AN ABSTRACT OF THE THESIS OF

Peter F. Cowell for the degree of Master of Science

in Geology presented on June 13, 1986

Title: STRUCTURE AND STRATIGRAPHY OF PART OF THE NORTHERN FISH CREEK RANGE, EUREKA COUNTY, NEVADA

Abstract Approved: Redacted for privacy

J. G. Johnson

Low-angle normal (denudational) faults separate the lithostratigraphic units in the study area into three structural packets. Lithostratigraphic and biostratigraphic studies indicate that the structural packets are locally derived. The lowest packet (Ordovician packet) is composed of, in ascending order, the Antelope Valley Limestone, Copenhagen Formation, Eureka Quartzite, and the Hanson Creek Formation. The main denudational fault in the area (MDF) separates the Ordovician packet from the overlying Lone Mountain Dolomite packet (composed mostly of Lone Mountain Dolomite) and Devonian packet. The Devonian packet is composed of, in ascending order, Beacon Peak Dolomite, the Bartine Member of the McColley Canyon Formation, Sadler Ranch Formation, Oxyoke Canyon Sandstone, Sentinel Mountain-Bay State Dolomites undivided, and the Devils Gate Limestone. Unconformably overlying the Sentinel Mountain-Bay State Dolomites undivided and the Devils Gate Limestone are members of the lower Oligocene Pinto Peak Rhyolite. The Lone Mountain Dolomite packet is separated from the overlying Devonian packet by another denudational fault (LMDDF) that merges with the MDF. A west-dipping

high-angle normal fault offsets the Lone Mountain Dolomite packet, the LMDDF, and the Devonian packet (including the volcanics). Stratigraphic offset along this fault is greatest at the center and decreases toward the ends where it merges with the MDF; thus the fault is shaped like a shovel. A right-lateral tear fault offsets the Lone Mountain Dolomite and Devonian packets, the LMDDF, and the shovel fault by about 3.25 km.

During the Cretaceous, the Ordovician packet was folded and an igneous body was emplaced under the Fish Creek Range. Heat generated from this igneous body caused low-grade metamorphism in the Ordovician packet. The Lone Mountain Dolomite and Devonian packets were probably detached from the Ordovician packet because they were not heated or folded. Uplift of the area accompanied deformation and the emplacement of the igneous body, and prompted initial gravity sliding of the Lone Mountain Dolomite packet off of the Ordovician packet along the MDF. The Devonian packet was carried on top of the Lone Mountain Dolomite packet. Igneous activity in the area was renewed in the early Oligocene. Uplift continued and caused movement of the Devonian packet (including volcanics) off of the Lone Mountain Dolomite packet along the LMDDF, and further movement of the Lone Mountain Dolomite packet over the Ordovician packet along the MDF. During this movement the shovel fault formed. Further movement of the upper two structural packets over the Ordovician packet produced the tear fault. Early Oligocene granite crosscuts the MDF, indicating that final emplacement of the upper two structural packets over the Ordovician packet was completed during the early Oligocene.

Structure and Stratigraphy of part of the Northern Fish Creek Range, Eureka County, Nevada

by

Peter F. Cowell

A THESIS

submitted to

Oregon State University

in partial fulfillment of
the requirements for tne
the degree of

Master of Science

Completed June 13, 1986

Commencement June 1987

APPROVED:

Redacted for privacy

Professor of Geology in charge of major

Redacted for privacy

Chairman of Department of Geology

Redacted for privacy

Dean of Graduate School

Date thesis is presented June 13, 1986

Typed by Peter F. Cowell for Peter F. Cowell

ACKNOWLEDGEMENTS

This project was funded in part by research grants from the Amerada Hess Corporation, Amoco Production Company, Marathon Oil Company, Dr. J.G. Johnson, and my parents.

I would like to thank Dr. J.G. Johnson for generously sharing his time and knowledge. Jana Thompson and Claudia Regier helped prepare conodont samples for identification by Dr. R.L. Ethington of the University of Missouri-Columbia and Dr. Gilbert Klapper of the University of Iowa. Dr. K.F. Oles and Dr. R.D. Lawrence reviewed the thesis and served on the thesis committee. Dr. Karen Lund provided insightful discussions on Basin and Range geology.

My fellow students at O.S.U. should be especially acknowledged for their support and input. Randy Keller and Greg Stromberg provided office facilities. Members of "The J Team" provided encouragement in the field and at O.S.U. Dana Willis provided drafting services. Special thanks to Amy "Hog" Hoover for all her support and input. I would especially like to thank Jana Thompson for her support and encouragement.

TABLE OF CONTENTS

LIST OF FIGURES

Figure

LIST OF PLATES
(in pocket)

STRUCTURE AND STRATIGRAPHY OF PART OF THE NORTHERN FISH CREEK RANGE, EUREKA COUNTY, NEVADA

INTRODUCTION

Purpose

Investigations in the region of the study area have been primarily stratigraphic in emphasis, and little was known about the structure of the area. The primary purpose of this project was to map and study part of the northern Fish Creek Range, Eureka County, Nevada (Fig. 1, 2, and 3). The most detailed map that includes the entire study area is the Eureka County geologic map of Roberts and others (1967). In compiling this map, Roberts and others modified information for much of the northern part of the Fish Creek Range from the map of Hague (1892). The units that Hague mapped are the Ordovician Pogonip Group, Ordovician Eureka Quartzite, Silurian Lone Mountain Dolomite, and "undifferentiated" Devonian rocks. Essentially, no structural interpretations were made by Hague, although he noted (1892, p.56, 57) that in the study area "different horizons" of the Lone Mountain and the Devonian "repose directly upon" and "overlap the quartzite." Roberts and others (1967) interpreted the contact between the Ordovician sequences and the Silurian and Devonian sequences shown on Hague's map as a thrust fault that juxtaposes the Ordovician sequences over the Silurian and Devonian sequences. Stewart (1980, p.81, Fig. 40) implied that the age of the thrust is

Mesozoic, perhaps based upon Nolan's (Nolan, 1962, p.28; Nolan and others, 1974) indication that the age of compressional deformation in the area of the Eureka mining district, located northeast of the study area (Fig. 3), is Mesozoic.

The main emphases of this study were: (1) to map and determine the stratigraphy of all rock units in the area; (2) to map the fault shown on the Eureka County map of Roberts and others (1967); (3) to determine the stratigraphic levels at which the fault(s) lie; and (4) to determine the relative movement and age of the fault(s) in order to relate movements to patterns of regional tectonics. Part of the geologic map of the Pinto Summit quadrangle (Nolan and others, 1974) is also revised in this study.

Location and Description

The Fish Creek Range is located in central Nevada, in the southeast corner of Eureka County and the northern part of Nye County (Fig. 2). The study area covers the southern half of the northern part of the range, and is centered approximately 18 km southwest of Eureka (Fig. 2). The area encompasses parts of U.S.G.S. Bellevue Peak and Pinto Summit 15-minute topographic quadrangle maps. Access to the area is via jeep trails that lead off U.S. Highway 50, which is located east of the area.

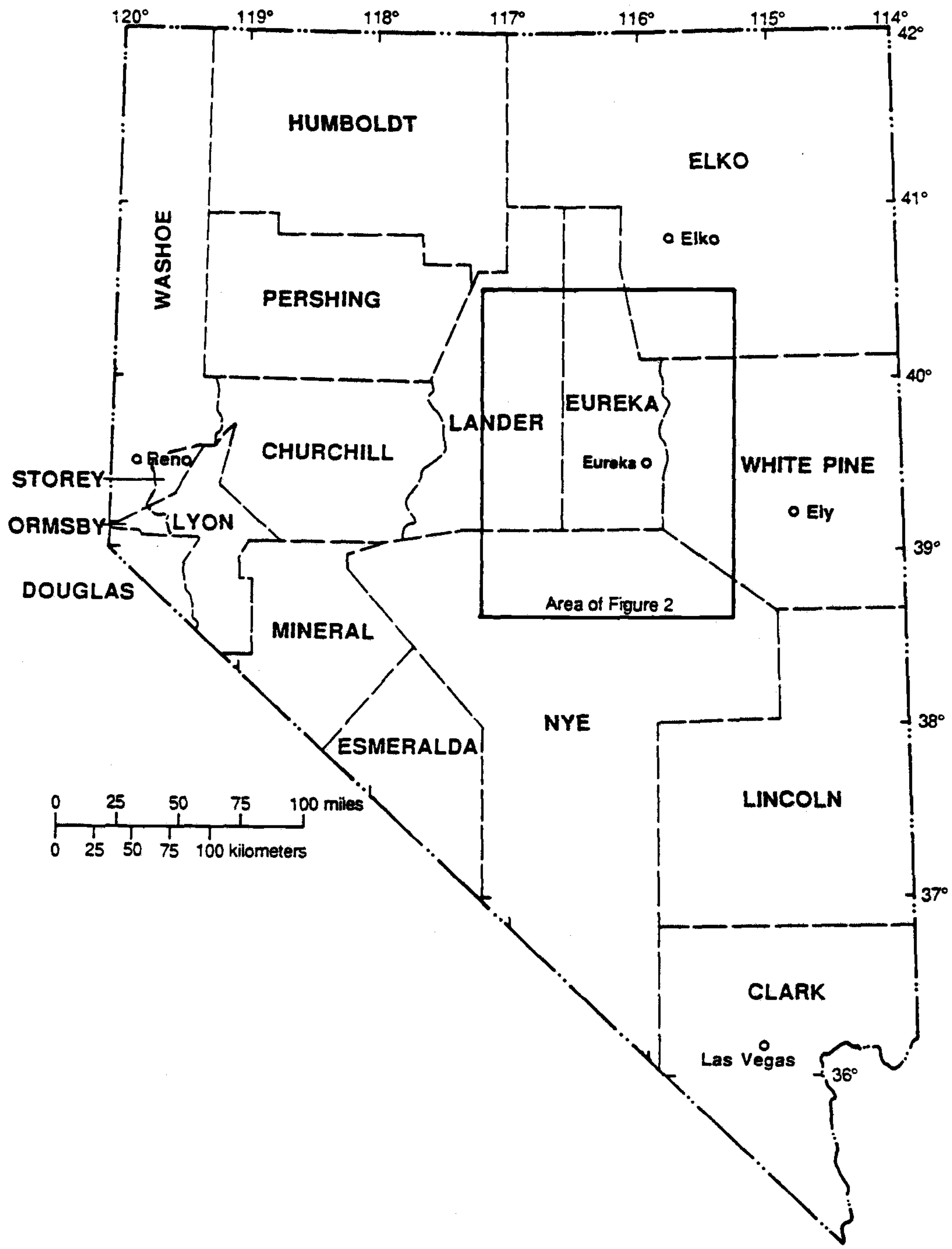

Figure 1. Index map of Nevada.

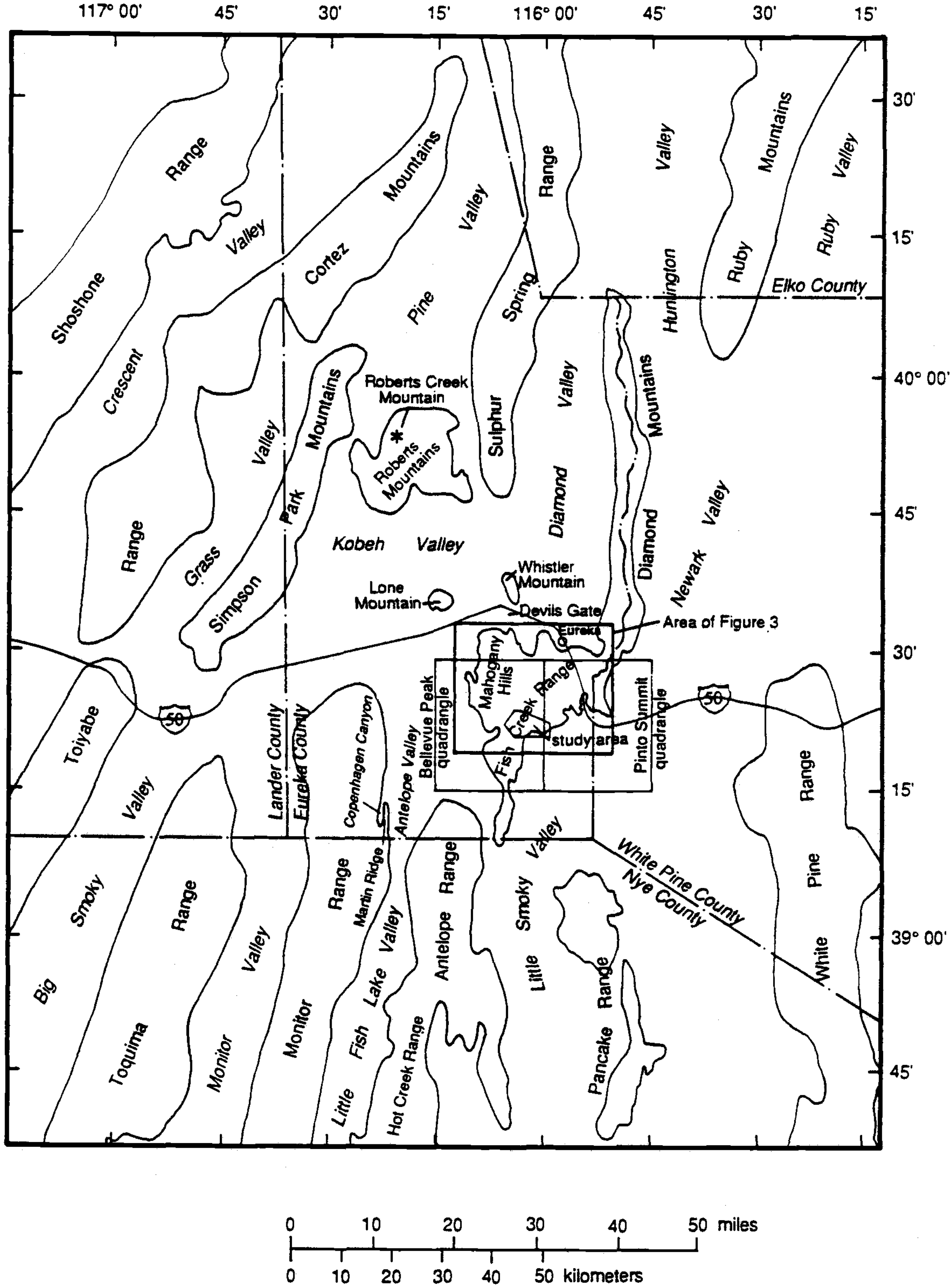

Figure 2. Index map of the Eureka region.

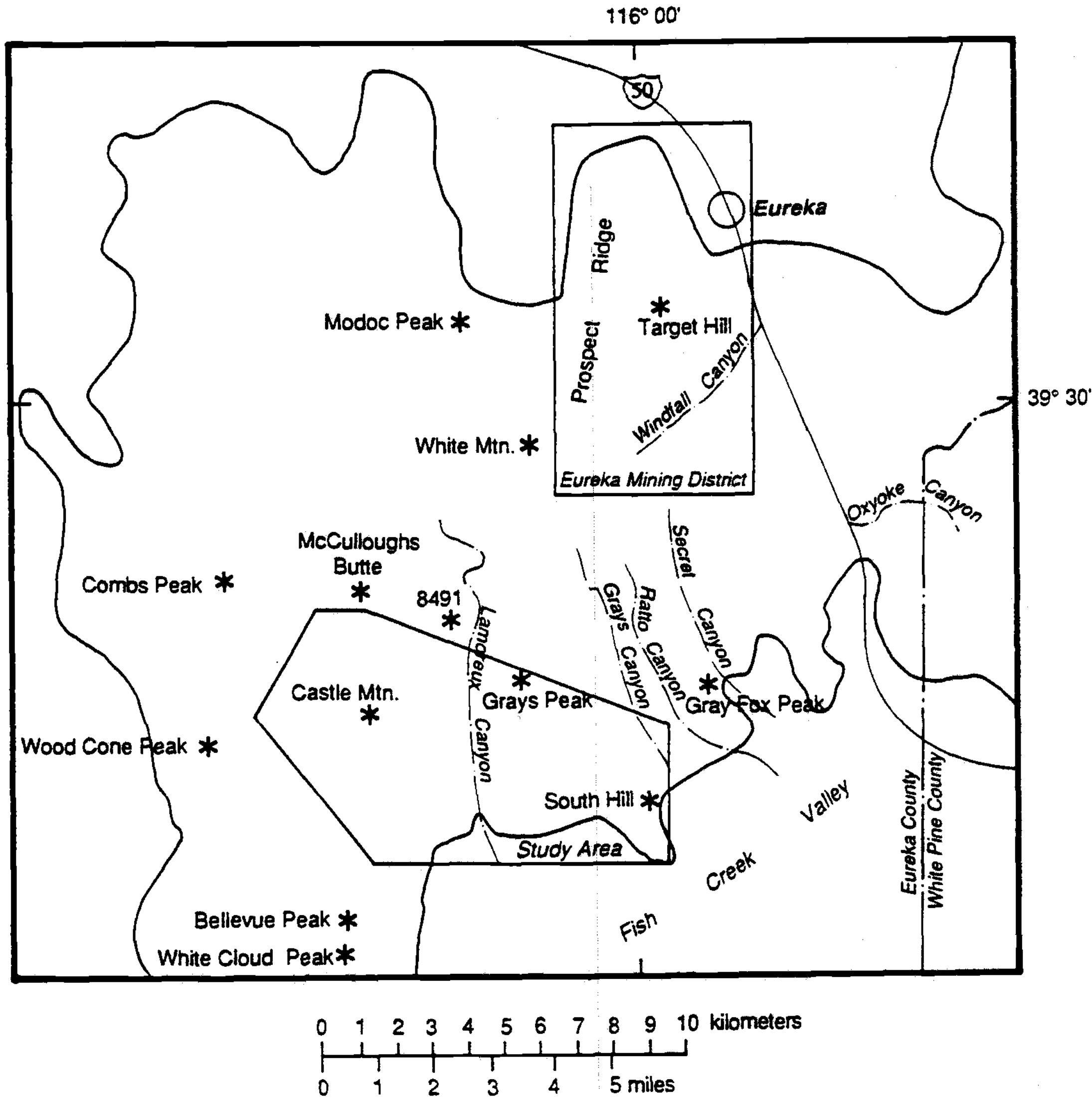

Figure 3. Index map of the northern Fish Creek Range and vicinity.

