AN ABSTRACT OF THE THESIS OF

John Paul Graham for the degree of Master of Science

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Title: Devonian and Mississippian Stratigraphy of the Southern Hot Creek Range, Nye County, Nevada.

Abstract approved: Redacted for Privacy

J. G. Johnson

Devonian carbonate rocks of the southern Hot Creek Range reflect a shift to the east of the carbonate shelf edge and a progressively deeper-water lithotope from late Early Devonian to early Middle Devonian time. Lower Devonian mudstones of the McColley Canyon Formation were deposited on an open shelf, dolomitized, and then flooded after gronbergi Zone time. As the sea transgressed eastward, a slope-basin lithotope developed. This deepening event is reflected in the hemipelagic deposition of the Middle Devonian Denay Limestone. Rapid flooding occurred during the Middle varcus Subzone with the onset of the Taghanic onlap. Pelletal and bioclastic-rich allodapic limestones represent distal deposition from turbidity currents originating on a slope to the east. A brief shallowing event is reflected in the rock succession at Empire Canyon and probably occurred in Interval 27, dengleri Zone. No Upper Devonian rocks have been identified in the southern Hot Creek Range or in the Warm Springs area.

Deep-water deposition resumed in the Mississippian. A rising Antler highland in the west provided sediments to a foreland trough during the Early Mississippian. Conodonts from the Warm Springs area were dated to range across the Kinderhookian-Osagean boundary, from the *isosticha*-Upper *crenulata* Zone into the *typicus* Zone. Allodapic mudstones to detrital quartz packstones, deposited in a submarine fan complex, represent distal turbidity currents originating to the west and ponding in the trough.

The stratigraphy of the southern Hot Creek Range has not been disrupted by the Roberts Mountains thrust. The limestones suggest deposition in a deepening marine lithotope as the carbonate shelf edge migrated to the east. Apparently, the Roberts Mountains allochthon progressed no farther east than the Dobbin Summit area in the central Monitor Range.

Devonian and Mississippian Stratigraphy
of the Southern Hot Creek Range,
Nye County, Nevada

by

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Professor of Geology in charge of major

Redacted for Privacy

Chairman of Department of Geology

Redacted for Privacy

Dean of Graduate School

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Typed by John Paul Graham

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G. Klapper of The University of Iowa introduced me to Dr. Johnson, and for this, I will always be grateful. I appreciated his patience and thoughtfulness when I was his student, and I appreciated his willingness to identify the conodonts in this thesis.

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I especially thank my parents for their continued, unwavering support for their wayward son. I thank Lynn, my wife, for still speaking to me after all she had to endure.

I thank A. Niem and K. F. Oles for reviewing the manuscript and for serving on the thesis committee.

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF PLATES

THE DEVONIAN OF NEVADA

It begins with a night of transgression
when placid Devonian sea bottoms
were borne up by wild thrusts
imbricately, one on top of the other, older on younger
(if there had been streets,
     they would have filled with dancing)
the toe of the big thrust prancing from the West
     over the bared flank of Nevada
(for all purposes, the punch was spiked
     but what happened seemed quite natural)
the sea plunged in, barriers burst,
the gentle arch was overlapped.
The extravagance of transgression,
(the bouquet of fine wine)
the Mississippian held in an unconformable alliance
was followed by the following morning,
     millinia later,
and the belch of epeirogeny,
the headache of regression--

Revelry abated, the land, much later,
crumpled, bent, displaced
by the passing of time
left forlorn and desolate,
baked by the sun,
loves of an ancient sea left to memory,
the Devonian became a mecca only for vultures,
     cacti, an occasional cat,
a wallflower at any party of pioneers,
poking its ragged outcrops
from mountaintops
strung out across Nevada,
bleached and buried by the bones of centuries
until, one day, the rocks in their silence
rang with the blow of hammers
began to fizz with renewed excitement,
awakening under the microscope
to begin to tell their tale.

J.Graham, 1982

Double rainbow over Warm Springs, Nevada.

DEVONIAN AND MISSISSIPPIAN STRATIGRAPHY OF THE SOUTHERN HOT CREEK RANGE, NYE COUNTY, NEVADA

INTRODUCTION

Purpose

Biostratigraphic and petrographic relationships of the Devonian and Mississippian carbonate rocks of the southern Hot Creek Range, Nye County, Nevada (Figure 1), are analyzed in this report. The area was chosen because detailed study of these rocks had not been done and because information gleaned from such a study could potentially help settle the dispute over the southern trace of the Roberts Mountains thrust in Nye County.

The objectives of this thesis include: 1) describing the geology and stratigraphy of Mississippian and Devonian rock sequences, 2) placing the rock units in a time-stratigraphic framework, 3) interpreting the depositional environments of these rocks, and 4) putting these rocks into a regional perspective with regards to the Antler orogeny and the Roberts Mountains thrust.

To meet these objectives, stratigraphic sections were measured and described at Warm Springs (Plate 1), Rawhide Mountain (Plate 2), and Empire Canyon (Plate 2). Samples were collected and processed for conodonts which then were used to date the rocks.

The reconstruction of a mid-Paleozoic paleogeography and depo-tectonic history for Nevada is an evolving concept. The southern Hot Creek Range provides pertinent information to the understanding of the history of the Antler orogeny and the emplacement of the Roberts Mountains thrust.

Location and Accessibility

Warm Springs, Rawhide Mountain, and Empire Canyon are all located in Nye County, Nevada (Figure 1). Paleozoic rocks exposed on the nose of the southern extension of the Hot Creek Range behind the town of Warm Springs are bracketed by U.S. Highway 6 to the east and south and by Tertiary volcanics to the west (Plate 1, Figure 2). In the NW 1/4 of the Warm Springs quadrangle, Paleozoic rocks straddle T4N, R50E, and T4N, R49 1/2E. Access is gained primarily by a dirt road leading up to the radio tower from Highway 6 and by foot. Mining roads on the eastern flank provide access to the jasperoid deposits.

Rawhide Mountain (Plate 2) is located 8 miles (13 km.) N10W from peak 6948 at Warm Springs (Figure 3) and is in the southwest quarter of the Tybo fifteen minute quadrangle, T5N, R49E. Access is gained mainly by foot after following a dirt road west from Highway 6 and a gravel, rocky road up Crystal Canyon until the gradient becomes too steep. A wall of jasperoid forms a natural barrier on the southern flank of Rawhide Mountain. A rugged road has been carved through this wall and stands as a monument to the determination of today's mining operations. However, even with this road, the Upper Devonian outcrops on the western flank remain virtually inaccessible.

Sections measured in Empire Canyon, approximately 9 miles (15 km.) N8W from peak 9169 at Rawhide Mountain, are located in the northwest quarter of the Tybo quadrangle, R49E, and are bisected by the boundary between T6N and T7N (Plate 2, Figure 4). Access is gained over rutted roads which traverse the saddles between Empire Canyon and either Tybo Creek to the south or Keystone Canyon to the north. The road leading to Tybo from

Highway 6 is less corrugated than the side road to Keystone Canyon.

Rawhide Mountain and Empire Canyon are located at elevations which accomodate both juniper trees and single-leaf pinyon pines. Sagebrush and cheat grass provide ground cover. Warm Springs, at an elevation of 6948 feet and on a southern exposure, contains scattered juniper trees, cholla and prickly pear cacti, sagebrush, and cheat grass. Wild horses and coyotes inhabit all three areas. Signs of cattle are ubiquitous.

Previous Work

Little serious attention has been paid to the Devonian and Mississippian facies relationships in the southern Hot Creek Range. Stewart and Poole (1974) referred to "siliceous mudstone, radiolarian chert, and sparse coralline limestone near Warm Springs" as belonging to a Devonian "shale and chert facies" (p. 47) which reflected deep marine conditions westward of the outer edge of the shelf. This interpretation was briefly reemphasized by Stewart (1980) in a paragraph wherein he also referred to the Devonian rocks near Warm Springs as "the easternmost outliers of the Roberts Mountains allochthon" (p. 35). No mention is made in Stewart (1980) of Mississippian rocks at the same locality, but the Geologic Map of Nevada (Stewart and Carlson, 1978) and the Preliminary Geologic Map of Northern Nye County (Kleinhampl and Ziony, 1967) do plot Mississippian limestone near Warm Springs. In 1979, a road log published in conjunction with the Basin and Range Symposium mentioned a Kinderhookian "klippen" behind the Warm Springs Cafe resting on Middle Devonian argillite and chert (French et al., 1979, p. 651).

Quinlivan and Rogers (1974) mapped the Paleozoics of Rawhide Mountain and Empire Canyon, but they did not delineate the formations of the Lower and Middle Devonian. McGovney (1977) measured three stratigraphic sections of Upper Silurian and Lower Devonian carbonate rocks in Hot Creek Canyon, 9 miles (14 km.) north of Empire Canyon, but he did not map the area. Detailed mapping and lithologic descriptions of Middle and Upper Devonian and Lower Mississippian rock units have been done in the northern Hot Creek Range by Potter (1975) and in the central Monitor Range, northern Nye County, by Wise (1976).

Terminology

Both Dunham's (1962) and Folk's (1962) carbonate classification schemes were scrutinized for use in this report. Dunham's classification, while useful for field descriptions of these carbonate rocks, fails to account for the diagenetic effects seen in thin section. Folk's classification, on the other hand, is useful for petrographic distinctions but is more cumbersone to use in the field. Combination of the two led to confusion and lack of clarity. As a consequence, Dunham's classification has been chosen on the following merits: 1) the classification has the practical advantage of easy recognition of rock types in the field, 2) the depositional texture has not been destroyed by later diagenesis, and 3) the rock name reflects the original ratio of mud : allochems from which a history of diagenesis may proceed.

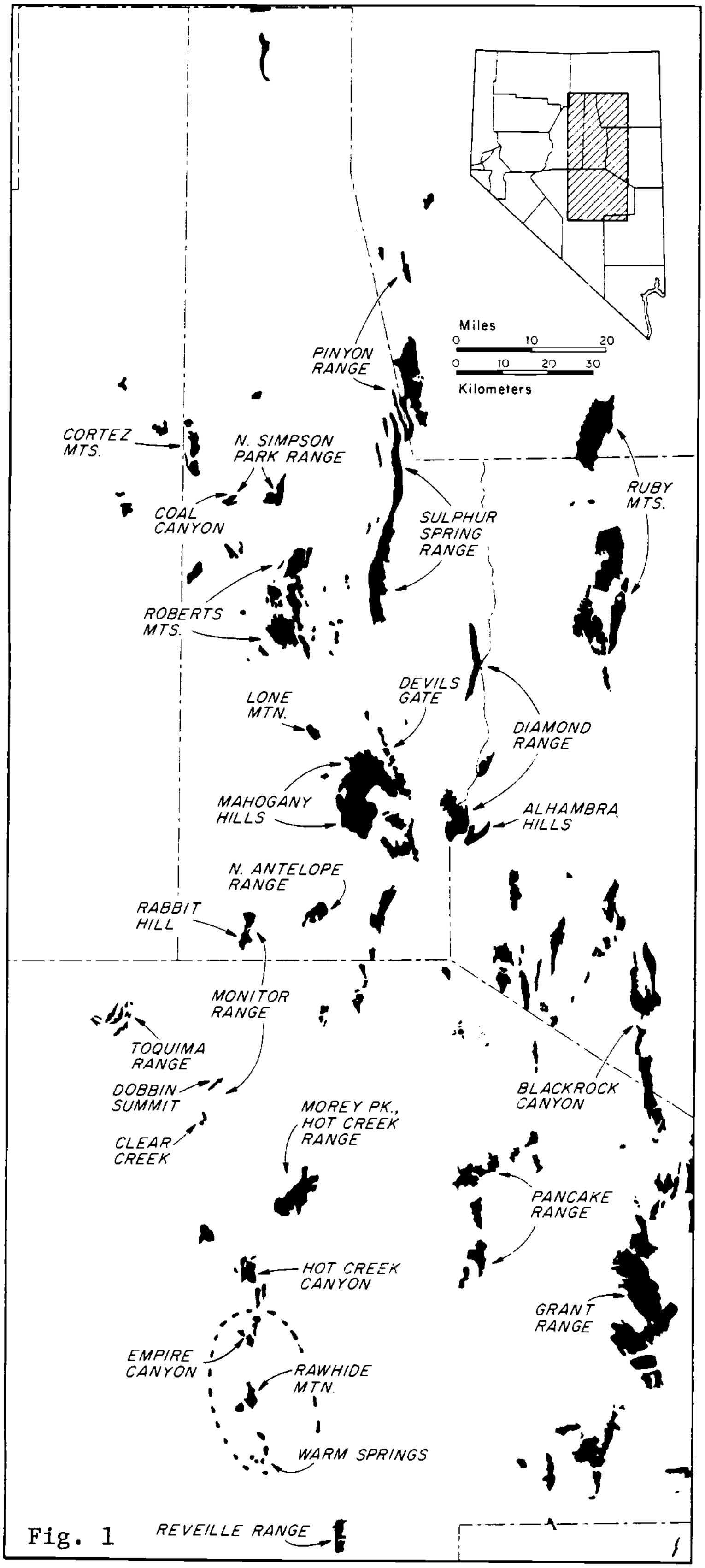
Bedding thickness is described using the terminology of McKee and Weir (1953) wherein laminated strata are 2 mm. to 1 cm. (about 1/16 to 1/2 in.) thick, very thin-bedded strata are 1 - 5 cm. (about 1/2 to 2 in.) thick, thin-bedded strata are 5 - 60 cm. (about 2 in. to 2 ft.) thick, thick-bedded strata are 60 - 120 cm. (about 2 to

4 ft.) thick, and very thick-bedded strata are greater than 120 cm. (4 ft.) thick.

Methods

Field work was conducted in the abnormally cold autumn of 1981 and the abnormally wet June of 1982. Seven sections were described and measured (using Jacob staff and Abney level; Plates 3-9). Paleozoic units in the Warm Springs area were mapped to a scale of 1:12,000 using enlarged 7.5 minute quadrangles as a base (Plate 1). Forty-two thin sections were analyzed, about half of which were stained to distinguish between calcite and dolomite. Seventy-two samples, exclusive of the 26 samples collected from Potter's (1976) thesis area in June, averaging 6.9 pounds (3.1 kg.) each were split and dissolved in dilute formic acid in order to extract conodonts. Each sample underwent heavy liquid separation using tetrabromoethene, and the heavy fraction was hand picked for conodonts. Claudia Regier and Jim Brown were invaluable guides during this process. Gilbert Klapper of the University of Iowa then identified the conodonts and assigned them to biostratigraphic zones. J. G. Johnson identified the brachiopods found in the samples. Brachiopods were, unfortunately, few and poorly preserved.

Figure 1: Index map of central Nevada with Devonian outcrops shown in black. Inset in upper right shows area covered. Thesis area is circled by dashed lines.


Miles
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Kilometers
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CORTEZ MTS.
N. SIMPSON PARK RANGE
COAL CANYON
SULPHUR SPRING RANGE
RUBY MTS.
ROBERTS MTS.
LONE MTN.
DEVILS GATE
DIAMOND RANGE
MAHOGANY HILLS
ALHAMBRA HILLS
N. ANTELOPE RANGE
RABBIT HILL
MONITOR RANGE
TOQUIMA RANGE
DOBBIN SUMMIT
CLEAR CREEK
MOREY PK., HOT CREEK RANGE
BLACKROCK CANYON
PANCAKE RANGE
HOT CREEK CANYON
GRANT RANGE
EMPIRE CANYON
RAWHIDE MTN.
WARM SPRINGS
REVEILLE RANGE

Fig. 1

Figure 2: Mississippian limestone capping peak 6948 behind Warm Springs, Nevada. Looking west from Highway 6. Tertiary volcanics in foreground.

Figure 3: Looking north from peak 6948 to Rawhide Mountain, the distant peak in the center of the photograph.

Figure 4: Empire Canyon. View from the east towards Cine Mountain, the flat-topped mountain on the right.

GEOLOGIC SETTING

During the early and middle Paleozoic, a north-south-trending Cordilleran geosyncline was the major structural feature of western North America. Within this geosyncline, deposition took place in three distinct lithofacies belts: 1) an eastern carbonate assemblage of dolomite, limestone, and quartz arenite deposited on a broad, shallow, subtidal, intertidal, and supratidal carbonate platform; 2) a limestone-clastic, transitional assemblage representing a subtidal basin or outer shelf environment oceanward from the carbonate platform; and 3) a western deep basin, siliceous clastic assemblage (Smith and Ketner, 1968; Matti et al., 1975). Stewart and Poole (1974) identified a fourth lithofacies belt, a chert province, beyond the siliceous clastic assemblage. Faunal changes, as well as lithic variations, indicate migrations of the carbonate shelf boundary during regressive and transgressive episodes within the Devonian (Johnson, 1970; Johnson and Sandberg, 1977; Klapper and Johnson, 1980).

Two basic tectonic models have evolved for the depotectonic setting of the western United States continental margin prior to the Middle Devonian. One model relies on a marginal basin-island arc structural setting with either an east-dipping (Burchfield and Davis, 1972; Poole, 1974; Stewart and Poole, 1974; Churkin, 1974; Poole et al., 1977b; Dickinson, 1977) or a west-dipping (Schweickert and Snyder, 1981) subduction zone. The other model, analogous to the Atlantic Coast today, postulates deposition occurring along a stable continental margin formed from Precambrian rifting (Stewart, 1972, 1976; Stewart and Poole, 1974; Poole, 1974).

After a major Lower Devonian regression, a transgression rapidly flooded the carbonate platform (Johnson, 1971; Johnson and Sandberg, 1977; Klapper and Johnson, 1980). The depositional setting in the area switched from a gentle ramp to a deep marine, slope-basin setting by late Middle Devonian time.

Effects of the Antler orogeny were felt in the area beginning in Late Devonian and extending into the Early M.ssissippian. The principal structural event of the Antler orogeny in Nevada was the emplacement of the Roberts Mountains allochthon which superposed western assemblage rocks onto eastern assemblage rocks by mid-Mississippian time (Dott, 1955; Roberts et al., 1958; Smith and Ketner, 1968; Johnson, 1971; Gutschick et al., 1980; Johnson and Pendergast, 1981). During the Early Mississippian, detritus from the allochthon was shed eastward into a subsiding foreland trough (Poole, 1974; Johnson and Pendergast, 1981). Pelagic sediments and sediments eroded from the rising highland formed a submarine fan complex, the distal lithotope of which is represented in the allodapic beds of Warm Springs.

Precipitation of jasperoid in association with gold, aresenic, silver, antimony and other economic deposits occurred during the Tertiary in conjunction with the high angle, north-northeast-trending normal faults which produced the Basin and Range fault block topography present today. Unfortunately, Tertiary volcanism has concealed nost of the Paleozoic facies relationships in the southern Hot Creek Range.

BIOSTRATIGRAPHY

Empire Canyon

Devonian strata in the measured sections in Empire Canyon were dated using conodonts and range from Pragian to Givetian in age (Plate 2, Figure 5). At least three and probably all four of the upper Lower Devonian phyletic lineage zones are represented in ECIII at the western end of the canyon (Figure 6). ECII, located at the eastern end of the canyon, contains conodonts indicative of the upper Middle Devonian, upper Middle _varcus_ to Upper _varcus_ Subzone (Appendix I). Devonian rocks above bed ECII-152 probably contain the youngest strata in Empire Canyon and Rawhide Mountain. However, no diagnostic conodonts or brachiopods were recovered to date these beds.

Section ECI (Appendix I), the easternmost section measured in Empire Canyon, contains the diagnostic conodont, _Tortodus kockelianus kockelianus_, in all but the highest sample, ECI-228. However, due to its spatial proximity to sample ECI-226 and the occurrence of _Polygnathus eiflius_, sample ECI-228 may also be in the _kockelianus_ Zone. Although found in the _ensensis_ Zone in the Eifelian Hills and the Rhenish Slate Mountains (Klapper and Ziegler, 1979; Klapper and Johnson, 1980), _P. eiflius_ has been found only in the _kockelianus_ Zone in Nevada (Klapper and Johnson, 1980; Johnson et al., 1980). Furthermore, sequential collections in the Devonian of the northern Antelope Range, central Nevada, suggest that _P. eiflius_ may be restricted to the lower part of the _kockelianus_ Zone in Nevada (Johnson et al., 1980, Tables 16, 17). Likewise, the sample collected two feet from the base of section ECII contained _T. k. kockelianus_, providing a Middle Devonian, _kockelianus_

Zone age assignment (Appendix I, Figure 5).

Three genera of brachiopods were also collected from ECI-226 and ECI-228 (Appendix II). They support an Eifelian age assignment for section ECI.

Although no zonal determination could be gleaned from the conodonts of ECII-150, the sample did yield brachiopods containing the Middle Devonian genus, Stringocephalus (Appendix II). A Givetian age was assigned to this collection.

Givetian age conodonts were recovered from the crinoidal wackestone at ECII-152 and assigned to the probable zonal range of the uppermost part of the Middle varcus Subzone to Upper varcus Subzone. The The Middle and Upper varcus Subzones are defined by the first occurrences of P. ansatus and P. latifossatus, respectively (Ziegler et al., 1976; Klapper, 1977). While P. latifossatus is not present in ECII-152, the collection does contain P. ansatus. Conodonts collected from ECII-152 are the same age as conodonts recently collected from sample X13 in the northern Antelope Range (J. G. Johnson, pers. comm., 1982; Trojan, 1978).

Sections ECI and ECII were measured from the eastern end of Empire Canyon. However, ECIII was measured at the westernmost exposure of Devonian strata, west of Cine Mountain (Plate 2). ECIII contains the oldest Devonian rocks found in the combined measured sections of Empire Canyon, Rawhide Mountain, and Warm Springs (Figures 6, 7, 9), and these have been dated as upper Lower Devonian (Appendix I). Pandorinellina exigua philipi was identified in the dolomitized mudstone of ECIII-22. Pand. exigua philipi ranges into the gronbergi Zone in Alaska and New South Wales, but is apparently restricted to the dehiscens Zone in Nevada (Klapper and Ziegler, 1979, Figure 2; Klapper and Johnson, 1980, Table 5). P. gronbergi, the nominative species for the next superjacent zone, was found in a collection from sample ECIII-60.

Above a thirty foot covered interval between ECIII-60 and ECIII-90 (Figure 6, Plate 5), sample ECIII-91, a pelletal grainstone, yielded both P. parawebbi and P. costatus costatus. The overlap of these two ranges in Nevada suggests an australis Zone age assignment (Johnson et al., 1980, Table 23).

Significantly, the thirty foot covered interval previously mentioned represents a condensed section in which the four conodont zones between the gronbergi Zone and the australis Zone are either missing or covered. Potter (1975) measured about 918 feet (280 m.) of strata between his pinyonensis Zone fauna (dehiscens Zone equivalent) and his Leptathyris cf. L. circula Zone fauna (australis Zone equivalent) in the northern Hot Creek Range (Potter, 1975, Plate 3). Trojan (1978) measured about 220 feet (67 m.) between the gronbergi and australis Zones and about 370 feet (113 m.) between the dehiscens and australis Zones in his measured section II (Trojan, 1978; Johnson et al., 1980).

Fossils indicative of the kockelianus Zone were found in the overlying samples ECIII-144, ECIII-184, ECIII-212, and ECIII-220. P. parawebbi, which does not range above the Lower varcus Subzone (Johnson et al., 1980), was recovered from ECIII-245. This sample was assigned to the kockelianus Zone to Lower varcus Subzone by Klapper on the basis of superposition (Figure 6).

Rawhide Mountain

Three probable Middle Devonian overlap range zones are represented in the one section from Rawhide Mountain (Plate 2, Figure 7, Appendix I). Sample RM-49 yielded the diagnostic conodont, Tortodus k. kockelianus. T. k.

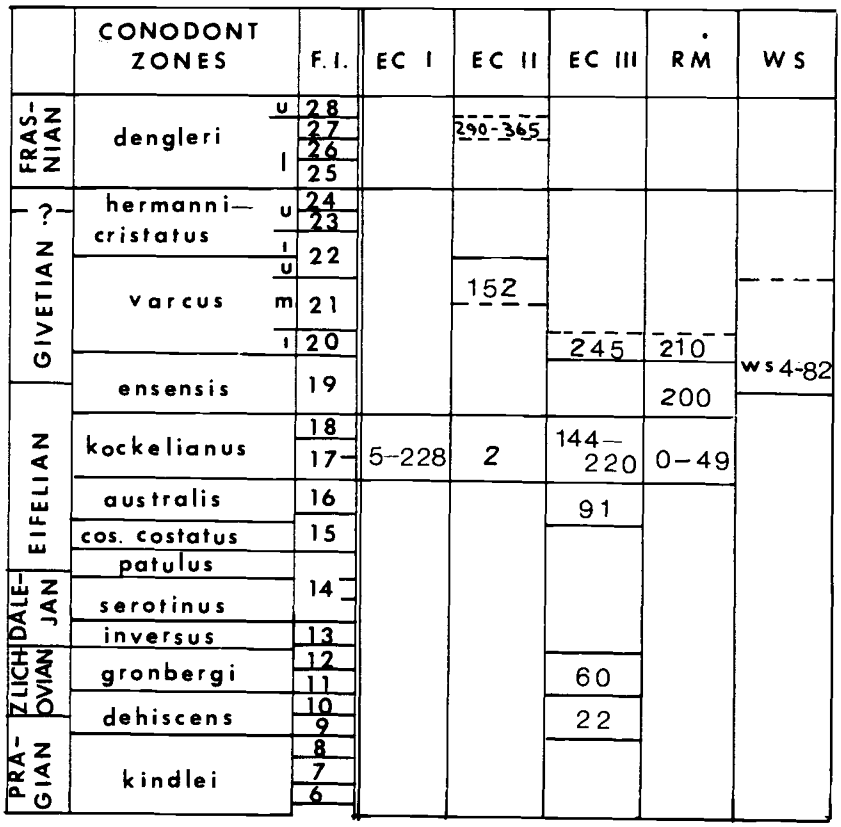
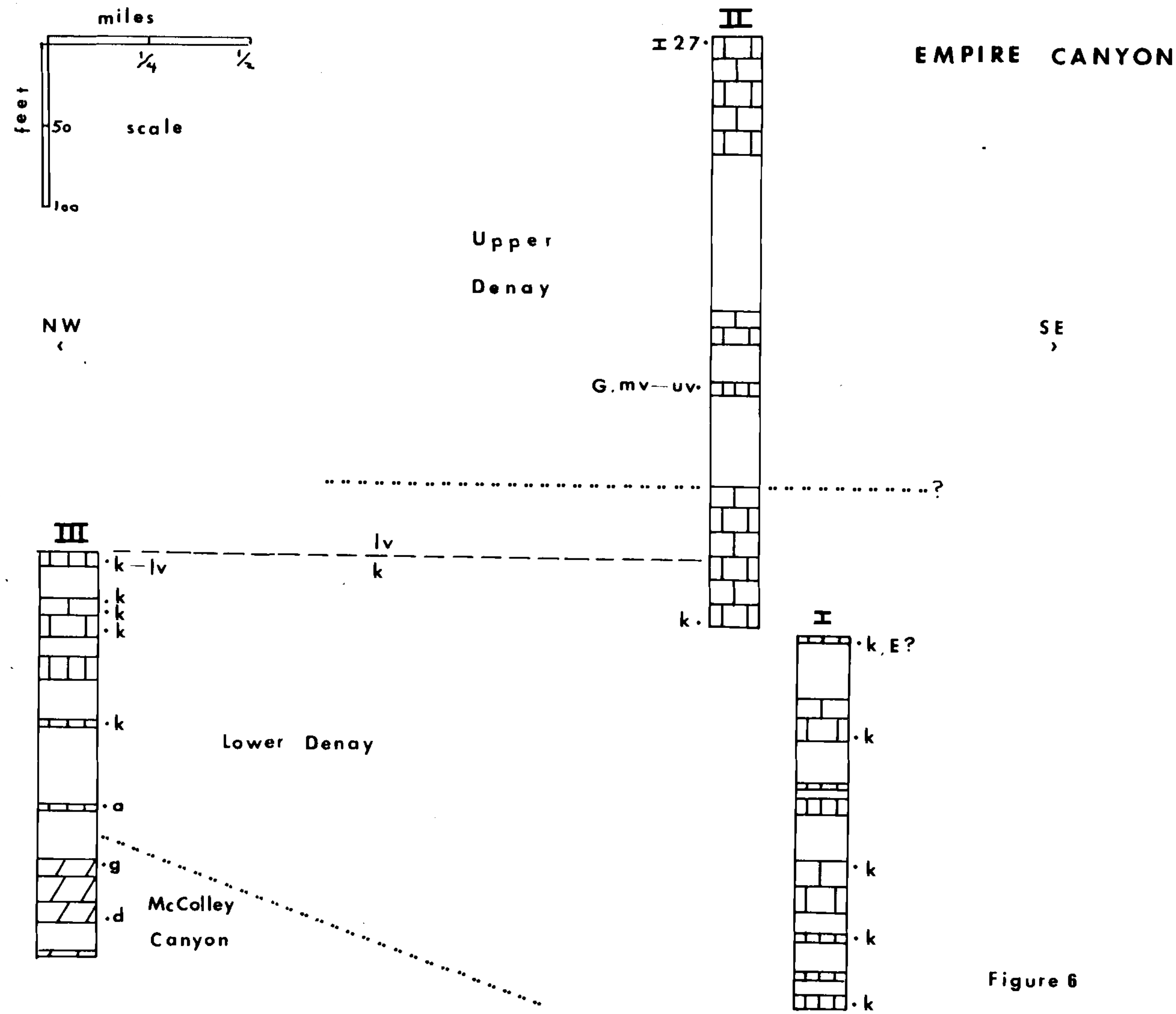


Figure 5: Biostratigraphic correlation chart for the Lower (part) and Middle Devonian of the southern Hot Creek Range. Numbers in vertical columns beneath sections ECI, ECII, ECIII, and RM refer to footages in measured sections where diagnostic fauna were collected. WS 4-82 is a random sample of undifferentiated Devonian limestone from Warm Springs which was dated using brachiopods. Numbers 6-24 are faunal intervals of Johnson et al., (1980). Both conodont zonation and faunal intervals are from Figure 5 of Johnson et al., (1980). EC refers to Empire Canyon; RM to Rawhide Mountain; and WS to Warm Springs.

Figure 6: Correlation chart of Lower and Middle Devonian units in Empire Canyon. Roman numerals at the top of the vertical columns refer to measured sections aligned in a northwest (left) to southeast (right) direction. Dots next to the columns indicate position of zonally diagnostic samples. Refer to Plate 2 for position of measured sections. Generalized lithology from Plates 3-5. Unlined areas are covered intervals. Small case letters refer to the following conodont zones: d, *dehiscens*; g, *gronbergi*; a, *australis*; k, *kockelianus*; lv, mv, uv, lower-middle-upper *varcus*, respectively. Large case letters refer to Stages identified from brachiopods as follows: E, Eifelian; G, Givetian. Large case letters followed by numbers (I27) refer to Intervals of Johnson et al., (1980). Dashed line represents a biostratigraphic boundary. Double dotted line is a lithologic boundary.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Lower Denay
a
g
d
McColley
Canyon
k
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k

Figure 6

kockelianus has been found to range into the superjacent *ensensis* Zone in the Eifelian Hills and the Rhenish Slate Mountains (Klapper and Ziegler, 1979, p. 210; Klapper and Johnson, 1980, Table 9), but *T. k. kockelianus* does not overlap *P. xylus ensensis* in Nevada (Johnson et al., 1980; Klapper and Johnson, 1980). Consequently, a *kockelianus* Zone, Eifelian age assignment may be established at the base of the section.

Undiagnostic conodonts were found in the next two samples, but sample RM-200 yielded conodonts probably associated with the *ensensis* Zone, specifically, *Polygnathus xylus ensensis*, the nominal form whose first occurrence marks the base of the zone. In the Eifelian type area, however, *P. x. ensensis* does range upward into the Givetian (Klapper and Ziegler, 1979, p. 210), and in Nevada, *P. x. ensensis* has been found to range into the Lower *varcus* Subzone in the northern Antelope Range (Johnson et al., 1980, Table 23; Klapper and Johnson, 1980, Table 10, p. 445). Without a characteristic *ensensis* fauna, the age assignment must remain a probability.

In the allodapic bed ten feet above RM-200, conodonts spanning the *ensensis* Zone to the Lower *varcus* Subzone were found. *P. x. ensensis* was not recovered, nor was a characteristic Lower *varcus* Subzone fauna, but *P. parawebbi* and *P. pseudofoliatus* were present. Although these are long-ranging undiagnostic species, neither ranges higher than the Lower *varcus* Subzone (Klapper and Ziegler, 1979; Klapper and Johnson, 1980; Johnson et al., 1980).

Brachiopods recovered from RM-200 and RM-210 also indicate a Middle Devonian age assignment (Appendix II, Figure 7). The specimens were preserved in such a state that they could not be identified beyond the generic level. Consequently, as with Empire Canyon, zonal assignment cannot be refined into the Interval concept which incorporates both conodont and brachiopod zones (Johnson, 1977; Johnson et al., 1980; Figure 5).

Warm Springs

A Mississippian fauna was collected from the measured sections at Warm Springs (Plate 1, Figures 8, 9). That the lowest zone is the *isosticha*-Upper *crenulata* Zone is indicated by the occurrence of *Gnathodus punctatus* and *G. delicatus* in WSI-1 (Appendix I). *G. punctatus* is first found in the *isosticha*-Upper *crenulata* Zone whose base is defined by the first appearance of *G. delicatus* (Sandberg, 1979).

The diverse conodont fauna present in some Warm Springs collections (Appendix I) reiterates the problems concerning zonal determination that were outlined by Dombrowski and Klapper in Johnson and Pendergast (1981, p. 656) with regard to similar Mississippian collections from the Swales Mountain area. As *G. punctatus* and *G. delicatus* both range across the Kinderhookian-Osagean boundary (Sandberg et al., 1978, p. 116; Sandberg, 1979, p. 100-101), a stage assignment is impossible unless certain *Siphonodella* species can be shown to be indigenous (Sandberg, 1979, Figure 5, p. 100; Johnson and Pendergast, 1981, p. 656). *Siphonodella* present in the collections, but which were extinct by Lower *crenulata* Zone time, include *S. cooperi* and *S. duplicata*. *S. obsoleta*, *S.* cf. *S. isosticha*, *S. quadruplicata*, and *S. lobata* first occur below the Lower *crenulata* Zone, but range into the *isosticha*-Upper *crenulata* Zone (Sandberg et al., 1978). If these latter *Siphonodella* species as well as *S. crenulata* and *S. isosticha*, which are restricted to the Lower crenulata to *isosticha*-Upper *crenulata* Zone, could be shown to be indigenous, then the

Figure 7: Correlation chart of Middle Devonian units at Rawhide Mountain and Empire Canyon. Measured section RM is positioned in a north (left) to south (right) trend along depositional strike from Empire Canyon. Symbols as for Figure 6 with the addition of the following: e, *ensensis*. Position of RM is located on Plate 2; general lithology is from Plate 6.

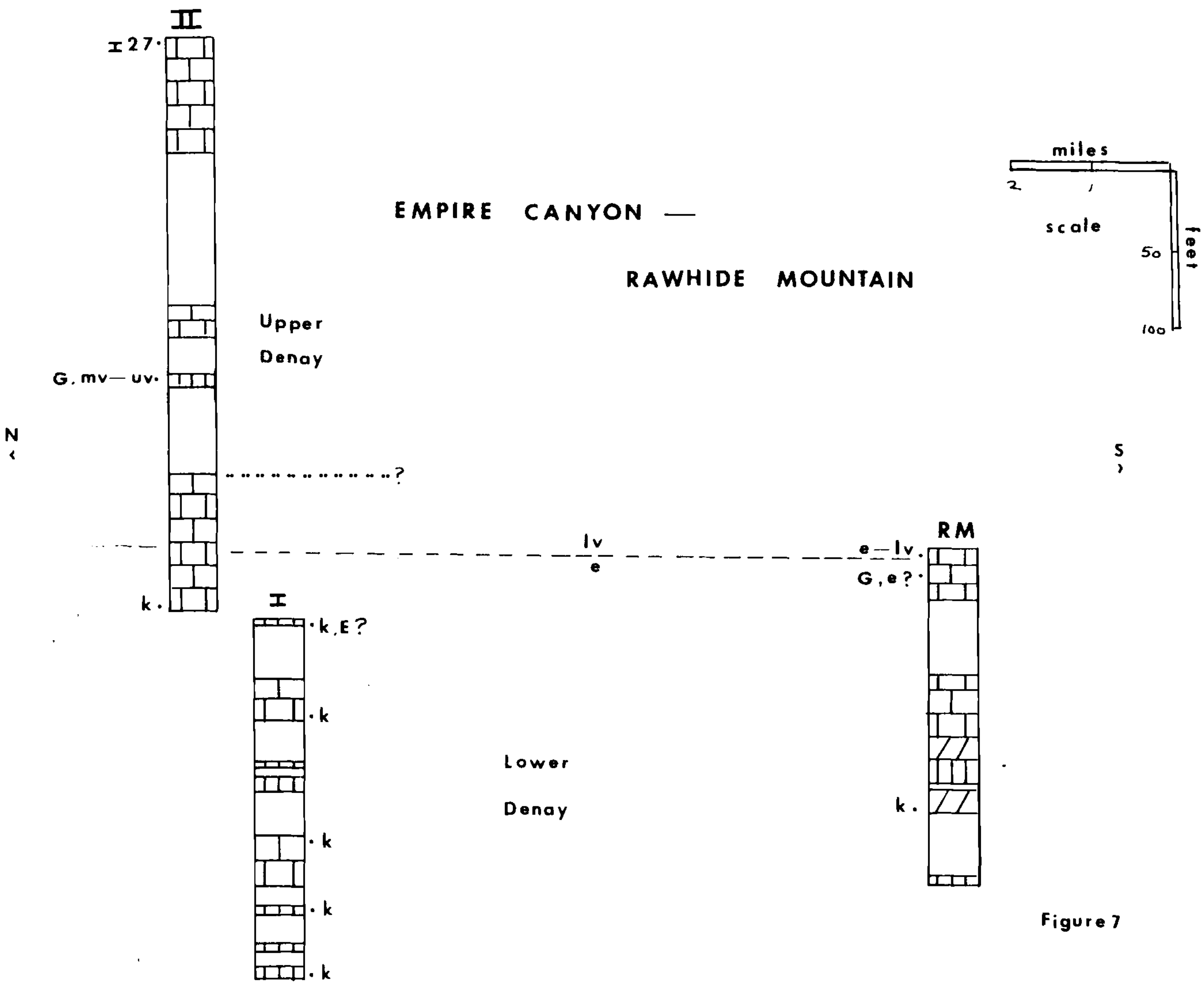
EMPIRE CANYON —
RAWHIDE MOUNTAIN
miles
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Figure 7

rocks could be assigned to the Kinderhookian. Gilbert Klapper of the University of Iowa is currently working on this problem.

Pseudopolygnathus marginatus and *Ps. triangulus triangulus* may also aid in zonal determination. Although rarely found in the *sandbergi-duplicata* Zone in the Midcontinent (Collinson et al., 1970, Figure 5, p. 368), *Ps. marginatus* and *Ps. triangulus triangulus* first appear in the Lower *crenulata* Zone in the Great Basin and Rocky Mountain regions (Sandberg, 1979, p. 100). One or both species were found in WSI-1, WSI-45A, WSI-45B, and WSII-332. Besides *Ps. marginatus*, the turbidite bed from which WSI-45B was collected also contained *G. punctatus* and *Ps. multistriatus*. The overlap of these two species narrows the zonal determination to the upper part of *isosticha*-Upper *crenulata* Zone to the *typicus* Zone (Land et al., 1980, Table 2; Text-figure 8).

The *Ps. marginatus* and *Ps. t. triangulus* in WSII-332 are probably reworked. Conodonts from WSII-216B, lower in the section, are from the *typicus* Zone (Appendix I). The zonal determination was made on the overlap of *G. puncatus* and *G. cuneiformis* (Lane et al., 1980; Klapper, written communication, 1982).

Of the seventeen samples collected from WSIII, only WSIII-4, WSIII-203, and WSIII-515 produced *Gnathodus* at the generic level (Appendix I). The occurrence of *Gnathodus* indicates a zonal assignment no lower than *isosticha*-Upper *crenulata* Zone. If *S. crenulata* has not been reworked then the sample would fall into this zone because *Siphonodella* does not range into the *typicus* Zone.

Tentaculites found on bedding surfaces of the limestone stratigraphically beneath the bedded chert at Warm Springs identify these limestones as Devonian (Bouček, 1964; Plate 1). Random samples collected from this limestone also produced a poorly preserved conodont fauna suggesting a Middle Devonian age (J. G. Johnson, pers. comm., 1982).

Brachiopods recovered from WS 4-82, a fossiliferous wackestone southwest from the town of Warm Springs (Appendix II), indicate a Givetian age ranging from upper Interval 19 to Interval 21 (J. G. Johnson, pers. comm., 1982; Figure 5). The Interval concept was developed in order to get a finer biostratigraphic resolution for the Lower and Middle Devonian (Johnson, 1977). Both conodont and brachiopod zones are composited into distinct faunal "intervals" which incorporate the zonal boundaries of the different assemblage, overlap range, and phyletic lineage zones (Johnson et al., 1980).

The assignment of WS 4-82 to the upper Interval 19 to Interval 21 is based on Subrensselandia which disappears in Nevada after Interval 21 and on Stringocephalus which enters the record in Interval 19 time (Johnson, 1977; Johnson et al., 1980, Tables 12, 13; J. G. Johnson, pers. comm., 1982). Interval 19 straddles the Eifelian and Givetian boundary; however, Stringocephalus has not been discovered in Eifelian age rocks in Nevada. Consequently, WS 4-82 probably correlates to the upper part of sections ECIII and RM and the middle of section ECII.

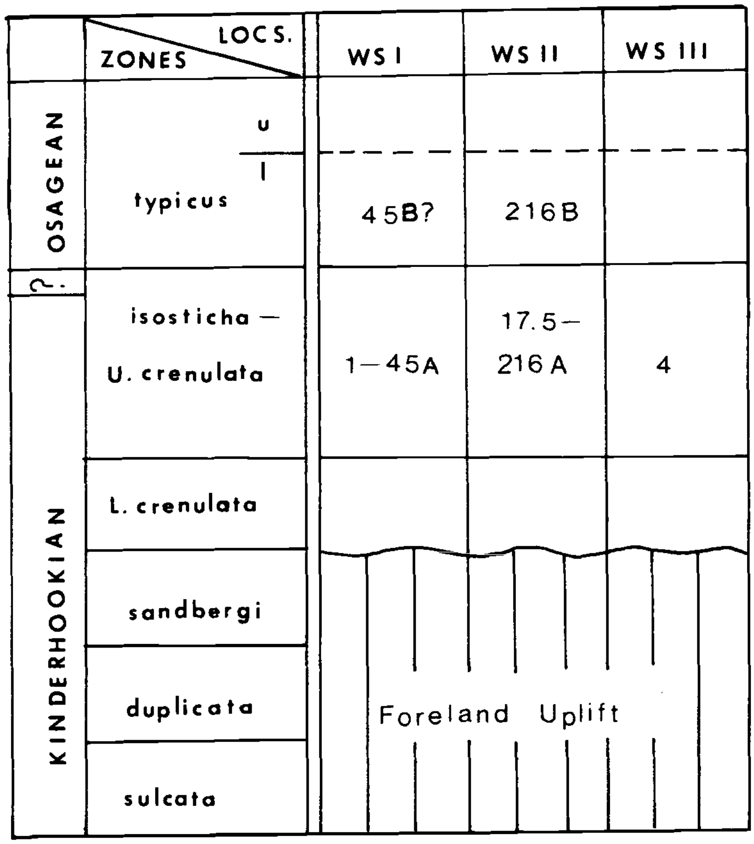

Figure 8: Biostratigraphic chart for the Lower Mississippian at Warm Springs. Numbers in vertical columns refer to footages in the measured sections where diagnostic fauna were found. Conodont zonation from Sandberg (1979); division of <u>typicus</u> into Upper and Lower Subzones from Johnson and Pendergast (1981, Figure 3). Question mark denotes indefinite stage boundaries.

Figure 9: Correlation chart of Lower Mississippian units measured at Warm Springs. Roman numerals at top of vertical columns refer to measured sections located on Plate 1 and aligned in a general west (left) to east (right) pattern. General lithology from Plates 7-9. Covered intervals are unlined. Dots indicate position of zonally diagnostic samples. Letters next to dots refer to the following: K, Kinderhookian; O, Osagean; i-uc, <u>isosticha</u>-Upper <u>crenulata</u> Zone; t, <u>typicus</u> Zone. Note: Section WSIII is broken between 200-530 feet for space considerations.

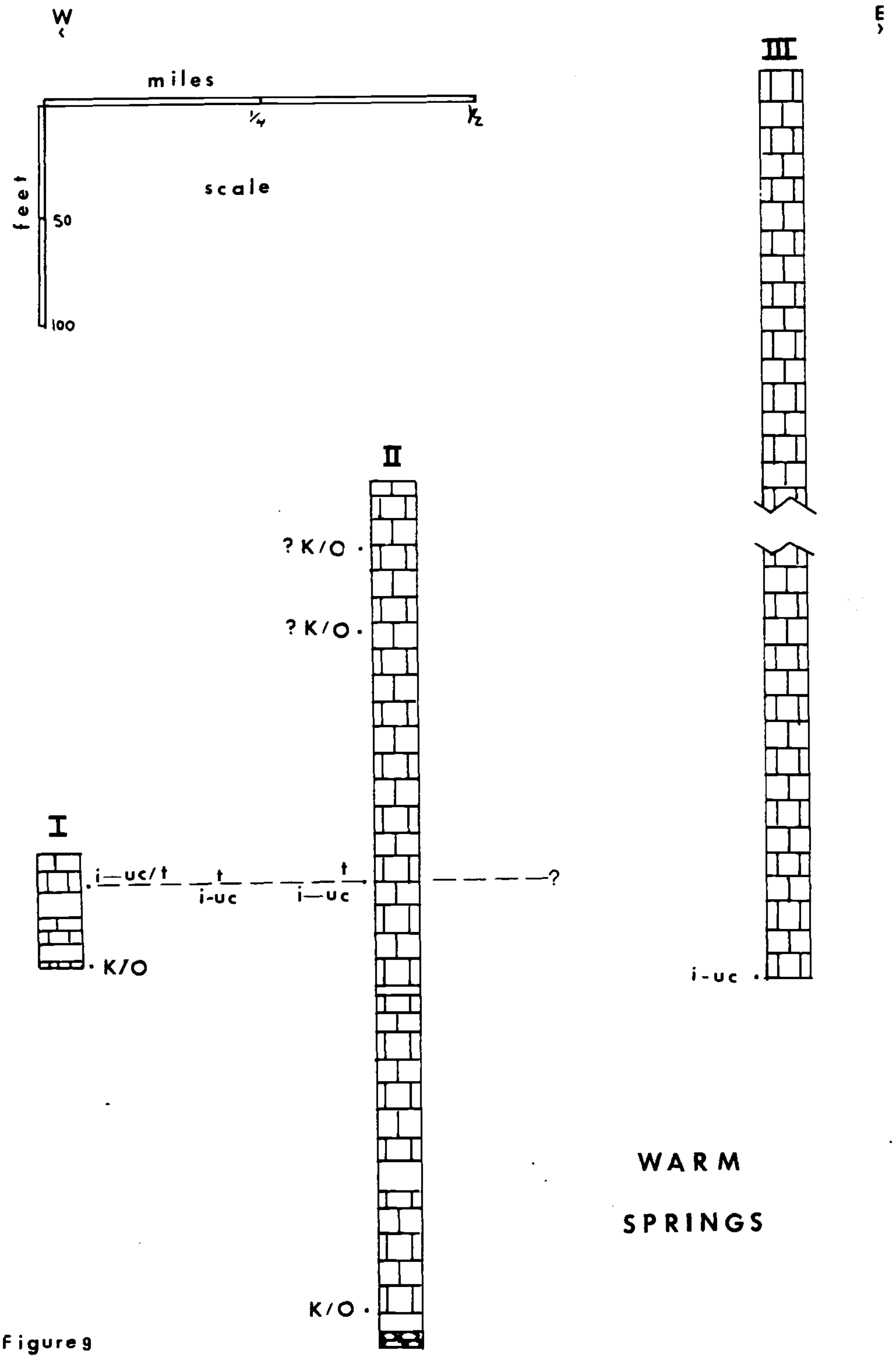
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Figure 9

LITHOSTRATIGRAPHY

Empire Canyon

The McColley Canyon Formation, the oldest lithologic unit measured in Empire Canyon, crops out at the base of ECIII, west of Cine Mountain (Plate 2). The base of this section is bounded by a fault so that only 60 feet (18 m.) of an unknown total thickness is available to be measured. Coarse sucrosic, dolomitized mudstone and wackestone form thin bedded benches and ledges (Figure 10). The rocks weather to a light gray to light yellowish color, but they appear dark gray on fresh fracture. