which the Roberts Mountains thrust is best known, brought fine-grained clastic and cherty rocks eastward tens of miles over the transitional and eastern facies carbonates (Smith and Ketner, 1968, 1975; Johnson, 1971, 1975).

Axial planes of folds in the Woodruff chert strike northeast and (for the overturned folds) dip west. Thus, compression acting along a northwest-southeast trend was responsible for the folding (and probably the thrusting), and an active vector from the northwest is likely. Such a regime is compatible with Roberts Mountains-type thrusting which occurred at about the Devonian-Mississippian boundary in the Carlin-Pinyon Range area (Smith and Ketner, 1968).

The thrusting in the map area may have been earliest Mississippian because the allochthonous rocks are Famennian - no younger rocks are included in the thrust plate. However, no younger age limit can be placed on the thrusting in the map area, because the Woodruff is not overlain by Paleozoic rocks in the area.

Relation of the Woodruff rocks to the Devils Gate Limestone in the map area is unresolved. Tentatively mapped as a thrust surface in some areas, the contact is nowhere well-exposed. If the allochthonous Woodruff Formation rocks are in thrust contact with the Devils Gate, then thrusting either occurred at widely separated times, or the thrust must have been at least partly an underthrust, with "slivering" of the allochthonous plate accounting for emplacement at two stratigraphic horizons.

Normal faulting

Hundreds of high-angle normal faults cut the entire stratigraphic sequence in the map area. Trends and displacements along the faults vary, but may be summarized as follows. Long, north to northeast trending faults of large throw broke the range into elongate blocks which are now high ridges. Later adjustment within these large blocks caused normal faulting along many trends, the main one being roughly perpendicular to the master faults.

The major north and northeast-striking faults caused the vertical displacement of the range relative to the basins on either side. The northeast-striking fault bounding the Antelope Valley Limestone on its west side has a probable vertical throw of 5200 feet, as indicated by the juxtaposition of Circula Zone fossils (Harry Dodge, pers. comm., 1975) of the Denay Limestone and the middle part of the Antelope Valley Limestone. The north-striking fault in the vicinity of North Sixmile Canyon Spring juxtaposes Lower Devonian or older Unnamed Dolomite and Middle Devonian dolomites, and thus has a throw of about 2000 feet. The apparent greater vertical displacement along such faults in the western side of the range results in the oldest rocks being exposed there, with generally younger rocks exposed towards the eastern side of the range.

The shorter intrablock faults caused minor (a few inches to tens of feet) vertical displacements. These faults are best documented where stratigraphy could be determined most precisely. Therefore, the faults depicted in the vicinity of both measured sections is probably representative of the structure in the whole area, but only in these limited areas could offsets of only a few feet be recognized.

It can be seen in the field that some of the major faults cut the Tertiary volcanics, which are Oligocene to Pliocene in age (Ekren and others, 1973). Therefore some faulting is post-Pliocene in age, but it cannot be demonstrated that all the normal faults are so young. However, work on the volcanic stratigraphy in the adjacent Moores Station Quadrangle (Ekren and others, 1973) shows that the great majority of normal faults are post-volcanic in that area.

CONCLUSIONS

Lithologies and fossils of the rocks studied indicate that shallow shelf to intertidal and possible supratidal conditions existed in the map area in Early Ordovician through Early Devonian time. Early and Middle Ordovician depositional regimes appear to have shallowed upward through the Antelope Valley Limestone, ending with possible supratidal conditions in the overlying Eureka Quartzite. Late Ordovician time was probably one of deepening of marine waters, as indicated by the transitional contact between the Eureka and the overlying dolomite of the Hanson Creek Formation. The dolomite of possible Silurian age lacks primary features and cannot be used as an environmental indicator. Next higher in the section is the Unnamed Dolomite which probably was deposited in the intertidal or supratidal zone. Progressively deeper shelf conditions are suggested by fossils and lithologies in the overlying Lower Devonian Kobeh and Bartine Members of the McColley Canyon Formation.

In a regional context, the Ordovician units have lithologic equivalents well to the west of the map area. However, the Unnamed Dolomite and the Kobeh Member of the McColley Canyon Formation are the westernmost known occurrences of Lower Devonian medium-thick bedded dolomites. An important bathymetric boundary (basin slope of Matti and others, 1975) existed a few miles to the west.

The possible tectonic origin of this slope (downdropping of the continental margin by extension in the Silurian marginal basin) has been discussed by Johnson and Potter (1975).

Coils Creek Member and Denay Limestone deposition in late Early and early Middle Devonian time occurred in significantly greater depths than those of earlier units. Slope and outer shelf environments are likely for these limestone units. An important feature of the Denay at this longitude is the existence of probable time-equivalent shallow-water limestones in the eastern part of the map area, suggesting that Denay deposition occurred near a break in slope, and that deeper conditions prevailed in the west.