Regional Geologic Setting

During early and middle Paleozoic time, present-day Nevada was part of the Cordilleran geosyncline that extended from California to the Canadian Arctic (Roberts, 1972). The continental margin was located in central Nevada and marginal basins and volcanic island arc systems were located to the west (Burchfiel and Davis, 1975). Three depositional belts were present during this time (Roberts and others, 1958). The deposits of the eastern belt (eastern assemblage) represent shelf deposits and are characterized by shallow-water dolomites, quartz arenites, and shelly limestones. A transitional zone represents deposition on the outer shelf and upper continental slope, and is characterized by carbonates and fine clastics. The western assemblage rocks represent an oceanic depositional complex composed of deep-water carbonates, bedded chert, fine clastics, and volcanics.

Early Paleozoic time in Nevada is considered to have been a time of little major tectonic activity (Burchfiel and Davis, 1975). During Late Devonian time, the stratigraphic sequence was disrupted by the Antler orogeny (Johnson and Pendergast, 1981).

The primary structural feature formed during the Antler orogeny was the Roberts Mountains thrust (Fig.4). From Late Devonian to earliest Mississippian time, the Roberts Mountains allochthon, composed of western assemblage deposits, moved relatively eastward; it rode over and deformed the slope and westernmost shelf deposits of the miogeosyncline (Roberts, 1972; Johnson and Pendergast, 1981). Large amounts of siliceous and volcanic debris derived from the newly formed highland were shed to the west into the eugeosyncline and eastward into a developing foreland basin (Poole, 1974; Johnson and Pendergast, 1981).

During Late Permian and Early Triassic time, the Sonoma orogeny occurred. It resulted in eastward movement of ocean-floor sediments over the eroded shallow-water deposits on the Antler highland. Compressional tectonism in Nevada continued during the Mesozoic. By Middle Jurassic time, much of Nevada was probably elevated and tectonically active (Stewart, 1980, p.76). During Jurassic to early Tertiary time, most of Nevada was in the hinterland of the Sevier orogenic belt (Armstrong, 1968), a north-northeast-trending belt of large-scale overthrusting and folding in southern Nevada and central Utah (Fig.4). During this time the hinterland was a site of deformation, plutonism, and regional metamorphism (Coney, 1980).

The sparsity of early Tertiary rocks in Nevada suggests that the region was elevated and undergoing erosion during the early Tertiary (Stewart, 1980, p.105). During the mid-Tertiary, the region was subjected to a complex history of volcanism and crustal extension (Zoback, 1981). At this time, low-angle normal fault movement was common in east-central Nevada (Armstrong, 1972; Coney, 1980), much of which is associated with the development of the metamorphic core complexes in this area. Basin and Range faulting that resulted in the present topography began in approximately mid-Miocene time and continues to the present (Zoback and others, 1981).

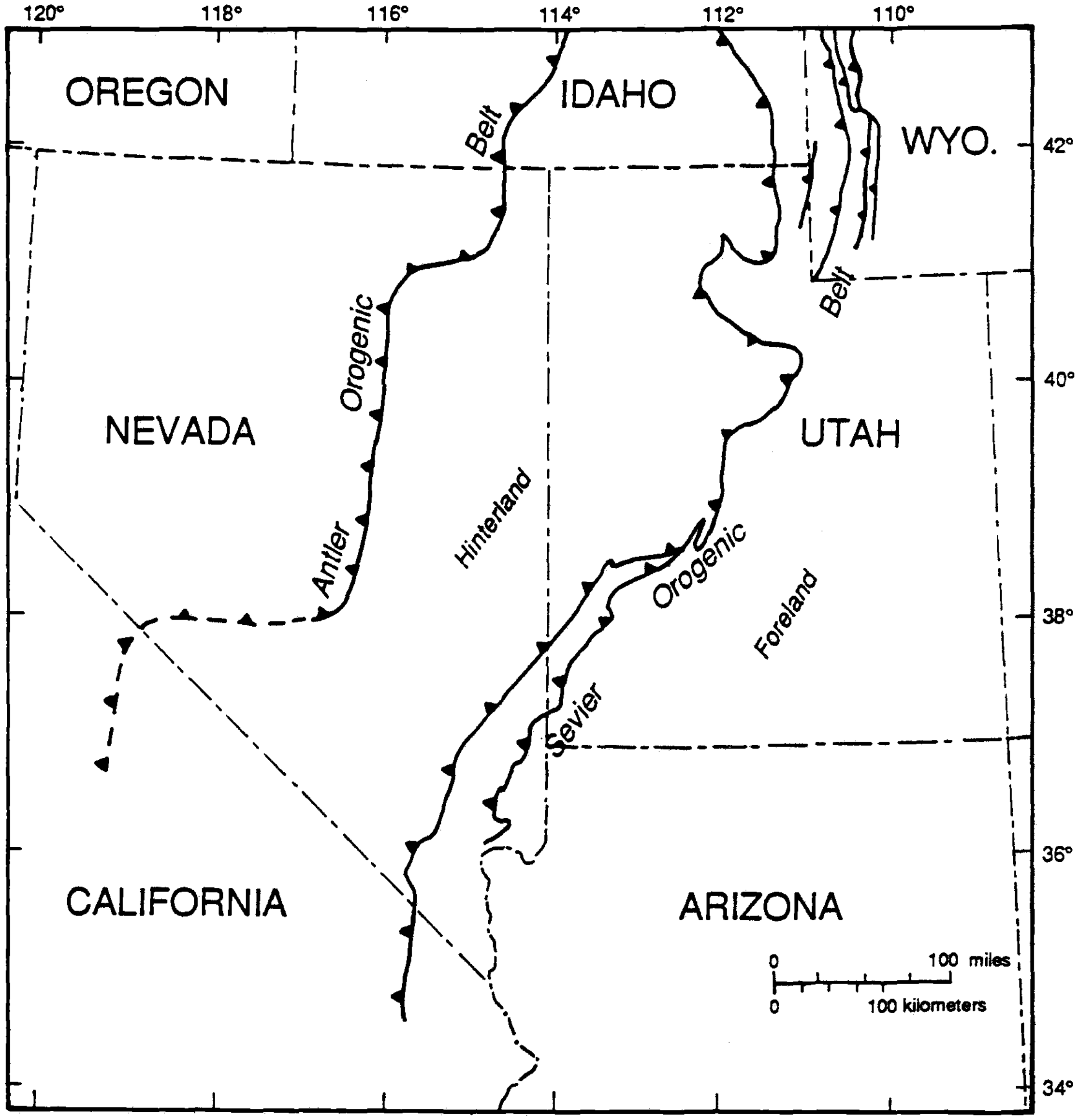

Figure 4. Map showing location of Antler orogenic belt, Sevier orogenic belt, and the foreland and hinterland of the Sevier oroenic belt. Modified from Armstrong (1972).

Paleozoic Stratigraphic Nomenclature

Nolan and others (1956) published a study of the Paleozoic and Mesozoic stratigraphic section in the vicinity of Eureka; they reviewed and modified the original work done in the area by King (1878), Hague (1883, 1892), and Walcott (1884), and later work done by Kirk (1933) and Merriam (1940). Nolan and others' (1956) Cambrian, Ordovician, and Silurian formational nomenclature is still the standard for the region. Devonian formational nomenclature of Nolan and others (1956) consisted of the Lower and Middle Devonian Nevada Formation overlain by the Devils Gate Limestone (of Merriam, 1940). Nolan and others defined five members of the Nevada Formation in the Diamond Range. In ascending order, they are: Beacon Peak Dolomite, Oxyoke Canyon Sandstone, Sentinel Mountain Dolomite, Woodpecker Limestone, and Bay State Dolomite. These units are not applicable to all areas west of the Diamond Range. Carlisle and others (1957) proposed a new set of names for rocks assigned to the Nevada Formation in the Sulphur Spring and Pinyon Ranges. In ascending order they are: McColley Canyon Member, Union Mountain Member, and Telegraph Canyon Member. Johnson (1962) raised the McColley Canyon Member to formational rank. Murphy and Gronberg (1970) divided this formation into three members.

Each of the five members of the Nevada Formation defined by Nolan and others was recently raised to formation status (Hose and others, 1982), thus raising the rank of the Nevada Formation to group status. This is a useful change, as noted by Kendall and others

(1983), "because the Nevada was a collection of distinctive rock units deposited during a particular interval of time in a variety of depositional environments, not only those that coexisted along depositional strike, but also environments that coexisted across depositional strike." Kendall and others defined three distinct belts of Lower and Middle Devonian lithofacies across the region: (1) a western belt, predominantly limestone and including slope deposited and basinal rocks (the Denay Limestone); (2) an eastern belt, predominantly dolomite of shallow-water origin (strata of the Diamond Range); and (3) a transitional belt in which the major lithofacies intertongue. Kendall (1975; Kendall and others, 1983) has shown that eastern and western formations, members, and tongues encompass all mappable rock units in the transitional belt, except for a new formation named by Kendall (1975; Kendall and others, 1983), the Sadler Ranch Formation. Johnson and Murphy (1984) proposed the most recent change in the Devonian formational nomenclature, the abandonment of the Union Mountain Formation and the Telegraph Canyon Formation.

Paleozoic rock units in the study area are all of the eastern assemblage (Roberts, 1972), and the Devonian units lie within the eastern and transitional belt as defined by Kendall and others (1983). Cambrian, Ordovician, and Silurian formational nomenclature used in this report is that of Nolan and others (1956). Devonian nomenclature used is that of Johnson (1962, McColley Canyon Formation), Murphy and Gronberg (1970, members of the McColley Canyon Formation), Hose and others (1982, members of the Nevada Formation of Nolan and others raised to formation status), Kendall and others

(1983, Sadler Ranch Formation), and the Devils Gate Limestone of Merriam (1940). The reader is referred to these works for definitions, regional descriptions, and detailed regional relations of units. Regional relations are shown in Plate 1. Units present in the study area and their generalized relations to each other are shown in Figure 5.

Terminology

The carbonate classification of Dunham (1962) and the sandstone classification of Dott (1964) are used for hand sample and thin section description. Crystal size designation for recrystallized carbonates follows the scheme proposed by Folk (1962): <0.004 mm aphanocrystalline, 0.004 mm to 0.016 mm very finely crystalline, 0.016 mm to 0.025 mm finely crystalline, 0.025 mm to 0.062 mm medium crystalline, 0.062 mm to 1.0 mm coarsely crystalline, 1.0 mm to 4.0 mm very coarsely crystalline. The terminology of McKee and Weir (1953) is used to describe bed thickness and splitting properties. Rock colors were determined by comparison with the Geological Society of America Rock-Color Chart (1963).

PERIOD	ROCK UNIT		THICKNESS (meters)	GENERALIZED RELATIONS
Quaternary / Tertiary	unconsolidated sediments		>40	T-Qu
Tertiary	Pinto Peak Rhyolite	Rhyolite Dome	?	Td
Tertiary	Pinto Peak Rhyolite	Sierra Springs Tuff	>30	Tss
Tertiary	Pinto Peak Rhyolite	Pinto Basin Tuff	>30	Trt
Devonian	Devils Gate Limestone		200 to 250 incomplete	Ddg
Devonian	Bay State Dolomite Sentinel Mountain Dolomite		310	Dsm-bs
Devonian	Oxyoke Canyon Sandstone		15 to 52	Doc
Devonian	Sadler Ranch Formation		73	Dsr
Devonian	McColley Canyon Fm	Coils Creek Mbr	37 to 73	Dmc
Devonian	McColley Canyon Fm	Bartine Member		Dmb
Devonian	Beacon Peak Dolomite		38	Dbp
Devonian / Silurian	Lone Mountain Dolomite		500	S-Dlm
Silurian	Hanson Creek Formation		91	O-Shc
Ordovician	Eureka Quartzite		167	Oe
Ordovician	Copenhagen Formation		?	Oc
Ordovician	Pogonip Group	Antelope Valley Limestone	275+ incomplete	Oav
Ordovician	Pogonip Group	Ninemile Formation	not exposed	On
Ordovician	Pogonip Group	Goodwin Limestone	not exposed	Og
Cambrian	Windfall Formation		?	€w
Cambrian	Dunderberg Shale		?	€d

Figure 5. Summary stratigraphic section for part of the northern Fish Creek Range, Eureka County, Nevada.

STRATIGRAPHY

Dunderberg Shale

Description

The Dunderberg Shale is the oldest stratigraphic unit in the study area and is located in the northeast corner (Plate 2). The unit consists of float, with the exception of a few road cuts in which disrupted bedding can be observed. The unit is composed of yellowish brown shale interbedded with thinly laminated light to medium light gray lime mudstone that weathers grayish orange to light brown. Bed thickness ranges from 0.6 cm to 45 cm.

Thickness

The nature of the exposures and the absence of exposed contacts does not allow for accurate thickness determination of the Dunderberg Shale. Dunderberg Shale that crops out approximately 0.2 km north of the exposures in study area is approximately 70 m thick (J.M. Bird, personal commun., 1986). Nolan and others (1956, p.19) measured 80 m of Dunderberg Shale near the type locality which is 8.5 km northeast of the map area at the head of Windfall Canyon, where the lower parts of the unit were removed by faulting (Palmer, 1960). At Cherry Creek, about 105 km northeast of Eureka, a complete section of Dunderberg Shale is 103 m thick (Palmer, 1960). In the central Nevada region, the unit rests conformably on the Hamburg Dolomite and is overlain concordantly by the Windfall Formation.

Age

Although the unit is fossiliferous elsewhere (Nolan and others, 1956, p.19), no fossils were found in it within the study area. Palmer (in Nolan and others, 1956, p.19) has reported an age of late Dresbachan and early Franconian (Late Cambrian) for the Dunderberg Shale.

Windfall Formation

Description

The Windfall Formation in the study area consists of float in the canyon just north of Grays Canyon in the northeast corner (Plate 2). The float at this locality contains angular lime mudstone, wackestone, and grainstone pieces up to 30 cm long. Pieces weather light to medium light gray and light brownish gray, and are moderate red to light to medium light gray on fresh fracture. Some of the mudstones are thinly laminated and contain rip-up clasts. Other mudstone pieces are mottled as a result of burrowing. Wackestones consist of fossil hash in lime mud matrix. The only recognizable fossil pieces are crinoid fragments as large as 6.0 mm in diameter. The grainstones are composed of hematite-stained ooids that average

0.25 mm in diameter set in a matrix of calcite cement. Most pieces of the Windfall Formation have calcite-filled fractures.

Thickness and age

In the Eureka area, the Windfall Formation is conformably overlain by the Goodwin Limestone (Nolan and others, 1956, Plate 2), and is 198 m thick at the type locality. Approximately 2.5 km north of exposures of the Windfall Formation described here are exposures of the unit that are approximately 190 m thick (J.M. Bird, personal commun., 1986). Palmer (in Nolan and others, 1956, p.20) reported the age of the Windfall Formation in the Eureka area as Franconian and Trempealeauan (Late Cambrian).

Antelope Valley Limestone

Description

In the study area, strata assigned to the Antelope Valley Limestone are generally covered by vegetation and exposure is poor, except for the west side of Castle Mountain (Plate 2), where exposure is good. The unit consists of interbedded lime mudstones, wackestones, packstones, and grainstones that have rough weathered surfaces. The contacts between rock types are gradational. The formation generally forms cliffy outcrops and the mudstones weather recessively.

The mudstones are laminated to thin-bedded with shaly splitting. They weather medium light gray to dark gray; some have yellowish gray and pink sparry calcite-filled fractures that compose as much as 25 percent of the rock. Fresh surfaces are light gray to medium light gray. The mudstones are composed of an average of 85 to 90 percent lime mud that commonly has been recrystallized to microspar and pseudospar. In some samples pseudospar is limited to burrows, in others it is predominant. Gastropod and pelecypod fragments are present in a few samples and constitute from 2.0 to 8.0 percent of the rock. Some of these fossils have been recrystallized to pseudospar and to sparry calcite. Crinoid stem pieces are common, a few have syntaxial overgrowths. Some samples contain as much as 12 percent well sorted, very fine sand- to silt-size quartz grains. Many of these grains have been partially replaced by calcite. Undulatory laminations up to a few mm thick are common in the mudstones, and are defined by different crystal sizes or by quartz grains. Many of the mudstones are mottled from burrowing. A few rip-up clasts and peloids are present. Approximately 2.0 percent fracture porosity exists in the mudstones.

Wackestones, packstones, and grainstones of the Antelope Valley Limestone weather light gray to grayish black and are medium dark gray to grayish black on fresh fracture. They are thin- to thick-bedded with blocky to massive splitting. Fragments of brachiopods, pelecypods, and pieces of crinoid stems with syntaxial overgrowths are the most common fossils. Some of the overgrowths have been abraded. Gastropod, trilobite, and ostracod pieces are also present.

The fossils are randomly distributed and randomly oriented. Many fossils have been recrystallized to sparry calcite. Silt-size quartz grains, some with quartz overgrowths on rounded quartz grains, compose between 5.0 to 7.0 percent of the rock. Up to 5.0 percent micritized pellets are also present. Rounded mudstone and wackestone intraclasts are common. Cement in the rocks consists of a mosaic of clear sparry calcite crystals. Porosity is about 2.0 percent, in the form of vugs, interparticle void spaces, and fractures.

Several mudstone samples exibit a crude foliation that is visible in thin section and at times in hand sample. The foliation is defined by parallel alignment of elongate, equal size calcite crystals that have fairly straight intercrystalline boundaries. Sizes of the crystals vary from sample to sample and range from very finely crystalline to finely crystalline. In one sample a quartz vein was folded, and folded calcite veins are visible at outcrop. Calcite crystals can be seen to "drape around" burrows in some samples. A foliated packstone sample collected near Castle Mountain (Plate 2) contains dark streaks and lenses of black, possibly carbonaceous(?) material that occur parallel to the foliation and in some places define the foliation. These lenses average 0.1 mm wide and as much as 0.5 mm long. The black material appears to be a "stain" on and a filling between the calcite crystals. Calcite crystals in the lenses are smaller than those of the surrounding rock. The deformation in these rocks is best developed on the west slope of Castle Mountain.

All of the Antelope Valley Limestone samples examined show hematite staining. Diagenetic quartz and chalcedony are also common.

Stylolites are abundant throughout the unit. The stylolites and the foliation indicate that the formation has undergone some form of stress.

Thickness and contacts

The estimated thickness of the formation in the study area is 275+ m. The lower contact of the unit with the underlying Ninemile Formation is not exposed, and the upper contact is covered by blocks of talus from the overlying Eureka Quartzite. At the type section in the Antelope Range, Merriam (1963, p.24) estimated the formation to be 366+ m thick. At White Mountain, located approximately 5.0 km northeast of the study area in the Eureka district, the unit is 335 m thick (J.M. Bird, personel commun., 1986). In the Antelope Valley area, the Antelope Valley Limestone is comformably overlain by the Copenhagen Formation, which is in turn overlain by the Eureka Quartzite (Merriam, 1963, p.19). In the Eureka district, the Eureka Quartzite overlies the Antelope Valley Limestone and the contact between the two is an unconformity (Nolan and others, 1956, p.25,28). At Lone Mountain, much of the upper part of the Antelope Valley is missing (Ross, 1970, p.33, 37). At the east side of Bellevue Peak, located just south of the southwest corner of the study area (Fig. 3), Merriam (1963, p.30) noted an edgewise mud-breccia, which he interpreted to be a result of pre-Eureka emergence, between the Eureka and the upper Antelope Valley Limestone. Although the contact between the two units in the study area is covered, the transition is abrupt. Fossil samples collected at or very near the Antelope Valley-Eureka contact are variable in age (see below and Appendix I). This, along with the abruptness of the contact between the formations, indicates that in the study area the contact is a disconformity.

Age

In the Antelope Valley area, the Antelope Valley Limestone is Whiterockian (Early Ordovician) in age (Merriam, 1963, p.25). Conodonts from the formation in the study area range in age from late Whiterockian to Chazyan (Middle Ordovician). This straddles the regional Antelope Valley Limestone - Copenhagen Formation time boundary. The lowest unit of the Copenhagen Formation, a calcite-cemented sandstone (unit A of Merriam, 1963, p.26), was not recognized in the study area. Because no recognizable stratigraphic break is present, the entire unit underlying the Eureka quartzite has been assigned to the Antelope Valley Limestone, even though some of the beds are equivalent in age to lower parts of the Copenhagen Formation (Plate 1).

Copenhagen Formation

Description

Limestones that are located along the western side of the southern border of the map area (Plate 2) are assigned to the Copenhagen Formation based on lithology and age. Exposure is fair, being hindered by an abundance of trees and float. The unit consists of lime mudstones, wackestones, and packstones that weather yellowish gray, grayish orange, and medium dark gray. Fresh surfaces are medium dark gray. Bedding varies from thin-bedded to laminated. The unit splits into angular slabby and flaggy blocks and shaly pieces.

Bivalve, crinoid, and brachiopod fossil debris occur and many fragments are recrystallized to sparry calcite. The mudstones exibit both laminations and mottling. Some rocks are hematite stained. Calcite-filled fractures are common. Lithologically, the rocks resemble those of unit B of Merriam (1963, p.26).

Contacts and thickness

The upper contact of the formation is covered by debris from the overlying Eureka Quartzite. As mentioned, the lower contact of the Eureka Quartzite with the Antelope Valley Limestone in the study area is a disconformity. This unconformable lower contact at the base of the Eureka probably is true for this locality as well. The lower contact of the Copenhagen Formation is not exposed, so thickness is undetermined. At the type section on Martin Ridge, approximately 30 km southwest of the study area (Fig. 2), Merriam (1963, p.26) measured 122 m of Copenhagen Formation that conformably overlies the Antelope Valley Limestone.

Age

The dominant species of conodonts recovered from this unit in the study area, P. aculeata, ranges from the Black Riveran through the Kirkfieldian (Middle Ordovician). This places the unit within the stratigraphic interval of the upper Copenhagen Formation (R.L. Ethington, written commun., 1985).

Eureka Quartzite

Description

Approximately one-third of the exposed rock in the map area is Eureka Quartzite (Plate 2). It is brecciated and jointed so that accurate determination of attitudes is difficult. Exposure ranges from poor to excellent. Locally, the quartzite has been recrystallized by mineralizing solutions. Hague (1883, p.262; 1892, p.54-57) originally named the formation for exposures in and surrounding the thesis area. In 1933, Kirk (p.33) redefined the type locality as Lone Mountain because the exposures in the Fish Creek Range are disrupted. The Eureka is very resistant and forms cliffy outcrops that cap the high ridges and peaks in the area. Bedding is very thick-bedded to unrecognizable in most cases. Locally, it is very thin-bedded to laminated. In the southwest corner of the area, float blocks with high angle (30 degrees) planar cross-laminated sets that average 4.0 cm thick, are common. The unit splits into massive to slabby angular blocks.

At Lone Mountain, Webb (1958, p.2341-2342) divided the Eureka Quartzite into three informal members. A 10.7 m thick member that consists of yellowish brown- to dark reddish brown-weathering vitreous quartzite and whitish weathering quartzite with purple banding is termed the "lower discolored quartzite member". The middle, or "white quartzite member" is 29 m thick and consists of white quartzite that often weathers red to orange on joint and out-crop surfaces. The middle member also contains a cross-bedded zone 14 m to 16 m from the base. The uppermost member (15.5 m thick) is the "upper gray quartzite member" and is composed of cross-bedded whitish quartzite and weathers light gray to brownish gray at its base. This is overlain by a dark brown to gray interval that is capped by a thin (about one m) bluish gray, medium-grained, quartz sandy dolomite. Kirk (1933), assigned the sandy dolomite beds to the overlying Hanson Creek Formation, a convention followed in this report. To the south and east of the Eureka area (Hot Creek Range, Grant and White Pine Range), there is a "shaly quartzite member" of the Eureka Quartzite (Webb, 1958, p.2348-2351). Webb noted that this unit represents a facies of pre-Eureka (type area) rocks and is separated from the main body of the formation by an unconformity.

This unit extends into Utah (Plate 1).

All of the lithologies observed by Webb at the type locality are present within the study area, but the disrupted exposure does not allow for recognition of stratigraphic relations or thickness determinations of the different lithologies. In general, the quartzite in the study area is white to medium dark gray on weathered and fresh surfaces. Moderate yellowish brown to grayish orange pink staining is common on weathered surfaces and in fractures. Most of the unit is medium-grained, moderately well sorted and rounded quartz arenite. In thin section, the quartz grains are clear, with the exception of numerous fluid inclusions and some rutile needles. Monocrystalline grains that exibit straight extinction predominate.

Regionally, the Eureka Quartzite is cemented by quartz overgrowths. However, in the majority of samples from this study, there is very little cement. The grain-to-grain contacts average 30 percent straight, 35 percent concavo-convex, and 35 percent sutured. The high percentage of interpenetrating contacts indicates that the sandstone has undergone a great deal of stress, resulting in pressure solution (Pettijohn, Potter, and Siever, 1973, p.92, table 3-5).

One sample contains approximately 2.0 to 3.0 percent muscovite that grew between grains or partially replaced grains. The muscovite may have crystallized out of a mineralizing solution, or it may be the result of metamorphism of clays already present in the quartzite. Pressure solution could have concentrated insoluble clays along the grain boundaries prior to or during metamorphism.

Hematite staining is common throughout the formation and may have caused the gray coloration. A few scattered, elongate zircon grains are present in most of the samples studied. Where cross-bedded, the bedding stands out on weathered surfaces and is a result of differences in grain size and sorting. Cross-bedded pieces have patches of poorly cemented grains which generally weather free from the rock; this results in approximately 7.0 percent porosity. In the rest of the unit porosity is 1.0 to 2.0 percent.

Contacts

The lower contact of the Eureka Quartzite is an disconformity. The quartzite caps most of the high peaks and ridges in the area and the upper surface is eroded. Where the normal stratigraphic sequence has been preserved, the Hanson Creek Formation overlies the Eureka Quartzite. Because the Eureka is resistant, it stands above the Hanson Creek Formation topographically and has shed debris onto the contact, obscuring it in most places. However, southeast of Dave Keane Spring and directly north of Castle Mountain (at fossil locality 10, Plate 2), the basal part of the Hanson Creek Formation can be seen to be in contact with the Eureka Quartzite. The basal Hanson Creek is medium dark to dark gray dolomitic quartz arenite. Sand content decreases and carbonate content increases in the Hanson Creek Formation up section away from the contact. This basal sandstone of the Hanson Creek Formation has been noted by workers in other areas. Potter (1976, p.27) stated that in the Hot Creek Range the upper contact of the Eureka Quartzite seems to be gradational with the overlying Hanson Creek Formation. Merriam (1940) noted the presence of the basal sandstone at Lone Mountain (p.21) and in the Roberts Mountains (p.10,11). He stated that there is no clear evidence to indicate post-Eureka erosion and suggested that the sandstone at the base of the Hanson Creek Formation may be transitional with the underlying Eureka Quartzite. However, Kirk (1933, p.28-31) reported that in the Roberts Mountains the basal sandstone of the Hanson Creek Formation rests upon an uneven surface of the Eureka Quartzite.

Thickness

The brecciated and jointed nature of the formation in the study area makes thickness determinations difficult. The maximum thickness of the Eureka Quartzite in the study area is estimated to be 167 m. Hague (1892, p.56) estimated the incomplete section of Eureka exposed on the southwest spur of Castle Mountain (Plate 2) to be 91 m thick.

Age

The age of the formation in the field area cannot be determined direcly, but it can be bracketed by the age of the underlying and overlying units. The youngest unit underlying the Eureka Quartzite is the upper part of the Copenhagen Formation. Conodonts recovered from the Copenhagen Formation could range up through the Kirkfieldian (late Middle Ordovician). Conodonts recovered from the overlying Hanson Creek Formation are Late Ordovician in age and could possibly be as old as the Shermanian (late Middle Ordovician). Thus, the Eureka Quartzite in this study is probably late Middle Ordovician to early(?) Late Ordovician in age. This is consistent with the absence of the "shaly quartzite member" and the disconformable lower contact with Middle and Lower Ordovician rocks (Plate 1).

Hanson Creek Formation

Description

In the study area, the Hanson Creek Formation is throughly brecciated, is dominated by dolomite and limestone, and is poorly exposed. The unit weathers recessively compared to the underlying Eureka Quartzite and the overlying Lone Mountain Dolomite. Weathered surfaces are very rough. Weathered and fresh surfaces are medium light gray to grayish black; some limestones weather grayish orange. Bedding ranges from massive to thin-bedded. Splitting is blocky to slabby. Sparry calcite has filled the numerous fractures in the rock. Porosity in the unit is 3.0 percent and pore spaces are localized along fractures.

The basal unit of the Hanson Creek Formation is a dolomitic quartz arenite. This sandstone consists of as much as 85 percent rounded and well sorted fine sand-size quartz grains. Some of the grains have quartz overgrowths that have been abraded. Many of the grain-to-grain contacts are concavo-convex and some are sutured, indicating that the rock has undergone some kind of stress. Many of the quartz grains have been partially replaced by dolomite. The dolomite in the sandstones is cloudy and has few enfacial triple junctions; it was probably originally lime mud rather than void-filling cement. Crinoid fragments are locally abundant in this basal sandstone. Up section from the contact, the percentage of quartz grains decreases and the amount of dolomite increases until the unit is composed completely of dolomite several meters above the contact.

The dolomites of the formation were probably originally lime mud that had been recrystallized to microspar and pseudospar, and was then dolomitized. Much of what was lime mud is in the form of rounded blebs approximately 0.06 mm in diameter, and may have originally been oolites or fecal pellets that were micritized. Fossils in the dolomites are rare, compose approximately 2.0 percent of the rock, and consist of recrystallized ghosts.

Limestones form the upper part of the Hanson Creek Formation in the study area, and consist of mudstones and crinoidal wackestones. The limestones form slopes covered with float. In many areas, moderate reddish brown chert is abundant in the float.

Thickness

The upper contact of the Hanson Creek Formation in the map area is a fault that separates it from the Lone Mountain Dolomite. In many places, the Hanson Creek has been completely removed by movement along this fault. Estimated maximum thickness of the Hanson Creek Formation in the study area is approximately 91 m. Throughout Eureka

County, the thickness of the unit varies from 91 m to 170 m (Dunham, 1977).

Age

Conodonts from the Hanson Creek Formation in the study area are Late Ordovician in age. In a study of the Hanson Creek Formation in the vicinity of Wood Cone Peak and in the Mountain Boy Range, Ross and others (1979) concluded that the formation is Late Ordovician and Early Silurian in age. Mullens and Poole (1972) and Murphy and others (1979) arrived at the same age range for the formation in other parts of Eureka County.

Lone Mountain Dolomite

Description

Exposure of the Lone Mountain Dolomite is good to excellent. Throughout most of the area, the unit is bounded by faults, and in places much of the section has been removed by faulting. Locally, the unit is fractured. The most complete section of the formation is located on the south flank of Grays Peak mountain (Plate 2), where three units can be recognized.

The lowermost unit (unit one) is a slope-former and is composed predominantly of thin-bedded brown-weathering limy dolomite mudstones and wackestones that are dark gray to brownish gray on fresh fracture. Rocks of unit one are best exposed 1.5 km southwest of Grays Peak (Plate 2), where the unit is approximately 130 m thick and is in fault contact with the Eureka Quartzite and the Hanson Creek Formation. The rocks emit a strong fetid odor when broken. The size of crystals in rocks of this unit range from very finely crystalline to medium crystalline. Weathered surfaces of the more finely crystalline rocks are locally laminated. Fossils are recrystallized and consist of disarticulated brachiopods, gastropods, and bivalves. Porosity is generally 1.0 percent or less, in the form of fractures, vugs, and fossil molds. Stylolites are rare.

Unit two conformably overlies unit one and composes most of the Lone Mountain Dolomite in the study area. Unit two forms massive cliffs and is composed predominantly of gray, massive, medium to coarsely crystalline, blocky weathering dolomite. This dolomite commonly is brecciated. Where brecciated the rock is composed of as much as 3.0 percent quartz cement that fills the fractures and forms a fine "net" of quartz. Outcrops weather yellowish gray, light olive gray, light gray, and very light gray. Fresh surfaces are medium light gray to very light gray. Weathered surfaces are rough and have a sugary texture. In thin section, the rocks are composed of a mosaic of clean, anhedral dolomite crystals. Generally, original features have been destroyed by dolomitization, but locally a few recrystallized bivalve fragments, crinoid stem pieces, brachiopod pieces, and tabulate corals are present. In places, unit two has as much as 6.0 percent porosity. Pore spaces are predominantly intercrystalline and vug. Moldic pore spaces are also present.

Stylolites are rare.

Locally, interbedded with the massive gray dolomite of unit two are laterally discontinuous sequences of interbedded light gray and brown limy dolomites (unit three). The best exposures of unit three are 2.0 km north of summit 7155 (Plate 2), where the unit obtains a maximum thickness of approximately 100 m. These rocks weather recessively compared to the enclosing massive gray dolomites and form benches. Most of unit three weathers dark yellowish brown and is dusky yellowish brown to medium dark gray on fresh fracture. Interbedded with these brown-weathering rocks are yellowish gray- to light gray-weathering rocks that are light olive gray to olive gray on fresh fracture. When fractured, the brown-weathering rocks emit a weak to strong fetid odor. Bedding in unit three varies from thin-bedded to thinly laminated. Crystal size ranges from finely crystalline to aphanocrystalline.

In thin section, the dolomite crystals of unit three are cloudy and brown in plane light. Original textures are preserved in the form of fine laminations and recrystallized fossils. Locally, the thinly laminated rocks have fenestral pore spaces that are filled by sparry cement and lime mud intraclasts up to 2.0 cm long and 1.0 cm wide, which suggests these rocks were originally algal mats. Where thin-bedded, unit three is locally a fossiliferous wackestone. Bivalves, gastropods, brachiopods, and ostracods(?) are common. Porosity in unit three is generally 1.0 percent. Pore spaces are in the form of fossil molds and fractures. Stylolites are rare.

Contacts

The lower contact of the Lone Mountain Dolomite in the study area is a fault that places the unit in contact with the Hanson Creek Formation and the Eureka Quartzite. Movement along this fault has removed unit one from much of the Lone Mountain Dolomite. The contact between unit one and unit two is gradational. Contacts between unit two and unit three are gradational, as are contacts between the light gray beds and the brown beds of unit three. The upper contact of the formation is a fault in most places, except in the vicinity of summit 7155. Near summit 7155, the massive gray dolomite that composes unit two of the Lone Mountain Dolomite is abruptly overlain by finely crystalline, light gray dolomites of the Beacon Peak Dolomite. The abrupt nature of the contact and the shallow water, high-energy deposits of the Beacon Peak Dolomite (described below) indicate that this contact is probably an unconformity. An unconformity between the Lone Mountain Dolomite and the Beacon Peak Dolomite has been noted by other workers in the region (Nolan and others, 1956, p.38, 41 [Eureka area]; Schalla, 1978, p.19-20 [Mahogany Hills]; Johnson and Murphy, 1984).

Thickness

Accurate thickness determination of the Lone Mountain Dolomite is difficult because the unit is disrupted by faults. Maximum thickness is estimated to be 500 m. Regionally, the thickness varies from 479 m at the type section at Lone Mountain to 668 m at Roberts Creek Mountain (Merriam, 1963, p.40).

Age

In the study area, conodonts from the Lone Mountain Dolomite range from Llandovery through early Wenlockian (Early Silurian) in age. In the Eureka area, Nolan and others (1956, p.39) indicate that the formation may range in age from the Silurian into the Early Devonian. At other localities, workers have documented that the unit ranges into the Early Devonian (Johnson and others, 1973 [Roberts Mountains]; Murphy, _in_ Colman, 1979 [Sulphur Spring Range]). With this in mind, it is likely that in the study area the formation ranges into the Early Devonian (Plate 1).

Beacon Peak Dolomite

Description

Exposure of the Beacon Peak Dolomite in the study area is good. The unit is composed of light gray, finely crystalline dolomite that forms benches and rubbly slopes and weathers recessively compared to the Lone Mountain Dolomite. Most of the outcrops are fractured and bedding is obscured. Sparry calcite and quartz crystals fill the fractures and compose as much as 5.0 to 10 percent of the rock. Near faults, the unit has been recrystallized to coarsely crystalline clear dolomite.

The recrystallization and the fractured nature of the outcrops often makes differentiation between the Beacon Peak Dolomite and the Lone Mountain Dolomite difficult. In faulted areas, the Beacon Peak Dolomite is brecciated, with angular clasts of the dolomite, as much as 18 cm long, set in a calcite matrix. Where undisturbed, the unit is thin- to very thin-bedded and splits into angular flaggy and slabby pieces.

The formation is predominantly composed of dolomite mudstones and intraclastic peloidal dolomite packstones and grainstones. Neither fossil material nor evidence of burrowing is present. Weathered surfaces are generally very light gray to medium light gray. In places, the unit weathers yellowish gray, olive gray, or pinkish gray. Fresh surfaces are commonly very light gray to light gray. Interbedded with the gray dolomites are limy dolomites that are pale yellowish brown to yellowish gray on weathered surfaces and medium dark gray to dusky yellowish brown on fresh surfaces. These interbeds are laterally discontinuous and form about 10 percent of the unit; when fractured they emit a strong fetid odor.

Most of the unit consists of finely crystalline dolomite that has abundant fenestra, peloids, and elongate intraclastic rip-ups that are as much as several cm long. The peloids are micritized and are composed of aphanocrystalline dolomite crystals that are brown in plane light. Peloids range from 0.6 mm to 0.05 mm in diameter and are ovoid to elongate. No internal structures are discernible in the peloids. In studies of the Beacon Peak Dolomite in the central Nevada region, workers have noted both ooids and pellets (Nolan and others, 1956, p.42; Kendall, 1975, p.20; Schalla, 1978 p.37; Long, 1973, p.48, 49). With this in mind, it is likely that the smaller peloids are remnants of both ooids and pellets, and the larger peloids are intraclasts.

The larger intraclasts are composed of laminated fenestral aphanocrystalline dolomite, the lithic type of much of the unit. An algal origin for the rock is indicated by the fenestra, the aphano-crystalline crystal size, and the lack of biotics and burrowing. Outcrops are commonly laminated, reflecting the algal material, fenestra, and variations in peloid size. Some laminae are undulatory.

The peloids and intraclasts generally occur in a finely crystal-line, cloudy dolomite matrix and form dolomite packstones. However, the interstices of some of the dolomites are filled with medium to coarsely crystalline clear dolomite spar. This spar exhibits a competitive growth fabric and many enfacial triple junctions, and was probably originally sparry calcite cement that formed a grainstone. The peloidal and intraclastic deposits, the presence of probable algal mats, and the lack of biotics suggest that much of the Beacon Peak Dolomite was deposited in shallow water, high energy conditions. Finely crystalline to aphanocrystalline dolomite mudstones with no internal structures also are present in the formation. Dolomite crystals in these mudstones are cloudy. These structureless dolomites could be the result of algal mats.

Outcrops of Beacon Peak located on the west side of the mouth of Lamoreux Canyon, and approximately 1.6 km south-southwest of the summit of South Hill (Plate 2), locally contain sandy quartz dolomite beds. Quartz grains are medium- to fine-grained, subangular to rounded, and well sorted. The grains are suspended in a fine to medium crystalline dolomite matrix. A few thin quartz arenite beds beds also are present in the Lamoreux Canyon outcrops.

Stylolites are rare in the formation. Hematite staining is common, possibly accounting for the pinkish coloration in areas. Porosity is generally 1.0 to 2.0 percent in the form of vugs, fractures, and fenestra.

Thickness and contacts

In the study area, the Beacon Peak Dolomite is about 38 m thick. Where the lower contact is depositional, the Beacon Peak rests unconformably on the Lone Mountain Dolomite. The upper contact of the formation is covered by float from the overlying Bartine Member of the McColley Canyon Formation. In the Mahogany Hills, this upper contact is abrupt and may represent a depositional hiatus consequent on deepening (Schalla, 1978, p.24). In the southern part of the Fish Creek Range, the contact is not as abrupt, but is gradational over a thickness of about 30 cm (Long, 1973, p.47).

Age

The overlying Bartine Member of the McColley Canyon Formation falls within the *gronbergi* conodont Zone (Early Devonian) at its base. The underlying Lone Mountain Dolomite is probably as young as Early Devonian. Thus, an Early Devonian age for the Beacon Peak Dolomite is estimated.

McColley Canyon Formation

Bartine Member

The McColley Canyon Formation is represented in the study area primarily by the Bartine Member. This unit is composed of abundantly fossiliferous limestone that weathers recessively compared to most other units in the area, and forms debris slopes. The Bartine is thin- to very thin-bedded and splits into angular platy to slabby pieces. The unit weathers medium dark gray to light gray and pale yellowish gray to very pale orange. On fresh surfaces, the limestone is medium light gray to light olive gray. The recessive weathering, abundant fossils, and the yellowish orange weathered color makes the Bartine Member easily identifiable in the field.

Porosity in the unit is generally less than 1.0 percent and is localized in vugs and fractures. Fractures are cemented with sparry calcite and quartz. Diagenetic chalcedony is present in a few samples studied and commonly replaces brachiopod and pelecypod shell material. Hematite staining is common. Stylolites are rare. Orange colored chert lenses, up to 30 cm thick, are present locally.

Fossil debris composes up to 60 percent of the rock. Brachiopods, whole and disarticulated, are the most abundant fossils. Whole solitary rugose and favositid corals also are present. Fragments of trilobites, pelecypods, crinoids, bryozoans, and gastropods are common. Texturally, the Bartine Member consists predominantly of mudstones, bioclastic pelloidal packstones, and pelloidal packstones.

Locally, dark gray encrinites are present. In places, pellets compose up to 30 percent of the rock, are evenly distributed, well sorted, and thoroughly micritized. Average pellet size varies from 0.04 mm to 0.06 mm. Microspar and pseudospar form the matrix material of the unit. Where the Bartine is a mudstone, it has a homogeneous texture, probably from bioturbation. Moderately to well sorted, very fine sand-size quartz grains are scattered through much of the unit and average about 3.0 percent. In a few places, white quartzose sandstone tongues, similar to the Oxyoke Canyon Sandstone, are present.

In the South Hill area, Nolan and others (1974) mapped the Bartine Member as the Grays Canyon Limestone (see Appendix II for details).

Coils Creek Member

The exposure of McColley Canyon Formation directly south of benchmark 7253 (Plate 2, mapped as McColley Canyon Formation undivided) consists of the Bartine Member underlying rocks that would best be classified as part of the Coils Creek Member of the formation. These rocks are thick- to thin-bedded laminated lime mudstones that are more resistant to weathering than the Bartine. Weathered surfaces are medium to light gray with a moderate red to light red tinge in areas. Fresh surfaces are brownish gray to medium gray. Pink argillaceous partings are common. Some areas are silicified. The contact with the Bartine Member is covered, but the transition is not abrupt, suggesting the contact is gradational.

To the west in the Mahogany Hills, the Coils Creek Member conformably overlies the Bartine Member of the formation (Schalla, 1978, p.42). Kendall (1975, p.80) noted that the Coils Creek at Modoc Peak is lithologically transitional between Bartine and type Coils Creek. Johnson and Murphy (1984, their Fig. 2) indicate that the Coils Creek Member overlying the Bartine pinches out between the Mahogany Hills and the Diamond Range (Plate 1). The exposure mapped as McColley Canyon Formation undivided (Plate 2) probably represents this event in the study area, the upper parts of the Bartine in the study area being correlative with the lower Coils Creek to the west (Plate 1).

Contacts

The contact between the Bartine Member and the underlying Beacon Peak Dolomite has been described. In the study area, the upper contact of the Bartine Member is covered by debris from the overlying Sadler Ranch Formation. In the southern part of the Fish Creek Range, the contact between the Bartine and the Sadler Ranch is gradational (Long, 1973, p.55). In the Sulphur Spring Range, the Sadler Ranch Formation overlies the Bartine Member with apparent disconformity (Kendall, 1975 p.52).

Thickness

The poor exposure of the formation makes accurate thickness determination difficult; thickness ranges from about 37 m to 73 m.

Age

Conodonts from the formation could range from the *dehiscens* to *serotinus* Zones (Early Devonian). Sample 360, from the southern end of South Hill, indicates that the Bartine lies within the *inversus* Zone. Brachiopods from the unit range from faunal intervals 11 through 13 of Johnson (1977) and Johnson and others (1980). These faunal intervals are coexistent with the *gronbergi* and part of the *inversus* conodont Zones. Thus, the Bartine in the study area can probably be restricted to the *gronbergi* and *inversus* Zones. This is consistent with the age range of the unit regionally (Plate 1). At Modoc Peak, north of the study area, the Bartine contains brachiopods from faunal intervals 11 to 12 at the base, and conodonts from the *inversus* Zone at the top (Kendall and others, 1983, Fig.5). At Oxyoke Canyon, east of the study area, a tongue of Bartine is interbedded with the upper Beacon Peak Dolomite and lies within the *inversus* Zone (Plate 1) (Kendall and others, 1983, Fig.5). In the southern part of Fish Creek Range, the Bartine lies within the *gronbergi* and *inversus* Zones (Johnson, 1978, p.21, Fig.4).

Sadler Ranch Formation

Description

The Sadler Ranch Formation consists of dolomite mudstones to crinoidal packstones that are dark gray to light olive gray on fresh surfaces and weather light brownish gray to light olive gray. The dark beds generally emit a strong fetid odor when fractured. The unit is thin- to thick-bedded, splits into angular slabby to blocky pieces, and forms slopes and ledges. Locally, the unit is highly fractured and the fractures are filled with sparry calcite cement. Kendall (1975, p.83; Kendall and others, 1983) has divided the formation into three units: 1) a lower dolomite, 2) a middle crinoidal dolomite, and 3) an upper dolomite. All three are present in the study area.

The dolomites of the formation are composed of finely crystalline anhedral to subhedral crystals that form an interlocking mosaic. The crystals are cloudy in plane light. Patches of medium crystalline dolomite that are clear in plane light may represent biotics.

The lower dolomite is composed of dark limy dolomite and dolomite mudstones. The mudstones are both structureless and finely laminated. Laminations result from variations in the amount of fossil debris and the cloudiness of dolomite crystals. Scattered crinoid ossicles are common in the lower dolomite.

The middle crinoidal dolomite is characterized by abundant crinoidal debris, mostly ossicles up to 2.0 mm across. Many of the ossicles are dilumenate. Both light and dark beds are present in the middle crinoidal dolomite which is composed primarily of dolomite wackestones and packstones. Mudstones also are present in the middle crinoidal dolomite. The transition between the middle crinoidal dolomite and the lower and upper dolomites is gradational.

The upper dolomite is similar to the lower dolomite with the exception of interbeds of light gray dolomite. These light gray dolomites closely resemble the dolomite interbeds of the overlying Oxyoke Canyon Sandstone and increase in abundance up section.

Hematite staining is common throughout the unit and locally stylolites are present. Porosity is approximately 2.0 percent and occurs as vugs, intercrystalline pore spaces, and fractures.

In the South Hill area, Nolan and others (1974) mapped the Sadler Ranch Formation as the Sentinel Mountain Dolomite (see Appendix II for details).

Thickness and contacts

The formation is approximately 73 m thick in the study area. The contact with the underlying McColley Canyon Formation is discussed in the previous section. The gradational upper contact with the overlying Oxyoke Canyon Sandstone is well exposed in most places. The abundance of light dolomite interbeds in the Sadler Ranch Formation increases near the upper contact; these grade into sandy dolomite and dolomitic sandstone of the Oxyoke Canyon Sandstone over an approximately 5 m thick interval. The mapped contact between the two units is at the base of the first quartzose dolomite.

Age

Conodonts from the Sadler Ranch Formation could range from the Early Devonian *serotinus* Zone to the Middle Devonian *costatus* Zone (Appendix I). Sample 221 was collected approximately 4.0 km southwest of benchmark 7253 from the middle crinoidal dolomite and contains conodonts characteristic of the *serotinus* Zone. Sample 220 was collected from the upper dolomite at the same locality and contains fauna of the *serotinus* to *patulus* Zones. These two collections, and the age of the underlying McColley Canyon Formation and the overlying Oxyoke Canyon Sandstone, suggest that in the study area the age of the Sadler Ranch Formation can be constrained to the *serotinus* and *patulus* Zones (Plate 1).

Oxyoke Canyon Sandstone

Description

The Oxyoke Canyon Sandstone is composed of dolomitic quartzose sandstones that are interbedded with fine- to medium-crystalline dolomite. Exposure is excellent. The quartzose beds form ledges and low cliffs and the dolomite beds generally form slopes. Weathered and fresh surfaces of sand-rich beds are yellowish gray to medium olive gray. Orange to brownish hematite staining is common in sandy units. Dolomite interbeds are yellowish gray to very light gray on weathered and fresh surfaces. Bedding is poor and ranges from thin-bedded to massive; the rocks split into angular blocks. Locally, quartz veins and quartz-cemented fractures are common, especially in the South Hill and summit 7478 areas (Plate 2). Jointing is common in the more quartz-rich beds. Porosity varies from 1.0 to 6.0 percent, in the form of vugs (as much as 2.0 cm long and 1.0 cm wide), fractures, fossil molds, and intercrystalline pore spaces. Stylolites are rare.

Quartz grains are predominantly medium sand size, rounded to well rounded, and moderately to well sorted. Fine and very fine sand size grains are present locally. Rounded syntaxial quartz overgrowths are common. The quartz is clear, monocrystalline, contains fluid inclusions, and exhibits straight extinction. Contacts between grains are commonly tangential or planar. In a few cases they are concavo-convex. The abundance of tangential and planar contacts and the rarity of stylolites indicates that the formation has undergone little to no pressure solution (Pettijohn, Potter, and Siever, 1973, p.92, table 3-5).

Dolomite is medium to finely crystalline spar. Crystals are somewhat cloudy, typically anhedral to subhedral, and form an interlocking mosaic that lacks a competitive growth fabric and enfacial triple junctions. Dolomite rhombs are scattered within the mosaic. Dark, medium to fine sand-size, rounded and elongate peloids of very finely crystalline to aphanocrystalline dolomite commonly form up to 6.0 percent of the unit. The peloids are structureless and probably represent intraclasts that underwent micritization and recrystallization. Faint recrystallized ghosts also might have been intraclasts. In some of the finely crystalline, quartz-poor parts of the unit, elongate patches of fenestral(?) pore spaces are filled with clear, medium crystalline spar. The lack of competitive growth fabric and enfacial triple junctions, and the cloudiness of the dolomite spar suggest that in the quartz-rich beds dolomite was probably a matrix, rather than a void-filling cement. However, recrystallization and dolomitization could have obscured these features.

Stratigraphic relations within the formation are best exposed in the vicinity of summit 7478 and approximately 1.0 km north-northeast of summit 7155, near fossil locality 387 (Plate 2). Near fossil locality 387, the lower and upper parts of the formation consist of interbedded fine to medium crystalline dolomites that contain about 6.0 percent quartz grains, and quartzose dolomites that contain between 15 and 40 percent quartz grains. Intraclasts are common and constitute as much as 6.0 percent of the rock. Locally, these dolomites are thinly laminated and contain fenestra, molds of brachiopods(?), and burrows(?). Laminations represent variations in the abundance of quartz grains, dolomite matrix, fenestra, and intraclasts. Contacts between lithologies in the lower and upper parts of the formation near fossil locality 387 are gradational. The middle of the formation at this locality is composed of dolomitic quartz arenites in which quartz composes between 60 and 85 percent of the rock. Interbedded with the quartz arenites are a few beds of dolomite and quartzose dolomite. Planar laminations and low-angle cross-laminations are locally abundant in the middle part of the unit and reflect variations in grain size, sorting, and the amount of dolomite matrix. Cliffs up to 6.0 m high are common in the middle of the unit at this locality. Contacts between the quartz arenites and the other lithologies that compose the middle of the unit are sharp and may represent erosional surfaces.

In the area of summit 7478 (Plate 1), the lowermost part of the formation consists of quartz arenites with a few quartzose dolomite interbeds. Up section, the abundance of dolomite increases until quartz is only present in a few quartzose beds and scattered in the matrix. Near the top of the formation, however, quartzose beds and quartz arenite beds become abundant locally.

In the South Hill area, Nolan and others mapped the Oxyoke Canyon Sandstone as the South Hill Sandstone (see Appendix II for details).

Contacts

The gradational contact with the underlying Sadler Ranch Formation has been described. The upper contact with the overlying Sentinel Mountain Dolomite also is gradational; quartzose dolomites of the Oxyoke Canyon Sandstone grade upward into light gray, finely crystalline dolomites of the Sentinel Mountain Dolomite over an interval of approximately 10 m. The mapped contact between the units was placed at the top of the quartzose dolomites.

Thickness

The thickness of the Oxyoke Canyon Sandstone in the study area is variable. In the exposures west of Lamoreux Canyon (Plate 2) the unit is approximately 37 m to 52 m thick. Northeast of summit 7155, the unit is about 52 m thick in the southern exposures, and thins northward along strike to about 30 m. In the South Hill area, the thickness of the unit changes along strike from approximately 50 m to 15 m. Kendall (1975) has shown that the Oxyoke Canyon Sandstone thickens and thins along strike in the Diamond Range and in the Sulphur Spring Range. To the west of the study area, within the Mahogany Hills, Schalla (1978) has documented that the unit is lost by facies transition (Plate 1).

Age

Conodonts from both the lower part of the formation (sample 387) and the upper part (sample 417B) can be restricted to the early Middle Devonian <u>costatus</u> conodont Zone (Appendix I).

Sentinel Mountain Dolomite - Bay State Dolomite

Description

The Sentinel Mountain Dolomite and the younger Bay State Dolomite are lithologically similar. The two formations originally were defined in the Diamond Range (Nolan and others, 1956), where the presence of the Woodpecker Limestone between them is used as the main criterion to separate them. The Woodpecker Limestone thins westward in the Diamond Range and is not present in the study area (Plate 1), thus the Sentinel Mountain Dolomite and the Bay State Dolomite were mapped as a single unit (Plate 2).

The unit as a whole consists of interbedded light and dark dolomites and exposure is fair to excellent. The light beds weather yellowish gray to light olive gray and are light gray to medium dark gray on fresh surfaces. The dark beds weather moderate brown to dusky yellowish brown and are medium dark gray to brownish black when fresh. Bedding ranges from very thin-bedded to thick-bedded; the rocks split into angular slabby and blocky pieces. Calcite-filled partings are present locally in the thin-bedded parts of the unit.

Above the gradational contact with the Oxyoke Canyon Sandstone, the Sentinel Mountain Dolomite is composed of rhythmically alternating light and dark dolomites and limy dolomites that form stairstep slopes. The layers of alternating colored dolomite range from less than 1.0 m to a few m thick. This rhythmically alternating sequence grades up section into the Bay State Dolomite. The Bay State is also composed of alternating light and dark dolomites and limy dolomites, but the alternations are not as regular; darker dolomites are more abundant. Alternating layers in the Bay State are a few m to several m thick. The Bay State also has a blockier, more cliffy topographic expression than the Sentinel Mountain. Contacts between light and dark layers are gradational in both formations. Also, individual light and dark layers are not persistent along strike. In the exposures of the unit west of Lamoreux Canyon (Plate 2), there is a thin zone of dolomitic sandstone at the top of the unit. The sandstone appears to pinch and swell in thickness along strike. Maximum thickness is approximately 18 m. Sandstones are not present in the unit elsewhere in the study area.

In thin section, the rocks are composed predominantly of medium crystalline dolomite. Crystals are both clear and cloudy brown. The cloudy crystals are generally anhedral to subhedral rhombs and form a dense interlocking mosaic. On fresh fracture, the rocks emit a strong to very strong fetid odor, especially the darker beds. This suggests that the dark coloring of the beds and cloudiness of the crystals result from a high organic content. Clear crystals are generally subhedral to euhedral rhombs that form a crystal-supported framework; intercrystalline spaces are void or filled with cloudy, less well-formed dolomite crystals or, in some cases, with silica cement. Mottling is common in much of the unit and represents intermixing of zones of clear and cloudy crystals and differences in crystal size. This mottling could be from differential dolomitization of the rock caused by differences in texture, porosity, or organic content. Laminations also are common, average 1.0 mm to a few mm thick, and are a result of alternating clear and cloudy dolomite layers. In a few of the laminated rocks, the voids between the clear dolomite crystals are filled with silica cement. In outcrop, the silica cemented laminae stand out in relief.

Locally, the unit is finely crystalline. In some places the rocks contain aphanocrystalline rip-up clasts. Dark, spheroidal patches (average diameter 0.06 mm) that are crosscut by dolomite crystal boundaries are present in a few samples. These dark patches might be the remains of micritized fecal pellets or intraclasts. Intraformational breccias are present locally. The breccias average 45 cm thick and are composed of angular pieces of dolomite (as much as 6.0 cm long) in a finely crystalline dolomite matrix.

With the exception of the features mentioned above, the dolomites are structureless; original textures were destroyed by dolomitization. The high organic content, the possible fecal pellets, and the fine laminations suggest that much of the unit was originally lime mud, possibly in the form of algal mats. No textural features observed in the rocks indicate the presence of original carbonate cement. However, textural evidence for cementation could have been destroyed by dolomitization. Locally, the unit is coarsely crystalline, especially where recrystallized near faults.

Fossils in the unit consist of fragments of gastropods, bivalves, crinoids, and brachiopods and are most abundant in the upper (Bay State) part of the unit. Texturally, the fossiliferous parts of the unit are wackestones and packstones. Along the upper eastern flank of South Hill (Plate 2) there are local occurrences of silicified brachiopods. Brachiopods from sample 425 are possibly <u>Geranocephalus</u> (Appendix I). Nolan and others (1974) reported abundant <u>Stringocephalus</u> high on the eastern flank of South Hill (these also may by <u>Geranocephalus</u>). Approximately 1.5 km south of the South Hill summit are what may be silicified <u>Syringopora</u> (sample 378A).

In some areas, the unit is very fractured and has been silicified by numerous silica veinlets and replacement chert. The chert is predominantly moderate yellowish brown, moderate red, and grayish orange. These silicified rocks are on the eastern flank of the hills located on the west side of the canyon that is west of South Hill, south of summit 7155, and south of and on the lower eastern flanks of South Hill (Plate 2). These areas are near faults along which silicifying solutions could have migrated. Hematite staining is common throughout the unit. Stylolites are present locally, but are not abundant. Porosity is generally 1.0 percent, but in places as high as 6.0 percent. Pore spaces are in the form of intercrystalline voids and vugs.

In the South Hill area, this unit was mapped by Nolan and others (1974) as the Bay State Dolomite (see Appendix II for details).

Contacts and thickness

The gradational contact with the underlying Oxyoke Canyon Sandstone has been described. The upper contact with the Devils Gate Limestone is also gradational. This upper contact is best exposed on the ridge 750 m west of the "o" in the word "Canyon" of Lamoreux Canyon (Plate 2). At this locality, the dolomites of the Bay State Dolomite grade upward into the dolomitic sandstone at the top of the formation. The dolomitic sandstone grades upward into moderate brown to yellowish gray limy dolomites and very thin-bedded grayish orange-weathering limestones of the Devils Gate Limestone. This transitional contact is approximately 30 m thick. The mapped contact between the unit and the overlying Devils Gate Limestone is placed at the top of the sandstone. Where the sandstone is not present, the contact is at the base of the first limestone bed. In the area west of Lamoreux Canyon, where the section is complete, the unit is approximately 310 m thick. In the South Hill area and on the west side of the canyon west of South Hill (Plate 2), much of the upper part of the unit has been eroded. In these areas, the unit is locally unconformable below Pinto Peak Rhyolite.

<u>Age</u>

Nolan and others (1956, p.46,47) assigned a Middle Devonian age to the Sentinel Mountain Dolomite, Woodpecker Limestone, and Bay State Dolomite sequence in the Diamond Range. Within the thesis area, "<u>Stringocephalus</u>" (from Nolan and others, 1974), <u>Gerano-cephalus</u>?, and <u>Syringopora</u>? were the only fossils identified. The age range of these fossils, and the fact that the unit is conformably underlain by the lower Middle Devonian (<u>costatus</u> conodont Zone) Oxyoke Canyon Sandstone, and conformably overlain by the upper Middle to Upper Devonian Devils Gate Limestone, confirms a Middle Devonian age for the Sentinel Mountain Dolomite-Bay State Dolomite in the study area (Plate 1).

Devils Gate Limestone

<u>Description</u>

The formation is composed predominantly of lime mudstones, wackestones, and packstones that form benches, slopes, and a few cliffy outcrops up to 10 m high. Exposure is fair to poor. Bedding varies from thick-bedded to laminated and the rock splits into angular platy to blocky pieces. The limestones weather medium light gray to olive gray to dark gray, and grayish orange to dark yellowish orange to very pale orange. Thicker bedded rocks are predominantly grays and thinner bedded rocks are commonly orange. Orangish to pink partings between beds are common. Fresh surfaces are medium light gray to dark gray and olive gray. Yellowish gray to light gray, finely crystalline thin-bedded dolomites and limy dolomites are interbedded with the limestone. Where faulted, the formation is brown to red from hematite staining, and is locally brecciated. Fractures filled with sparry calcite cement are common. Stylolites are rare. Because of abundant sparry calcite cement in pore spaces, porosity is low, and is generally less then 1.0 percent.

Fossil debris composes as much as 50 percent of the rock in some places, and consists of fragments of ostracods, brachiopods, gastropods, bivalves, rugose corals, and crinoids. Where present, the fossils are randomly oriented and randomly distributed. Dendroid tabulate corals are common in layers parallel to the bedding surface. Solitary domal stromatoporoids are present locally. Many fossil fragments have been replaced by sparry calcite cement.

Peloids are common in much of the unit. In some samples, the peloids are poorly sorted, randomly distributed, and are as much as 5.0 cm long. In other samples, the peloids are well sorted, evenly distributed, and average 0.15 cm across. In all samples the peloids have been completely micritized. The larger peloids are intraclasts and, where mixed with fossil debris, the small peloids are probably pellets and intraclasts. In some of the laminated mudstones that contain peloids, a faint cross-lamination is visible on weathered surfaces. These cross-laminated peloids originally may have been ooids. In the Diamond Range, ooids are present in the formation (Nolan and others, 1956, p.50). Fenestral pore spaces that have been filled with sparry calcite cement also are present in some laminated mudstones. Nodular algal mats(?) are present locally. The laminae in these rocks are generally less than 2.0 mm thick.

Texturally, the allochemical constituents form bioclastic wackestones and packstones, peloidal bioclastic wackestones and packstones, intraclastic packstones, and fenestral mudstones. Generally, the matrix material is either dense, brown lime mud or microspar. No burrow structures were found, but the abundance of structureless mudstones suggests that bioturbation was common. Burrows are present in the lower Devils Gate Limestone in the northern Mahogany Hills area (Drake, 1978, p.49).

Poor exposure in the study area makes study of stratigraphic relations within the Devils Gate difficult. Generally, at the base, the formation consists of a few meters of moderate brown to yellowish gray limy dolomites that grade into grayish orange, platy lime mudstones that exhibit what may be ripple marks. Above these limestones are thin- to thick-bedded, gray lime mudstones and intraclastic packstones with interbedded dolomites and limy dolomites that are thin-bedded. Higher in the section, the formation becomes fossiliferous, forms cliffs, and is composed of interbedded lime mudstones, wackestones, and packstones that are laminated to thick-bedded. Dolomites, quartzose dolomites, and a few quartz arenites that resemble the Eureka Quartzite are interbedded locally.

Contacts

The gradational lower contact of the formation with the underlying Bay State Dolomite was described in the previous section. West of Lamoreux Canyon (Plate 2), the upper contact of the formation is a fault. Elsewhere, the upper contact is erosional. West of Lamoreux Canyon and on the west side of the canyon west of South Hill, the formation is unconformably overlain by members of the Pinto Peak Rhyolite. In the Eureka area and at the type section at Devils Gate Pass, the formation is conformably overlain by the Pilot Shale.

Thickness

Exposures of the formation west of Lamoreux Canyon are approximately 200 m to 250 m thick. At Devils Gate Pass, where the base is not exposed, the Devils Gate Limestone is 335 m thick (Drake, 1978, p.33) At Newark Mountain, a complete section of the formation is 366 m thick (Nolan and others, 1956, p.49).

Age

In the thesis area, conodonts from the Devils Gate Limestone indicate a possible Frasnian (early Late Devonian) age. In the Eureka and Antelope Valley areas, the Devils Gate Limestone is late Middle Devonian and Late Devonian in age (Nolan and others, 1956, p.50; Merriam, 1963, p.54). In the southern Mahogany Hills, the base of the formation is late Givetian to early Frasnian in age (Schalla, 1978, p.75) (Plate 1).

Pinto Peak Rhyolite

A lower Oligocene rhyolite assemblage exposed at Pinto Peak in the Pinto Summit quadrangle (Fig. 2) was named the Pinto Peak Rhyolite by Iddings (in Hague, 1892, p.374-379) and the name was adopted for formal usage by Nolan and others (1974). Nolan and others recognized six units in the assemblage, two of which they formally named as members. Three of the units are present in the study area: the Pinto Basin Tuff Member, the Sierra Spring Tuff Member, and an intrusive rhyolite dome.

Pinto Basin Tuff Member

The Pinto Basin Tuff is the most widespread member of the Pinto Peak Rhyolite. The tuffs probably once covered an area of more than 1600 square km (Nolan and others, 1974), and were erupted from vents that extend from Target Hill (Fig. 3) in the Eureka quadrangle south to Grey Fox Peak in the Pinto Summit quadrangle. Nolan and others (1974) noted local unconformities and intra-unit minor faulting, and suggested that the tuff was erupted over a considerable period of time. Blake and others (1975) estimated a maximum thickness of about 150 m.

In the study area, the unit consists of air-fall vitric-crystal tuff, lithic-pumice crystal tuff, and tuffaceous sandstone that is generally white to light gray and locally light greenish gray or pink. Biotite crystals commonly compose as much as 4.0 percent of the unit and occur as hexagonal books that average 2.0 mm in diameter. Clear, anhedral quartz phenocrysts compose about 1.0 percent of the unit and are as large as 2.0 mm in diameter. Sanidine generally averages less than 1.0 percent and is in the form of white subhedral crystals as large as 5.0 mm long. Pumice fragments are generally rounded and are 1.0 to a few cm across, but are as large as 30 cm locally. Sand contained within the unit is commonly coarse-grained and moderately sorted and rounded.

The Pinto Basin Tuff is present in the canyon west of South Hill (Plate 2), where it is buried by Tertiary-Quaternary unconsolidated sediments but crops out on the flanks of gullies. The tuff unconformably overlies the Sentinel Mountain-Bay State Dolomites and the Devils Gate Limestone on the hills of the west side of the canyon that is west of South Hill, and the Sentinel Mountain-Bay State Dolomites north and south of South Hill. Northeast of South Hill, in the northeast corner of the map area (Plate 2), the unit is well exposed, and the northernmost exposures appear to lie upon the Windfall Formation and Dunderberg Shale. The tuff also crops out west of summit 6738. Exposed parts of the tuff are at least 30 m thick.

Four K-Ar age determinations of the unit within the Pinto Summit quadrangle gave results of 34.6 to 35.6 million years (McKee and others, 1971, p.36-38, recalculated using new IUGS decay constant).

Sierra Springs Tuff Member

The Sierra Springs Tuff Member is an ash-flow tuff that overlies the Pinto Basin Tuff Member, and has an estimated outcrop area of about 130 square km (Blake and others, 1975). At the type section in Ratto Canyon, located near the northeast corner of the study area (Fig. 3), Nolan and others (1974) recognized three cooling units with a total thickness of about 182 m. The two lower units are similar. Each has a nonwelded base that grades through a highly welded zone, with development of eutaxitic vitrophyre, to a nonwelded top. At the type section, the average phenocryst composition for the middle unit is 13 percent quartz, 3 percent sanidine, 59 percent plagioclase, and 25 percent mafic minerals; these include biotite, hornblende, augite, hypersthene, and magnetite (Nolan and others, 1974). The upper unit is more mafic than the lower ones and is characterized by a black glassy eutexite near the base. The unit is compositionally a quartz latite (Nolan and others, 1974).

Within the field area, the unit can be recognized by its characteristic orange and brown weathering color and the locally well developed eutaxitic texture. Phenocryst abundances are similar to those noted by Nolan and others (1974) at the type section, with abundant (up to 5.0 percent) euhedral biotite books prominent throughout the unit. Sulphur-filled vesicles are common in the more crystal-rich parts of the unit. Exposed parts of the tuff are at least 30 m thick.

The unit is present in the study area in the canyon west of

South Hill, where it overlies the Pinto Basin Tuff. On the west side of Lamoreux Canyon, the unit unconformably overlies th Devils Gate Limestone. The unit is also present north of summit 6738 (Plate 2).

Two K-Ar age determinations of minerals in the Sierra Springs Tuff done by McKee and others (1971, p.35) give an average age of 35.7 million years (recalculated using new IUGS decay constant).

Rhyolite Dome

A rhyolite dome is located approximately 1.0 km northeast of the summit of South Hill, where it intruded the Sierra Springs Tuff Member (Plate 2). The dome is composed of flow-banded porphyritic rhyolite with a cryptocrystalline groundmass. Bands are 1.0 to 5.0 mm thick and are pale red, grayish orange, and grayish orange pink. Phenocrysts average 1.0 mm across and consist of clear, anhedral quartz (less than 1.0 percent), euhedral biotite (1.0 to 2.0 percent), and equant anhedral to subhedral, white to clear sanidine (1.0 to 2.0 percent). This dome is the southernmost of a series of intrusive rhyolite domes and plugs that extends south from target Hill (Nolan and others, 1974). Eight K-Ar age determinations made by McKee and others (1971, p.36-38) on the intrusive rhyolites range from 33.9 to 35.9 million years (recalculated using new IUGS decay constant), suggesting that the intrusions were probably episodic over time, as were the tuffs (Nolan and others, 1974).

Porphyritic Granite

A few small exposures of porphyritic granite are present in the northwest part of the study area. The exposures are located directly north of and on the north and east flanks of summit 7933 of Castle Mountain, 0.6 km northwest of benchmark 7253, and 2.6 km west of Castle Mountain (Plate 2). The granite is poorly exposed; contacts are approximately located. Weathered surfaces of the unit are yellowish to pinkish gray in color. Brownish gray oxide staining is common. Fresh surfaces are pinkish gray to white.

The unit consists of clear anhedral quartz phenocrysts (average 2.0 mm across) and pinkish subhedral feldspar phenocrysts (average 1.5 mm across) in a fine-grained to microcrystalline groundmass of potassium feldspar, plagioclase, and quartz. Biotite is present in minor amounts and is generally subhedral and prismatic in form. Mineral volumetric proportions, visually estimated, are as follows: quartz phenocrysts 5.0 percent; feldspar phenocrysts 6.0 percent; biotite 2.0 percent; groundmass quartz 40 percent; groundmass feldspar (approximately equal(?) proportions of potassium feldspar and plagioclase) 41 percent; sericite 5.0 percent; calcite 1.0 percent.

Most of the quartz phenocrysts have been embayed by the groundmass and are irregular in shape. Feldspar phenocrysts are somewhat rectangular and consist of both(?) potassium feldspar and plagioclase. None of the feldspar phenocrysts studied are zoned. The potassium feldspar phenocrysts are perthitic and albite intergrowths occur as irregular blebs with indistinct boundaries. Carlsbad and albite twinning are common in the plagioclase phenocrysts. Anorthite content of the plagioclase is approximately 60 percent (labradorite).

The groundmass is microgranitic in texture; grains average 0.08 mm across. Quartz occurs as anhedral blebs, and feldspars are subhedral and somewhat tabular. Feldspars have undergone extensive alteration to sericite, making distinction between potassium feldspar and plagioclase difficult. Replacement of the feldspars by calcite is also common.

The granite intrusions in the study area are dikes and offshoots of the granitic intrusion of the Wood Cone Peak area (Hague, 1892, p.123), located 4.0 km southwest of Castle Mountain. The extent of this intrusion is shown on Plate XI of Hague (1892) and on the Eureka County map of Roberts and others (1967). Hague (1892, p.221-229) and Iddings (in Hague, 1892, p.339-345) give a detailed description of the intrusion, which they refer to as a granite porphyry. Dikes directly connected to the intrusion are present just west of the edge of the map area, at the southwest end of the southwest spur of Castle Mountain (Plate 2). At this location, the dikes are a few meters wide and stand several meters above the enclosing Antelope Valley Limestone.

Hague (1892, p.122) stated that exposures of the intrusion extend as far as the Bellevue Peak area, approximately 1.5 km south of the southwest corner of the map area (Fig. 3). There, the unit crops out as a dike between Bellevue and White Cloud Peaks and has domed the strata, presumably because of an underlying mass of crystalline rock. Locally, dikes related to the intrusion contact metamorphosed the country rock (limestone and dolomite) into white marble, but the zone of alteration surrounding dikes is not of great extent (Hague, 1892, p.224).

The age of biotite in the granitic intrusion of Wood Cone Peak was determined to be 34.1 $\pm$1.5 million years (early Oligocene) by Marvin and others (in Marvin and Cole, 1978, sample 45), and is assumed by them to be the age of emplacement. The granite intrusions in the field area are probably of the same age.

Tertiary and Quaternary Unconsolidated Sediments

Unconsolidated sediments in the study area are undifferentiated (Plate 2). The most prominent deposits consist of piedmont gravels that are locally cemented by caliche. Locally, these sediments are deeply dissected by present-day stream channels that expose as much as 40 m of coarse gravel. Other deposits in the area include stream alluvium, undissected piedmont gravels, and slope wash.

STRUCTURE

Introduction

Lithostratigraphic units in the field area are separated into three structural packets: the Ordovician packet, the Lone Mountain Dolomite packet, and the Devonian packet (Fig. 5, Plate 2).

A low-angle normal (denudational) fault, hereafter referred to as the main denudational fault (MDF), has moved the Lone Mountain Dolomite and Devonian packets onto the upper part of the Ordovician packet (Plates 2, 3). This is the fault that Roberts and others (1967) interpreted as a thrust fault that juxtaposed the Eureka Quartzite over Silurian and Devonian rocks, and that Stewart (1980, p.81, Fig. 40) suggested is Mesozoic in age. In most places, the Lone Mountain Dolomite and Devonian packets are separated by another low-angle normal (denudational) fault, hereafter referred to as the LMDDF, that merges with the MDF (Plates 2, 3).

In the canyon west of South Hill and west of the mouth of Lamoreux Canyon, the Lone Mountain Dolomite and Devonian packets, and the LMDDF, are cut by a high-angle normal fault that dips west and merges with the MDF. Stratigraphic offset along this fault is greatest at the center and decreases toward the ends where it merges with the main denudational fault; thus, the fault is shaped like a shovel and will be referred to as the shovel fault (Plates 2, 3).

A tear fault with right lateral displacement has offset the Lone Mountain Dolomite and Devonian packets, the LMDDF, and the shovel fault. In the northeast corner of the area, possible high-angle normal faults have brought the Devonian packet in contact with Antelope Valley Limestone, Eureka Quartzite, and the Upper Cambrian units (Plates 2, 3).

Ordovician Packet

The Ordovician packet forms the lowest structural unit and consists of, in ascending order, the Antelope Valley Limestone, Copenhagen Formation, Eureka Quartzite, and Hanson Creek Formation. Rocks of the Ordovician packet have been deformed into a series of open folds. North and east of Castle Mountain, a southeast-plunging syncline is inferred (Plates 2, 3). High-angle normal faults cut the Ordovician packet and are best exposed west of Castle Mountain and in the Grays Peak area. Structures in the Ordovician packet are confined to that structural level and therefore formed before or during movement along the MDF.

Low-grade metamorphic features were developed in rocks of the Ordovician packet. These include foliation in the Antelope Valley Limestone, metamorphic muscovite in the Eureka Quartzite, and sutured grain contacts in the Eureka Quartzite and Hanson Creek Formation. Paleotemperature estimates based on the conodont alteration index of conodonts in the Ordovician packet indicate that temperatures ranged from 300 to >600^{0} C (R.L. Ethington, written commun., Appendix I). The low-grade metamorphic features listed above are not present in rocks of overlying Lone Mountain Dolomite and Devonian packets, thus metamorphism occurred before or during movement along the MDF.

Main Denudational Fault

The relations of the MDF to the three packets are similar, in a very simple way, to three cards stacked one on top of the other. If the lowest card in the stack is tilted at one end, the two upper cards will slide off the stack and spread out in front of the lowest card in the direction of tilt. The Ordovician packet represents the lowest card in the stack.

In most places, movement along the MDF has juxtaposed the Lone Mountain Dolomite over the Eureka Quartzite; the Hanson Creek Formation was faulted out because it is a thin, relatively easily deformed sequence (Ross and others, 1979) and may have acted as a "gliding surface" between the more rigid Ordovician and Silurian packets. The deformed surface of the Ordovician packet probably controlled or modified both the level of the MDF and the development of structures in the overlying Lone Mountain Dolomite and Devonian packets. Where the Ordovician packet forms a plunging syncline, the MDF cuts up section from the Eureka into the Hanson Creek Formation, thus preserving the Hanson Creek; this is best developed in the area of Dave Kean Spring, southwest of Grays Peak, and in the Castle Mountain area (Plate 2). Northeast of Rock Spring and south and east of Grays Peak, the Ordovician packet dips steeply under the overlying packets. At these localities, the MDF cuts up section in the hanging wall block, merges with the LMDDF, and cuts out the Lone Mountain Dolomite packet and parts of the Devonian packet so that Devonian rocks are in contact with the Ordovician packet (Plates 2, 3).

At Castle Mountain, the MDF is crosscut by dikes of the early Oligocene porphyritic granite.

Lone Mountain Dolomite Packet

The Lone Mountain Dolomite packet is composed mostly of Lone Mountain Dolomite. As mentioned, much of the Lone Mountain Dolomite has been removed by faulting. In most places, unit one (the lowest unit) of the Lone Mountain Dolomite has been cut out by movement along the MDF. There are many high-angle normal faults within the Lone Mountain Dolomite packet that do not continue across the denudational faults; this indicates that the high-angle faults either merge with, or are truncated by, the denudational faults. Southeast of Castle Mountain, movement along several high-angle normal faults has resulted in the preservation of some of the Devonian packet (mostly Beacon Peak Dolomite) within the Lone Mountain Dolomite packet (Plates 2, 3).

Denudational Fault Between Lone Mountain Dolomite and Devonian Packets

Generally, the LMDDF is located near the top of the Lone Mountain packet and at the base of the Devonian packet, but in many places is located within the Devonian packet. Southeast of Dave Kean Spring and southeast of Grays Peak, the LMDDF merges with the MDF. Northeast of summit 7155 and south of benchmark 7253 the LMDDF splays and the uppermost splay is in the lowest part of the Sentinel Mountain-Bay State Dolomites. The LMDDF is displaced by the tear fault west of summit 7155 and south of benchmark 7253. On the west side of South Hill, in the footwall block of the shovel fault, the denudational fault lies between the Lone Mountain Dolomite and the Beacon Peak Dolomite, but south of South Hill it has cut up section in the Devonian and placed the Sentinel Mountain-Bay State Dolomites on top of Beacon Peak and Bartine (sections E-E', and F-F', Plate 3). This sequence is offset along the tear fault and is resumed on the west side of the mouth of Lamoreux Canyon (in the footwall block of the shovel fault) where the upper part of the Devonian packet and the fault have been eroded away.

Devonian Packet

The Devonian packet is composed of, in ascending order, the Beacon Peak Dolomite, McColley Canyon Formation (mostly Bartine Member), Sadler Ranch Formation, Oxyoke Canyon Sandstone, Sentinel Mountain-Bay State Dolomites undivided, Devils Gate Limestone, and members of the Pinto Peak Rhyolite. As mentioned, the LMDDF rises into the Devonian packet locally and splays further disect the Devonian packet. Specifically, these splays place the Oxyoke Canyon Sandstone against the Bartine and the Sadler Ranch Formation against the Beacon Peak Dolomite west of Dave Kean Spring, southwest of benchmark 7253, in the summit 7155 area, and in the Grays Canyon area (Plate 2). The Devonian packet is disrupted by high-angle normal faults. Southwest of Grays peak, these faults cut into the Lone Mountain Dolomite packet and displace the LMDDF and other earlier formed normal faults within the Devonian packet. The abundance of low-angle normal fault splays in the Devonian packet compared to the other packets is perhaps(?) because of the abundance of thin units within the lower parts of the Devonian packet. The thinner units form less rigid layers of strata that are more easily deformed compared to the more rigid layers above (Sentinel Mountain-Bay State Dolomites and Devils Gate Limestone) and below (Lone Mountain Dolomite).

Shovel Fault

The main trace of the shovel fault is in the canyon west of South Hill where it juxtaposes volcanics of the Devonian packet in the hanging wall on the west and the Lone Mountain Dolomite and Devonian packet in the foot wall on the east (Plates 2, 3). The LMDDF is displaced by the shovel fault. The shovel fault is also present at the west side of the mouth of Lamoreux Canyon where it juxtaposes the Devils Gate Limestone in the hanging wall on the west with Silurian and Devonian rocks in the footwall on the east. The shovel fault merges with the MDF at the north end of the canyon west of South Hill and near benchmark 6852 (Plate 2).

Tear Fault

The tear fault in the Lone Mountain Dolomite and Devonian packets displaces structures and lithologic units in the two packets, including the shovel fault, by approximately 3.25 km of right lateral offset, and therefore formed after them. The trace of the tear fault trends north-northwest in lower Lamoreux Canyon and trends northwest along the gully between the Lone Mountain Dolomite and Ordovician packets east of Castle Mountain, where it merges with the MDF (Plate 2, 3). This nearly linear trace suggests that the fault has a steep dip. Drag along the tear fault folded rocks of the Lone Mountain Dolomite and Devonian packets where they merge with the fault; north and east of summit 7155 the two packets are folded into a southeast-plunging anticline and west of Lamoreux Canyon, south and east of benchmark 7253, they are folded into a southeast plunging-syncline.

Formation of the tear fault in the Lone Mountain Dolomite and Devonian packets was probably controlled by the preexisting structure of the underlying Ordovician packet. West of Castle Mountain and near benchmark 7253, the Ordovician packet is folded into a southeast-plunging anticline with a steeply dipping west limb. Movement of the Lone Mountain Dolomite and Devonian packets over the Ordovician packet along the MDF probably produced the tear in the upper two packets because of the relative change in elevation of the top of the Ordovician packet on either side of the steeply dipping limb of the anticline.

Discussion

Regional considerations

The eastern edge of the Roberts Mountains allochthon may have been located either in the study area or just to the west. Rocks of the allochthon are present 16 km northwest of the study area in the northern Mahogany Hills-Devils Gate area (Drake, 1978, p.81), and Mississippian strata in the southern part of the Fish Creek Range, 24 km south of the study area, contain Devonian blocks that were detached from the front of the allochthon and slid eastward into the foreland trough (Sans, 1986). There is no evidence in the study area for Late Devonian-earliest Mississippian age Antler orogenic activity. Likewise, there is no evidence for Late Permian-Early Triassic Sonoman orogenic activity.

In the Eureka mining district, northeast of the study area (Fig. 3), there are quartz diorite plugs that are late Early Cretaceous (102 $\pm$3 m.y.) in age (Nolan, 1962, p.2, 13; Nolan and others, 1974; Marvin and Cole, 1978, p.9). Data from mines and drill holes indicate that these intrusions are much more extensive at depth than the surface exposures indicate and that they are probably part of a larger intrusive mass that underlies the mining district (Nolan, 1962, p.2, 13). Outcrops of granitoid intrusive units are present just north of the study area (Fig. 3) at McCulloughs Butte and on the flanks of summit 8491 (J.M. Bird, personal commun., 1986). The intrusion exposed at McColloughs Butte is Late Cretaceous in age (Barton, 1986). Southeast of the study area, in the Pancake Range, an Early Cretaceous (110 $\pm$5 m.y.) intrusive dacite is present (Nolan and others, 1974; Marvin and Cole, 1978, p.9). Northwest of the study area, intrusive rocks of Late Jurassic age occur at Whistler Mountain (Armstrong, 1963, p.163). An Early Cretaceous intrusion is suspected to underlie the Roberts Mountains (Winterer, 1968), and ones of Late Jurassic or Early Cretaceous age are known in the Cortez Mountains (Armstrong, 1963, p.165; Gilluly and Masursky, 1965, p.71). In light of the presence of these intrusive masses northwest, north, northeast, and southeast of the study area, it is probable that the northern part of the Fish Creek Range is underlain by a Cretaceous intrusive mass at depth.

Northeast of the study area, the Cambrian, Ordovician, Mississippian, Pennsylvanian, and Permian sections in the northwest corner of the Pinto Summit quadrangle (Fig. 2) have undergone extensive compressional deformation that resulted in folding and faulting (Nolan, 1962, p.1, 28; Nolan and others, 1974). In places, they are unconformably overlain by the Early Cretaceous Newark Canyon Formation. The Silurian and Devonian sections are not present in this part of the Pinto Summit quadrangle. Nolan divided the quadrangle into a series of north-trending linear structural blocks (antiforms and synforms). The following field relations suggest that most of this deformation occurred in the Late Mesozoic:

Nolan, 1962, p.28:

> The sedimentary rocks of the Newark Canyon formation are clearly younger than all the faults of this episode, but the lithologic character of the sediments - coarse poorly sorted lenticularly bedded conglomerates alternating with fresh-water limestones and siltstones - and the surface of considerable relief on which they were deposited combine to suggest that deposition of the formation occurred during a period of crustal instability. This, combined with the probable Cretaceous age of the intrusive quartz diorite, leads to the speculation that the folding and thrusting at Eureka may have occurred just prior the deposition of the Newark Canyon formation and that its deformation and the emplacement of the igneous mass are the final stages in an episode of deformation that was most intensive at the beginning of Early Cretaceous time.

Nolan and others, 1974, p.11:

> The formation of the structural blocks appears to date from Cretaceous time; outcrops of the Newark Canyon Formation are to a large extent localized within the synforms and the lithology of the unit indicates that it was largely deposited in depressions that received their debris from the adjoining antiforms. The megabreccias exposed in the Eureka quadrangle to the north (Nolan and others, 1971, p.3-4) also point to this conslusion.

Causes and timing of deformation in the study area

Lithostratigraphic and biostratigraphic studies of the Paleozoic section exposed in the study area indicate the section is in place relative to surrounding areas; therefore, strata of the three structural packets are locally derived. Based upon the evidence presented, the following hypothesis is proposed for the development of the study area: Folding in the Ordovician packet probably occurred during the Cretaceous. During this time, central Nevada was in the hinterland of the Sevier orogenic belt and was undergoing deformation, plutonism, and regional metamorphism (Coney, 1980). The igneous mass that presumably underlies the range was emplaced during this orogenic episode. Heat generated by plutonism caused the low-grade metamorphism in the Ordovician packet.

The Lone Mountain Dolomite and Devonian packets were probably detached from the Ordovician packet and not folded. This is substantiated by the lack of heating in the packets and the presence of lower Oligocene volcanics overlying the Devonian packet at approximately the same stratigraphic level throughout the study area. Heat from the pluton may have been transferred to the country rocks by migrating pore water. The thick, massive Lone Mountain Dolomite possibly provided a permeability barrier to the migrating pore water, thus sparing the overlying packets from heating. Also, this permeability barrier may have allowed increased pore-water pressure at the Lone Mountain - Hanson Creek interface. Uplift of the area probably accompanied deformation and the emplacement of the igneous body. This, and the possible increased pore-water pressure at the boundary between the Lone Mountain Dolomite and Ordovician packets, may have promoted initial gravity sliding of the Lone Mountain Dolomite packet off of the Ordovician packet. Alternatively, the Lone Mountain Dolomite and Devonian packets may have resided in a nearby area (perhaps in the northwest part of the Pinto Summit quadrangle) that was not affected by the low-grade metamorphism. High-angle normal faulting in the Ordovician packet, and possibly the overlying packets, may have occurred during the uplift because the faults predate final MDF movement.

The Upper Devonian through Permian(?) section was probably removed during the uplift. Because the upper surface of the preserved packets (Devonian packet) was at a somewhat uniform stratigraphic level (Middle and Upper Devonian), the overlying Upper Devonian through Permian(?) section may have been removed by gravity sliding, although there is no evidence to substantiate this. It is probable that the Early Cretaceous Newark Canyon Formation was not deposited in the study area because the area was then an upland source for Newark Canyon detritus. The Newark Canyon Formation is present south of the study area in the southern part of the Fish Creek Range (Sans, 1986, p.42), northwest in the Mahogany Hills (Schalla, 1978, p.94), and to the northeast and east in the Pinto Summit quadrangle (Nolan and others, 1974).

Removal of the Upper Devonian through Permian(?) strata from the study area was completed by early Oligocene because the lower Oligocene volcanics unconformably overlie the Middle and Upper Devonian strata in the study area. The presence of these volcanics indicates that igneous activity in the region was renewed or continued into the Oligocene. Uplift probably continued and perhaps further deformed the Ordovician packet. The uplift probably caused the sequential development of the following features within a relatively short period during the early Oligocene: (1) continued movement along the MDF; (2) development of the LMDDF (and splays); (3) development of the shovel fault; and finally (4) development of and movement along the tear fault. At least some of the high-angle normal faults in the Lone Mountain Dolomite and Devonian packets probably developed during this episode of movement along the denudational faults.

Movement between the Lone Mountain Dolomite and Devonian packets did not occur before the early Oligocene because the lower Oligocene volcanics are not present on the Lone Mountain Dolomite packet. Also, most movement of the Lone Mountain Dolomite and Devonian packets over the Ordovician packet occurred along the MDF because large amounts of section were removed along this fault relative to the small amounts of section lost along the other denudational faults. This episode of movement and final emplacement of the Lone Mountain Dolomite and Devonian packets is restricted to the early Oligocene because lower Oligocene volcanics were displaced by this movement and granite of early Oligocene age (younger than the volcanics) crosscuts the MDF along which movement occurred.

Widespread plutonic-volcanic activity of magmatic arc affinity affected Nevada during the mid-Tertiary (Coney, 1980). This is also when much of the low-angle normal faulting that affected east-central Nevada occurred (Armstrong, 1972; Coney, 1980), and Nevada as a whole was undergoing extension (Zoback and others, 1981).

The Upper Cambrian units exposed in the northeastern corner of the study area are overlain by lower Oligocene volcanics, which indicates that the Cambrian units were exposed prior to the Oligocene. Presently, the Cambrian units are separated from the Devonian packet by a possible high-angle normal fault that bifurcates to separate the Cambrian rocks on the east from Ordovician rocks that are separated from Devonian rocks on the west (Plate 2). These faults continue to the north where they show the same sense of displacement and have an nearly linear trace that suggests a steep westward dip (J.M. Bird, personal commun., 1986). Because the Devonian rocks were not emplaced in their present positions until the early Oligocene, and because the Cambrian section is in contact with Devonian and Ordovician rocks along high-angle normal faults, movement along these normal faults is probably post-early Oligocene and related to basin and range faulting. Nolan and others (1974) noted that recent block faulting related to basin and range extension in the Eureka area is at least as young as post-Oligocene because these block faults cut Oligocene volcanics. Recent faulting in the region has also produced well-marked fault scarps in the piedmont gravels.

REFERENCES

Armstrong, R.L., 1963, Geochronology and geology of the eastern Great Basin [Ph.D. thesis]: New Haven, Connecticut, Yale University, 202p.

——1968, Sevier orogenic belt in Nevada and Utah: Geological Society of America Bulletin, v.79, p.429-458.

——1972, Low-angle (denudational) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah: Geological Society of America Bulletin, v.83, p.1729-1754.

Barton, M.D., 1986, Lithophile element mineralization associated with Late Cretaceous two-mica granites in Nevada and California: Geological Society of America Abstracts with Programs, p.84.

Berry, W.B.N., 1977, Some Siluro-Devonian biofacies patterns in the western United States in Stewart, J.G., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic Paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 1, p.217-240.

Blake, M.C., Jr., McKee, E.H., Marvin, R.F., Silberman, M.L., and Nolan, T.B., 1975, The Oligocene volcanic center at Eureka, Nevada: Journal of Research, U.S. Geological Survey, v.3, p.605-612.

Burchfiel, B.C., and Davis, G.A., 1975, Nature and controls of Cordilleran orogenesis, western United States: extensions of an earlier syntheses; American Journal of Science, v.275-A, p.363-396.

Carlisle, Donald, Murphy, M.A., Nelson, C.A., and Winterer, E.L., 1957, Devonian stratigraphy of Sulphur Springs and Pinyon Ranges, Nevada; American Association of Petroleum Geologists, v.41, p.2175-2191.

Colman, R.L., 1979, The carbonate petrology and conodont biostratigraphy of the Old Whalen Member of the Lone Mountain Dolomite (Lower Devonian), Sulphur Springs Range, Nevada [M.S. thesis]: Riverside California, University of California, 116p.

Coney, P.J., 1978, Mesozoic-Cenozoic Cordilleran plate tectonics in Smith, R.B., and Eaton, G.P., eds., Cenozoic tectonics and regional geophysics of the western Cordillera: Geological Society of America Memoir 152, p.33-50.

——1980, Cordilleran metamorphic core complexes: an overview in Crittenden, M.D., Coney, P.J., and Davis, G.H., eds., Cordilleran metamorphic core complexes: Geological Society of America Memoir 153, p.7-34.

Dott, R.H., Jr., 1964, Wacke, graywacke and matrix - what approach to immature sandstone classification?: Journal of Sedimentary Petrology, V.34, p.625-632.

Drake, E.A., 1978, Paleozoic stratigraphy of the Devils Gate - northern Mahogany Hills area, Eureka County, Nevada [M.S. thesis]: Corvallis, Oregon, Oregon State University, 110p.

Dunham, J.B., 1977, Depositional environments and paleogeography of the Upper Ordovician, Lower Silurian carbonate platform of central Nevada in Stewart, J.G., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic Paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 1, p.157-164.

Dunham, R.H., 1962, Classification of carbonate rocks according to depositional texture in Ham, W.E., ed., Classification of carbonate rocks, a symposium: American Association of Petroleum Geologists Memoir 1, p.108-121.

Folk, R.L., 1962, Spectral subdivisions of limestone types in Ham, W.E., ed., Classification of carbonate rocks, a symposium: American Association of Petroleum Geologists Memoir 1, p.62-84.

Gilluly, J., and Masursky, H., 1965, Geology of the Cortez quadrangle, Nevada: U.S. Geological Survey Bulletin 1175, 117p.

Hague, Arnold, 1883, Abstract of report on the geology of the Eureka district, Nevada: U.S. Geological Survey 3d Annual Report, p.237-272.

——1892, Geology of the Eureka district, Nevada: U.S. Geological Survey Monograph 20, 419p., with an atlas of 13 sheets.

Hintze, L.F., 1973, Geologic history of Utah: Brigham Young University Geology Studies, v.20, p.1-181.

Hose, R.K., Armstrong, A.K., Harris, A.G., and Mamet, B.L., 1982, Devonian and Mississippian rocks of the northern Antelope Range, Eureka County, Nevada: U.S. Geological Survey Professional Paper 1182, 19p.

Johnson, D.B., 1978, Analysis of Lower Devonian conodont ecology, Eureka County, Nevada [Ph.D. thesis]: Iowa City, University of Iowa, 186p.

Johnson, J.G., 1962, Lower Devonian - Middle Devonian boundary in central Nevada: American Association of Petroleum Geologists Bulletin, v.46, p.542-546.

——1965, Lower Devonian stratigraphy and correlation, northern Simpson Park Range, Nevada: Bulletin of Canadian Petroleum Geology, v.13, p.365-381.

——1977, Lower and Middle Devonian Faunal Intervals in central Nevada, based on brachiopods in Murphy, M.A., Berry, W.B.N., and Sandberg, C.A., eds., Western North America Devonian: University of California, Riverside, Campus Museum Contributions 4, p.16-32.

Johnson, J.G., and Lane, N.G., 1969, Two new Devonian crinoids from central Nevada: Journal of Paleontology, v.43, p.69-73.

Johnson, J.G., and Murphy, M.A., 1984, Time-rock model for Siluro-Devonian continental shelf, western United States: Geological Society of America Bulletin, v.95, p.1349-1359.

Johnson, J.G., and Pendergast, Anne, 1981, Timing and mode of emplacement of the Roberts Mountains allochthon, Antler orogeny: Geological Society of America Bulletin, v.92, p.648-658.

Johnson, J.G., and Sandberg, C.A., 1977, Lower and Middle Devonian continental-shelf rocks of the western United States, in Murphy, M.A., Berry, W.B.N., and Sandberg, C.A., eds., Western North America Devonian: Riverside California, University of California, Campus Museum Contribution 4, p.121-143.

Johnson, J.G., Boucot, A.J., and Murphy, M.A., 1973, Pridolian and early Gedinnian age brachiopods from the Roberts Mountains Formation of central Nevada: University of California Publications in Geological Sciences, v.100, 75p.

Johnson, J.G., Klapper, G., and Trojan, W.R., 1980, Brachiopod and conodont successions in the Devonian of the northern Antelope Range, central Nevada: Geologica et Palaeontologica, v.14, p.77-116.

Johnson, J.G., Klapper, G., and Sandberg, C.A., 1985, Devonian eustatic fluctuations in Euramerica: Geological Society of America Bulletin, v.96, p.567-587.

Kendall, G.W., 1975, Some aspects of Lower and Middle Devonian stratigraphy in Eureka County, Nevada [M.S. thesis]: Corvallis, Oregon, Oregon State University, 199p.

Kendall, G.W., Johnson, J.G., Brown, J.O., and Klapper, Gilbert, 1983, Stratigraphy and facies across Lower Devonian - Middle Devonian boundary, central Nevada: American Association of Petroleum Geologists Bulletin, v.56, p.503-527.

King, Clarence, 1878, Systematic geology: U.S. Geological Exploration 40th parallel, v.1.

Kirk, Edwin, 1933, The Eureka Quartzite of the Great Basin region: American Journal of Science, 5th Series, v.26, p.27-44.

Long, J.F., 1973, Stratigraphy and depositional environments of shoal water carbonate rocks in the Fish Creek Range, central Nevada [M.S. thesis]: Riverside California, University of California, 151p.

Marvin, R.F., and Cole, J.C., 1978, Radiometric ages: compilation A, U.S. Geological Survey: Isochron/West, no. 22, p.9

McKee, E.H., Silberman, M.L., Marvin, R.F., and Obradovich, J.D., 1971, A summery of radiometric ages of Tertiary rocks in Nevada and eastern California, Part I, central Nevada: Isochron/West, no.71-2, p.21-42.

McKee, E.P., and Weir, G.W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geological Society of America Bulletin, v.64, p.383.

Merriam, C.W., 1940, Devonian stratigraphy and paleontology of the Roberts Mountains region, Nevada: Geological Society of America Special Paper 25, 114p.

——1963, Paleozoic rocks of Antelope Valley, Eureka and Nye Counties, Nevada: U.S. Geological Survey Professional Paper 423, 67p.

——1973, Paleontology and stratigraphy, Rabbit Hill Limestone, Lone Mountain Dolomite, Nevada: U.S. Geological Survey Professional Paper 808, 50p.

Merriam, C.W., and Anderson, C.A., 1942, Reconnaissance survey of the Roberts Mountains region, Nevada: Geological Society of America Bulletin, v.53, p.1675-1728.

Mullens, T.E., and Poole, F.G., 1972, Quartz-sand bearing zone and early Silurian age of the upper part of the Hanson Creek Formation in Eureka County, Nevada: U.S. Geological Survey Professional Paper 800-B, p.21-24.

Murphy, M.A., and Gronberg, E.C., 1970, stratigraphy and correlation of the lower Nevada Group (Devonian) north and west of Eureka, Nevada: Geological Society of America Bulletin, v.81, p.127-136.

Murphy, M.A., McKee, E.H., Winterer, E.L., Matti, J.C., and Dunham, J.B., 1978, Preliminary geologic map of the Roberts Creek Mountain quadrangle, Nevada: U.S. Geological Survey Open-File Report 78-736.

Murphy, M.A., Dunham, John, Berry, W.B.N., and Matti, J.C., 1979, Late Llandovery unconformity in central Nevada: Brigham Young University Geology Studies, v.26, pt.1, p.21-36.

Nolan, T.B., 1962, The Eureka mining district, Nevada: U.S. Geological Survey Professional Paper 406, 78p.

Nolan, T.B., Merriam, C.W., and Williams, J.S., 1956, The stratigraphic section in the vicinity of Eureka, Nevada: U.S. Geological Survey Professional Paper 276, 77p.

Nolan, T.B., Merriam, C.W., and Brew, D.A., 1974, Geologic map of the Eureka quadrangle, Eureka and White Pine Counties, Nevada: U.S. Geological Survey Map I-612.

Nolan, T.B., Merriam, C.W., and Blake, M.C., Jr., 1974, Geologic map of the Pinto Summit quadrangle, Eureka and White Pine Counties, Nevada: U.G. Geological Survey Map I-793.

Palmer, A.R., 1960, Identification of the Dunderberg Shale of Late Cambrian age in the eastern Great Basin: U.S. Geological Survey Professional Paper 400-B, p.B289-B290.

Pettijohn, F.J., Potter, P.E., and Siever, Raymond, 1973, Sand and sandstone: New York, Springer Verlag, 618p.

Poole, F.G., 1974, Flysch deposits of the Antler foreland basin, western United States: Society of Econimic Paleontologists and Mineralogists Special Publication 22, p.58-82.

Potter, E.C., 1976, Paleozoic stratigraphy of the northern Hot Creek Range, Nye County, Nevada [M.S. thesis]: Corvallis, Oregon, Oregon State University, 129p.

Roberts, R.J., 1972, Evolution of the Cordilleran fold belt: Geological Society of America Bulletin, v.83, p.1489-2004.

Roberts, R.J., Preston, E.H., Gilluly, J., and Ferguson, H.G., 1958, Paleozoic rocks of north - central Nevada: American Association of Petroleum Geologists Bulletin, v.42, p.2813-2857.

Roberts, R.J., Montgomery, K.M., and Lehner, R.E., 1967, Geology and mineral resources of Eureka County, Nevada: Nevada Bureau of Mines Bulletin 64, 152p.

Ross, R.J., 1970, Ordovician brachiopods, trilobites, and stratigraphy in eastern and central Nevada: U.S. Geological Survey Professional Paper 639, 103p.

Ross, R.J., Nolan, T.B., and Harris, A.G., 1979, The Upper Ordovician and Silurian Hanson Creek Formation of central Nevada: U.S. Geological Survey Professional Paper 1126-C, 22p.

Sans, R.S., 1986, Origin of Devonian rock units in the southern Fish Creek Range, Nye County, Nevada [M.S. thesis]: Corvallis, Oregon, Oregon State University, 69p.

Schalla, R.A., 1978, Paleozoic stratigraphy of the southern Mahogany Hills, Eureka County, Nevada [M.S. thesis]: Corvallis, Oregon, Oregon State University, 118p.

Stewart, J.H., 1980, Geology of Nevada: Nevada Bureau of Mines and Geology Special Publication 4, 136p.

Stewart, J.H., and Poole, F.G., 1974, Lower Paleozoic and uppermost Precambrian Cordilleran miogeocline, Great Basin, western United States: Society of Economic Paleontologists and Mineralogists Special Publication 22, p.28-57.

Walcott, C.D., 1884, Paleontology of the Eureka district: U.S. Geological Survey Monograph 8, 297p.

Webb, G.W., 1958, Middle Ordovician stratigraphy in eastern Nevada and Western Utah: American Association of Petroleum Geologists Bulletin, v.42, p.2335-2377.

Winterer, E.L., 1968, Tectonic erosion in the Roberts Mountains, Nevada: Journal of Geology, v.76, p.347-357.

Zoback, M.L., Anderson, R.E., and Thompson, G.A., 1981, Cainozoic evolution of the state of stress and style of tectonism of the Basin and Range province of the western United States: Philosophical Transactions of the Royal Society of London A-300, p.407-434.

APPENDICES

APPENDIX I

Faunal Lists

Ordovician conodont collections were identified by R.L. Ethington, 1985. Silurian and Devonian conodont collections were identified by Gilbert Klapper, 1985, 1986. Brachiopods and corals were identified by J.G. Johnson, 1986. Gastropods were identified by Robert Blodgett, 1986. Collection localities are indicated on Plate 2. The letters oc following the sample number indicate that the sample was collected from outcrop, fl indicates it was collected from float.

Sample 2 oc: Hanson Creek Formation — Conodonta:

CAI is 6, ca. 500° C.
Three specimens, all fragmented, representing a species of *Oulodus* or of *Plectodina* - Sample is post-Whiterockian and probably at least as young as Shermanian; *Plectodina tenuis*, the most likely species, ranges from the Shermanian through to the top of the Ordovician.

Sample 10 oc: Hanson Creek Formation — Conodonta:

CAI is 5.5, ca. 450° C.
Sample is dominated by a species of *Panderodus*, a genus whose elements show a vexing degree of variation. It will require careful examination of each specimen to attempt identification; I am not confident that these specimens, all of which are broken, are adequate. A fragment of a belodinid is inadequate for generic identification, but seems to conform to morphologes that are known best from Sweet's work on the Bighorn Group. Two other fragments may be basal parts of a species of *Culumbodina* but that is a very "iffy" diagnosis. On balance, this seems to be an Upper Ordovician collection.

Sample 14 oc: Antelope Valley Limestone — Conodonta:

These have really been cooked, many having reached the colorless, glassy stage. CAI is 8, more than 600° C, probably much more. Most are generalized simple cones, and not diagnostic, given the state of preservation. I can identify elements of *Scolopodus gracilis* Ethington and Clark, *Scandodus sinuosus* Mound, and *Multiostodus auritus* (Harris and Harris), which indicate Whiterockian age. If another specimen is an element of *Paraprioniodus costatus* (Mound), I can limit it to upper Whiterockian (preChazy) but that again is very "iffy".

<u>Sample 17 oc: Antelope Valley Limestone Conodonta:</u>

These specimens are quite thermally altered and also are quite distorted.

I have not been able to identify any of the specimens in a cursory examination, but a more thorough study may change that. These are not Upper Ordovician forms and they are too advanced for the Lower Ordovician. They are Chazyan or younger Cincinnatian. Several elements represent either <u>Phragmodus</u> or <u>Periodon</u> and, if I can establish which, they should tie this down.

<u>Sample 20 fl: Devils Gate Limestone Gastropoda:</u>

Middle to Upper Devonian

c.f. <u>Scalitina</u> sp.

<u>Sample 23 fl: Antelope Valley Limestone Conodonta:</u>

Conodonts have lost much of their color so that they have a CAI of the order of 6.5, i.e. exposed to temperatures between 400 and 450° C. The specimens are very fragmentary, so that identifications are very tentative. These conodonts are no older than late Whiterockian. If the severely broken specimen identified as <u>C. rigbyi</u> is identified correctly, the fauna is of late Whiterockian age; if that identification is not correct, these conodonts could be as young as the lower part of the Copenhagen (Chazyan; I use Whiterockian in the classic sense as older than Chazyan rather than in the sense of Rube Ross who included the Chazyan in the Whiterockian).

<u>Ansella sp</u>.

?<u>Belodina monitorensis</u> Ethington and Schumacher

?<u>Chosonodina rigbyi</u> Ethington and Clark

<u>Drepanoistodus</u> sp.

<u>Panderodus</u> sp.

<u>Protopanderodus</u> sp.

<u>Sample 26 oc: Antelope Valley Limestone Conodonta:</u>

CAI is 4.5, i.e., paleotemperatures up to 250° C.

The fauna is meager but the conodonts are well preserved so that age is unequivocally late Whiterockian. These conodonts occur in upper Antelope Valley in the Monitor Range and in the Lehman Formation of western Utah.

<u>Erraticodon balticus</u> Dzik

<u>Paraprioniodus costatus</u> (Mound)

<u>Scandodus sinuosus</u> Mound

<u>Sample 29 fl: McColley Canyon Fm., Bartine Member Brachiopoda:</u>

Interval 12-13 (Johnson, 1977; Johnson and others, 1980)

<u>Brachyspirifer pinyonoides</u>

<u>Chonetes</u> sp.

Sample 30 fl: McColley Canyon Fm., Bartine Member Brachiopoda:

Interval 12-13 (Johnson, 1977; Johnson and others, 1980)
Brachyspirifer pinyonoides
Schizophoria sp.

Sample 34 oc: Lone Mountain Dolomite Conodonta:

CAI is 2.
Silurian, amorphognathoides Zone, late Llandovery-early Wenlockian
Pterospathodus amorphognathoides Walliser
P. pennatus (Walliser)
Panderodus sp.

Sample 35 oc: Hanson Creek Formation Conodonta:

CAI 6+; maximum paleotemperatures in excess of 400° C.
The specimens are quite worn, perhaps mechanically prior to sedimentation and consolidation of the rock from which they were obtained. I can only hazard a guess as to the age of these forms because they are so poorly preserved. The most important forms are those identified with Pseudobelodina; I believe this is a reasonable although not necessarily reliable assignment. If it is correct the fauna is of Late Ordovician age, i.e. from Hanson Creek or its equivalent. If, however, these are specimens of Belodina, the sample could be from rocks as old as Copenhagen.
Panderodus aff. P. feulneri (Glenister)
?Pseudobelodina inclinata (Branson and Mehl)
?angulate pectiniform elements--perhaps from a species of Oulodus, Plectodina, or Bryantodina

Sample 37 fl: McColley Canyon Fm., Bartine Member Conodonta:

CAI is 2.
Lower Devonian, dehiscens to serotinus Zones
Icriodus trojani Johnson and Klapper
I. nevadensis? Johnson and Klapper
Pandorinellina sp. indet.
Belodella sp.
Panderodus sp.

Sample 40 oc: Sadler Ranch Formation Conodonta:

CAI is 2.
Lower-Middle Devonian, in the range from the serotinus Zone to the costatus Zone (Klapper and Johnson, 1980, tables 6, 7)
Pandorinellina expansa Uyeno and Mason
Polygnathus serotinus Telford
Panderodus sp.

__Sample 52 oc: Copenhagen Formation Conodonta:__
CAI is 5+; maximum paleotemperature above 300° C.
Specimens are broken, showing strong surface etching, and commonly have silt adhering to their surfaces. The dominant species, P.aculeata, ranges from the Black Riveran through the Kirkfieldian so that this collection falls within the stratigraphic interval of the upper Copenhagen.

Panderodus sp.
Plectodina aculeata (Stauffer)

__Sample 57 oc: Antelope Valley Limestone Conodonta:__
CAI approaching 8, many specimens are colorless and glassy; paleotemperatures of as much as 600° C which seems high unless the sample was collected adjacent to an intrusive body or just beneath a welded tuff. In addition, the specimens are quite deformed, I suppose tectonically. As a result the species are hard to decipher. Collectively they suggest lower Copenhagen, above the sandstone member.

Ansella sp.
Panderodus sp.
Phragmodus sp.

__Sample 58 oc: Antelope Valley Limestone Conodonta:__
CAI approaching 8.
This sample is latest Whiterockian or oldest Chazyan, depending on how that boundary eventually is recognized. This fauna is very like that in uppermost Antelope Valley on Martin Ridge in the Monitor Range, but these forms easily could range up into the basal sandstone member of the Copenhagen from which no conodonts have been recovered. The fauna occurs in the Crystal Peak Dolomite in western Utah.

Ansella sp.
Oulodus aff. O. serratus (Stauffer)
Phragmodus "preflexuosus" sensu Sweet, 1984 (= P. flexuosus of Ethington and Clark, 1982)
Panderodus sp.

__Sample 59 oc: Antelope Valley Limestone Conodonta:__
CAI is 5; maximum paleotemperature around 300° C.
This is a young Whiterockian fauna. I did not find the species of Histiodella and of Juanognathus in the Lehman Formation in Utah but their stratigraphic position there might be in the overlying Watson Ranch Quartzite. They are present in upper Antelope Valley Limestone at Ikes Canyon in the Toquima Range.

Histiodella n. sp. 2 of Harris et al., 1979
Juanognathus aff. J. variabilis Serpagli of Harris et al., 1979
?Oistofus sp.
Scandodus sinuosus Mound

Sample 205 oc: Hanson Creek Formation Conodonta:

CAI is 7+.

Many specimens have quartz grains adhering to their surfaces. The population is dominated by _Panderodus_ which is not diagnostic as to age in our present understanding of that vexing genus. Two elements represent species of _Plectodina_ and of _Phragmodus_ respectively. If they do represent the species with which I have compared them, this sample is from rocks equivalent to some part of the Hanson Creek. If those identifications are not reliable, the sample could be from rocks as old as the Copenhagen.

Panderodus sp.
?_Phragmodus_ _undatus_ Branson and Mehl--S element lacking posterior process
?_Plectodina_ _florida_ Sweet--Sbb element

Sample 211A oc: Oxyoke Canyon Sandstone Conodonta:

CAI is 2.
Probably Middle Devonian
Polygnathus sp. indet.

Sample 217 oc: Sadler Ranch Formation Conodonta:

CAI is 2.
Lower-Middle Devonian
Pandorinellina sp. indet.
Panderodus sp.
Pseudooneotodus sp.

Sample 218 fl: McColley Canyon Fm., Bartine Member Conodonta:

CAI is 2.
Latest Silurian-Lower Devonian
Icriodus sp. indet.
Panderodus sp.

Sample 220 oc: Sadler Ranch Fm., upper dolomite Conodonta:

CAI is 2.
Probably Lower Devonian, _serotinus_ to _patulus_ Zones (Klapper and Johnson, 1980, tables 6, 7)
Pandorinellina _expansa_
Polygnathus _n_. sp. B? of Klapper
_P_. _linguiformis_ _bultyncki_ Weddige

Sample 221 oc: Sadler Ranch Fm., middle crinoidal member Conodonta:

CAI is 2.
Probably Lower Devonian, _serotinus_ Zone
Polygnathus _serotinus_
_I_. _trojani_?
Panderodus sp.

<u>Sample 222 fl: McColley Canyon Fm., Bartine Member</u> Conodonta:
CAI is 2.
Lower Devonian, <u>dehiscens</u> to <u>serotinus</u> Zones
<u>I. trojani</u>
<u>I. nevadensis</u>?
<u>Panderodus</u> sp.

<u>Sample 224 oc: McColley Canyon Fm, Coils Creek Member</u> Conodonta:
CAI is 2.
Probably late Lower-early Middle Devonian
<u>I</u>. sp. indet.
<u>Pandorinellina</u> <u>expansa</u>?
<u>Pand.</u> sp. indet.
<u>Panderodus</u> sp.
<u>Belodella</u> sp.

<u>Sample 241A fl: Antelope Valley Limestone</u> Conodonta:
CAI about 4; paleotemperatures somewhat above 200° C.
This fauna clearly is of Chazyan age; comparable conodonts occur in the lower Copenhagen and in the Crystal Peak Dolomite.
<u>Ansella</u> <u>nevadensis</u> (Ethington and Schumacher)
<u>Curtognathus</u> sp.
<u>Drepanoistodus</u> <u>aff</u>. <u>D</u>. <u>suberectus</u> (Branson and Mehl)
<u>Erraticodon</u> <u>balticus</u> Dzik
<u>Goverdina</u> <u>alicula</u> Fahraeus and Hunter
<u>Panderodus</u> sp.
<u>Phragmodus</u> <u>flexuosus</u> Moskalenko
<u>Staufferella</u> <u>aff</u>. <u>S</u>. <u>falcata</u> (Stauffer)
<u>N</u>. gen., <u>n</u>. sp. Raring of Harris et al., 1979, Pl. 3, fig. 8.

<u>Sample 257A fl: Antelope Valley Limestone</u> Conodonta:
CAI cannot be established from material at hand; specimens are badly fragmented.
The M element is of a type that is common in the Middle and lower Upper Ordovician of many places in North America. I have not seen this kind of element in the Whiterockian so I am inclined to believe that it is at least as young as Chazyan. Lofgren has reported similar elements from upper Llanvernian in Sweden, rocks that must be equivalent to the upper Whiterockian, so that I cannot say unequivocally that this is not a late Whiterockian fauna.
M element of <u>Oistodus</u> <u>venustus</u> type.
<u>Panderodus</u> sp.

Sample 258A fl: Antelope Valley Limestone Conodonta:

CAI 5; 300° C.
This is an abundant and diverse late Whiterockian fauna. The assemblage is very similar to that in the Lehman and most of these forms are present high in the Antelope Valley on Martin Ridge in the Monitor Range.

Ansella sp.
Chosonodina rigbyi Ethington and Clark
Drepanoistodus angulensis (Harris)
Erraticon balticus Dzik
Histiodella holodentata Ethington and Clark
?aff. Loxodus sp. of Ethington and Clark, 1982, Pl. 5, fig. 4.
?Multioistodus auritus (Harris and Harris) of Ethington and Clark, 1982, Pl. 6, figs. 6, 7.
Multioistodus compressus Harris and Harris
Paraprioniodus costatus (Mound)
Pteracontiodus gracilis Ethington and Clark
Scandodus sinuosus Mound
"Scolopodus" gracilis Ethington and Clark

Sample 313B oc: Devils Gate Limestone Conodonta:

CAI is 1.5.
Possibly Frasnian
Mehlina gradata? Youngquist
I. sp. indet.

Sample 359 oc: McColley Canyon Fm., Bartine Member Conodonta:

CAI is 1.5.
Probably Lower Devonian, possibly serotinus Zone
Icriodus latericrescens robustus? Orr
I. trojani?
Panderodus sp.
Belodella sp.
scolecodonts on slide

Sample 360 fl: McColley Canyon Fm., Bartine Member Conodonta:

CAI is 1.5.
Lower Devonian, inversus Zone
Polygnathus inversus Klapper and Johnson
P. laticostatus Klapper and Johnson
Pand. exigua exigua (Philip)
I. nevadensis
I. trojani
Belodella sp.
Panderodus sp.

<u>**Sample 378A oc: Sentinel Mountain-Bay State Dolomites Coelenterata:**</u>
Middle Devonian
?*Syringopora* sp.

<u>**Sample 386 oc: Sadler Ranch Fm., lower dolomite Conodonta:**</u>
CAI is 2.
Lower-Middle Devonian, *serotinus* to *costatus* Zones
Pandorinellina expanse
I. latericrescens robustus
Steptotaxis sp.
Polygnathus serotinus
P. linguiformis bultyncki?
Panderodus sp.
Belodella sp.

<u>**Sample 387 fl: Oxyoke Canyon Sandstone Conodonta:**</u>
CAI is 1.5.
Middle Devonian, *costatus* Zone
I. norfordi Chatterton

<u>**Sample 401 fl: McColley Canyon Fm., Bartine Member Brachiopoda:**</u>
Interval 11-13 (Johnson, 1977; Johnson and others, 1980)
Brachyspirifer pinyonoides

<u>**Sample 417B oc: Oxyoke Canyon Sandstone Conodonta:**</u>
CAI is 1.5.
Middle Devonian, *costatus* Zone
I. norfordi
Pand. expansa

<u>**Sample 425 oc: Sentinel Mountain-Bay State Dolomites Brachiopoda:**</u>
Fossils identified from photograph of outcrop
Interval 19-20 (Johnson, 1977; Johnson and others, 1980)
?*Geranocephalus* sp.

<u>**Sample 489 oc: Lone Mountain Dolomite Conodonta:**</u>
CAI is 3.
Silurian, *celloni* Zone, Llandovery
Pterospathodus celloni (Walliser)
Ozarkodina plana (Walliser)
Panderodus sp.

APPENDIX II

Revision of Devonian Stratigraphy South of Eureka, Nevada

Peter F. Cowell and J.G. Johnson

Introduction

Work in the South Hill area clarifies Lower and Middle Devonian stratigraphy. This note explains the circumstances that lead to the naming of the anomalous formation called "South Hill Sandstone" by Nolan and others (1974, p.3) and why that name should now be abandoned.

Nolan and others (1956) proposed five members for the Nevada Formation; they were given formation status by Hose and others (1982, p.1). These formations, Beacon Peak Dolomite, Oxyoke Canyon Sandstone, Sentinel Mountain Dolomite, Woodpecker Limestone, and Bay State Dolomite, occur in ascending order at several sections in the Pinto Summit quadrangle, but at South Hill, along the western margin of that quadrangle, Nolan and others (1974) recognized a different succession: Beacon Peak Dolomite, Grays Canyon Limestone (new), Sentinel Mountain Dolomite, South Hill Sandstone (new), and Bay State Dolomite (Fig. 6). The South Hill Sandstone was thought to be in the position of the Woodpecker Limestone. This was anomalous because the South Hill, a shallow-water quartzose dolomite, stated to be very similar to the Oxyoke Canyon Sandstone (Nolan and others, 1974, p.3), would be in an offshore position to the Woodpecker Limestone, which formed in response to a deepening event (Johnson and others, 1985). To further complicate matters, the South Hill was mapped as in thrust (or low-angle normal) fault contact with formations above and below (Nolan and others, 1974).

The unique succession reported by Nolan and others (1974) at South Hill, and at Grays Canyon in the adjacent Bellevue Peak quadrangle, involved several interrelated misinterpretations. Changing any one created problems with the others. Below, we examine the interpretations on which the Nolan and others (1974) stratigraphy is based and reidentify the units in question.

Stratigraphic problems

The Grays Canyon Limestone has the *Eurekaspirifer pinyonensis* Zone brachiopod fauna and the characteristic lithology of the Bartine Member of the McColley Canyon Formation (Murphy and Gronberg, 1970). The name Grays Canyon Limestone has therefore been abandoned (Johnson and Murphy, 1984, p.1349). The problem lies not with the name, but with Nolan and others' (1974, p.3) interpretation that this unit is the western correlative of the Oxyoke Canyon Sandstone. The same interpretation was published elsewhere by Merriam (1963, Fig. 8, p.52). More recent work (Kendall and others, 1983), based on conodont biostratigraphy, has demonstrated that the Sadler Ranch Formation, above the Bartine Member, is correlative with the lower part of the Oxyoke Canyon Sandstone (Fig. 6, Plate 1).

Above their Grays Canyon Limestone, Nolan and others (1974) identified the Sentinel Mountain Dolomite (the unit expected above the Oxyoke Canyon Sandstone). These rocks instead belong to the Sadler Ranch Formation and have the characteristic 2-hole crinoid ossicles of Gasterocoma? bicaula (Johnson and Lane, 1969). Further, the conodont Polygnathus serotinus is present, indicating a position significantly older than Sentinel Mountain Dolomite.

Above the Sadler Ranch beds are the rocks Nolan and others (1974) called South Hill Sandstone. They supposed that their South Hill Sandstone was in the position of the Woodpecker Limestone because rocks below were identified by them as Sentinel Mountain Dolomite and because thay found no Woodpecker Limestone. The rocks called South Hill have yielded conodonts (Pandorinellina expansa and Icriodus norfordi) indicative of the costatus Zone, far older than Woodpecker and known elsewhere below and in facies relationship with the upper, coarse crystalline beds of the Oxyoke Canyon Sandstone (Kendall and others, 1983, Fig. 5). The South Hill Sandstone is in fact a tongue of upper Oxyoke Canyon Sandstone; this explains the remark of Nolan and others (1974, p.3) that these rocks are very similar to the Oxyoke Canyon Sandstone.

Above their South Hill Sandstone, Nolan and others (1974) identified the Bay State Dolomite, the formation expected above the Woodpecker Limestone. We identify this unit as Sentinel Mountain-Bay State undivided. The two units are similar lithologically and at most places cannot be separately distinguished if the Woodpecker does not intervene. The absence of the Woodpecker at the western edge of the carbonate section is now well known (Johnson and Sandberg, 1977, Fig. 7).

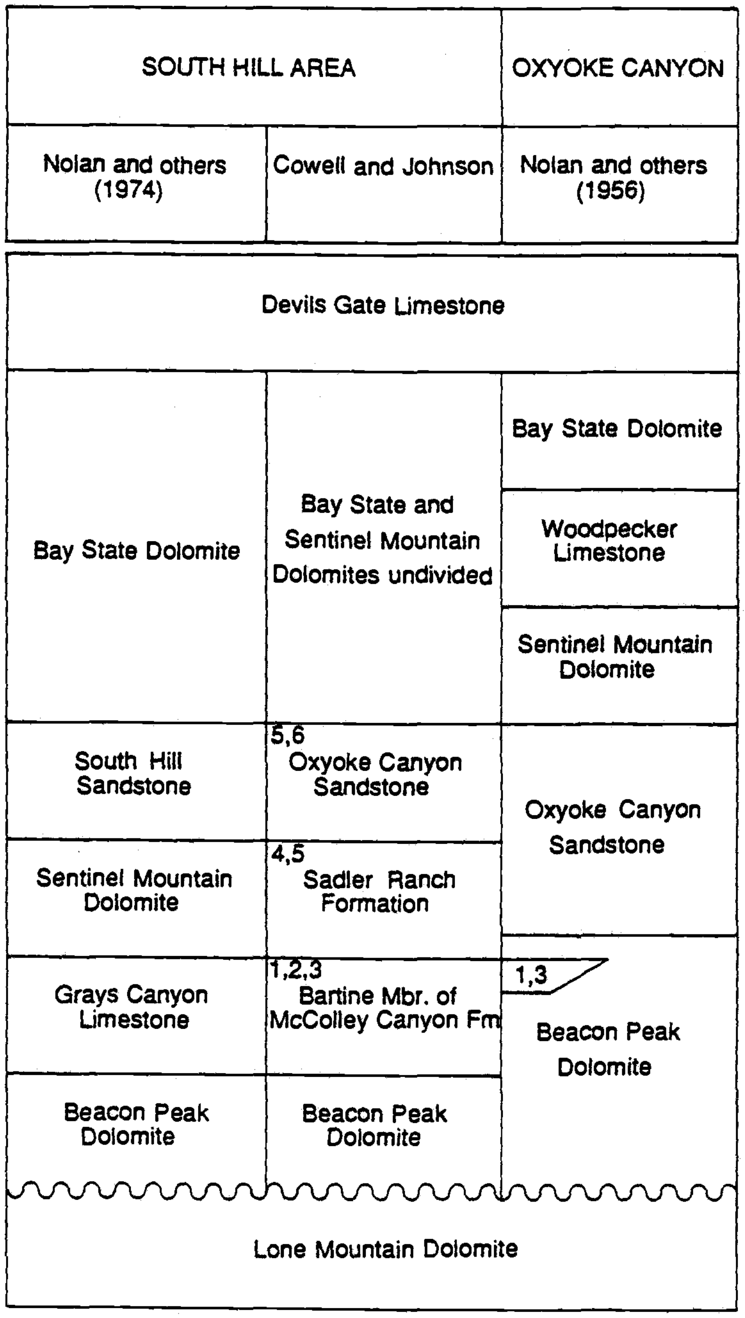

Figure 6. Correlation chart of Devonian rocks (except Pilot Shale) south of Eureka. The two columns on the left show the different stratigraphic nomenclatures of the South Hill area, discussed here. The fossiliferous tongue of Bartine Limestone in Beacon Peak Dolomite at Oxyoke Canyon was reported by Kendall and others (1983). Numbers refer to age-significant conodont species, as follows: 1 is Pandorinellina exigua exigua; 2 is Polygnathus laticostatus; 3 is Polygnathus inversus; 4 is Polygnathus serotinus; 5 is Pandorinellina expansa; 6 is Icriodus norfordi.