The slope break as evidenced by the Denay Limestone must have shifted at least 10 miles east, from its Early Devonian position between the Monitor and Hot Creek Ranges. This slope migration need not have been tectonic in origin, because progressively deepening conditions are evidenced by lithologies and fossils from the Lower Devonian Unnamed Dolomite upward through the Middle Devonian Denay Limestone. Therefore, transgression followed by progradation westward of shallow-water sediments probably accounts for the position of the slope break during Denay time.

Shallow regimes (some were probably intertidal), represented by the Middle Devonian dolomite and the Devils Gate Limestone, returned after Denay time.

Mississippian limestones and fine clastic rocks bear clasts of radiolarian chert and volcanic rocks, indicating that the Antler Orogenic Highland to the west significantly affected sedimentation in the map area, as in all of the Great Basin. An important effect of the Antler Orogeny was the thrust emplacement of the Late Devonian Woodruff Formation over Middle Devonian dolomites and possibly over the Devils Gate Formation as well.

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APPENDIX

Faunal Lists and Localities

All localities in Morey Peak 7-½ minute quadrangle
Identifications, age assignments, and notes by
J. G. Johnson except as noted.

Section: SUMMIT 8782
Location: Approximately 12,000 feet N 20°W from Mahogany Peak summit.
Sample: EP-101 (silicified)
Footage: 40 feet

- Dalejina sp. 7
- Levenea sp. 4
- Coelospira cf. pseudocamilla 4
- Meristella sp. 34
- Howellella cf. cycloptera 9
- indet. solitary tetracorals 16

Age: A Trematospira community analog, perhaps a little older than Trematospira Subzone. There is a fauna like this below the Trematospira Zone at McColley Canyon, Sulphur Spring Range. It has conodonts indicative of Spinoplasia Zone age. This is appropriate because of the Howellella in this collection.

Sample: EP-102 (silicified)
Footage: 265 feet

- Meristella sp. 4
- Costispirifer dobbinensis 319
- Indet. solitary tetracoral 1

Age: Costispirifer Subzone

Sample: EP-104
Footage: 500 feet

- Dalejina? sp. 5
- Carinagypa loweryi 3
- Strophonella cf. punctulifera 2
- Atrypa sp. (coarse ribs) 4
- indet. spiriferid sp. (coarse ribs) 1
- Phacops sp. 5
- Conocardium nevadensis 13

Age: Lower Pinyonensis Zone

Sample: EP-105.1
Footage: 660 feet

Muriferella sp. 6
"Strophochonetes" ? sp. C 18
Innuitella? sp. 2

Age: Early Devonian
Note: This is the "Strophochonetes"? sp. that occurs at 700 feet at Lone Mtn.

Sample: EP-105.5
Footage: 1041 feet

"Strophochonetes"? sp. C 22

Age: Uncertain, but may be E. Dev.
Note: This is the "Strophochonetes"? sp. C that occurs at 700 ft. at Lone Mtn. Probably in the range of Polygnathus laticostatus at Lone Mtn. (LM-27-33) - G. Klapper.

Sample: EP-105.7
Footage: 1320 feet

indet. dalmanellid sp. (carinate?) 1
Phragmostrophia? sp. 2
Chonetes? sp. 3
Atrypa sp. 4
byrozoans, indet. 6

Age: Eifelian based on conodonts Polygnathus costatus costatus (identified by Klapper)

Sample: EP-105.9
Footage: 1503 feet

Leptathyris cf. L. circula 3 slabs+42

Note: Collection is partly silicified. Some specimens look more elongate than L. circula.
Age: probably Circula Zone

Sample: EP-107
Footage: 1843 feet

indet. atrypid sp. 1
ambocoeliids? indet. 13 slabs

Age: Middle Devonian or Frasnian.

Sample: EP-200 Conodonts identified by G. Klapper
Location: 15,900 feet S55E from NW corner of quadrangle
Anchignathodus typicalis
A. isarcicus
Note: *A. isarcicus* occurs in a limited interval in the earliest Triassic
Age: Early Triassic (Griesbachian)
Spirorbis (identified by E. Yochelson)
indet. ammonoids - abundant

Section: DEVILS CAVE RIDGE
Location: approximately 4800 feet N 62°W of the SW corner of Sect. 31, T. 10 N., R. 51 E.
Sample: EP-301 (silicified)
Footage: 75 feet
Meristella sp. 7
Howellella cf. *cycloptera* 7
Megakozlowskiella sp. 1
Age: this appears to be the same fauna as EP-101.

Sample: EP-302 (silicified)
Footage: 280 feet
Meristella sp. 4
Costispirifer dobbinensis 25
crinoid "roots" 3
Age: *Costispirifer* Subzone

Sample: EP-303 (silicified)
Footage: 430 feet
Acrospirifer kobehana 55
Age: *Kobehana* Zone

Sample: EP-204
Footage: 460 feet
Parachonetes macrostriatus 11
Meristina sp. 1
indet. large ostracode 1
Age: *Pinyonensis* Zone

Sample: EP-400 Conodonts identified by G. Klapper
Location: 5900 feet 833E from the NW corner of the quadrangle
Mixed fauna
"Youngest element (*Siphonodella crenulata*, several species of *Gnathodus*) is roughly about at the Kinderhook-Osage boundary. Oldest element is lower Famennian. I cannot

decide whether this is reworking or stratigraphic leak without knowing something about the stratigraphic and pertologic evidence" - Klapper

Sample: EP-500 Brachiopods identified by J. L. Carter
Location: $SE\frac{1}{4}$, $NE\frac{1}{4}$, Sect. 29, T. 10 N., R. 51 E.

Avonia (?Quasiavonia) n. sp. (large collection but ornament not well preserved)
?Eomarginifera sp. (just a hunch - poorly preserved)
?Antiquatonia sp. (one specimen seems to have an ear ridge - a diagnostic character, but the ribbing is too fine to suit me)
Ovatia sp. (cosmopolitan in Tournaisian-Visean)
Echinoconchus or Overtonia (I just can't make out the speine pattern well enough to decide which)
Rugosochonetes n. sp. (an unusual elongate specimen)
?Rhipidomella sp. (one lousy specimen)
Martinia sp.
Eumetria sp. (definite generic ID here - only known from Tour.-Visean to me anyway)
Indeterminate reticulariacean
Indeterminate choristitinine

Age: Visean

Sample: EP-700 (silicified)
Location: 10,900 feet N12W of Mahogany Peak summit

Schizophoria sp. 76
Indet. smooth gypidulid sp. 1
Parapholidostrophia? sp. 1
Indet. atrypid sp. 1
Echinocoelia sp. 1
Indet. solitary tetracoral 1

Age: probably Middle Devonian

Sample: EP-800 Brachiopods identified by J. L. Carter
Location: $NW\frac{1}{4}$, $NE\frac{1}{4}$, Sect. 20, T. 10 N., R. 51 E.

Pseudosyrinx sp. (strictly Osagian in North America)
Dimegelasma or Verkhotomia (can't determine which with these specimens but would be Osagian in any case)
Indet. impunctate spiriferid
Echinoconchus alternatus (typical Upper Burlington or Keokuk form)
Buxtonia sp.
Indet. stenoscismatacean

Age: Osagian-Upper Tournaisian

Sample: EP-900 (silicified)
Location: 15,200 feet N1E of Mahogany Peak summit
Cassidirostrum? sp. 2
Leiorhynchus sp. 1
Spinatrypa sp. 4
Variatrypa cf. exoleta
Nucleospira sp. 4
Warrenella sp. 6
Cyrtina sp. 1
Age: Probably Givetian, about Castanea Zone

Sample: KDL-MP-1
Location: 11,900 feet N10.5W of Mahogany Peak summit
Leiorhynchus sp. 59
Leptathyris circula 307
Warrenella cf. columbina columbina 14
Age: L. circula Zone, Eifelian

Sample: KDL-MP-2
Location: 200 feet E of NW corner, Sect. 7, T. 10 N., R. 51 E.
indet. brachiopods
indet. chonetid
tentaculites (thin-shelled) abund.
Age: E. Devonian or younger; probably Devonian

Sample: KDL-MP-3 (silicified)
Location: same as EP-303
Dalejina? sp. 1
Meristella? sp. 1
Acrospirifer kobehana 12
Age: Kobehana Zone

Sample: KDL-MP-4 (silicified)
Location: 11,600 feet N11W of Mahogany Peak summit
Warrenella praekirki (numerous)
Age: Early Middle Devonian

Sample: J-7 (silicified)
Location: 10,600 feet N20W of Mahogany Peak summit - from float
Teichertina americana 1
Pentamerella sp. 2
Variatrypa sp. 3
Spinatrypina sp. 6
Age: L. circula Zone, Eifelian

Sample: J-10 (silicified)
Location: same as J-7; also a float sample
Pentamerella wintereri 1
Variatrypa licta 4
Spinatrypina sp. 1
indet. smooth brachiopod 1
Age: *L. circula* Zone, Eifelian

Sample: J-32 Tabulate coral identified by R. A. Flory
Location: 3500 feet N3W of SW corner of Sect. 31, T. 10 N., R. 31 E.
Thamnopora altanevadana
indet. stromatoporoids
indet. crinoids
thin-shelled tentaculites
Age: "Eifelian, probably in the lower part" - Flory

Sample: 1505bh identified by W. A. Oliver, Jr.
Location: at 1505 feet in the Summit 8782 Section
Hexagonaria cf. H. ? sp. h Merriam

Height in section	Styliolina fissurella	Striatostyliolina roemeri (?)	Nowakia sp. indet (?)	N. barrandei (?)	N. holynensis	N. richteri (?)	N. zlichovensis (?)	N. otomari (?)	Viriatella procera
1580	X					X			
1520	X					X			
1468	X					X			
1440	X					X			
1370	X					X			
1310	X	X						X	X
1260	X	X	X				X		
1235	X	X	X	X	X				
1070	X						X		
1030	X						X		
860	X						X		

Occurrences of thin-shelled tentaculites in the Summit 8782 measured section. Collected and identified by C. S. Eldridge (1974). Heights in section refer to Plate 3.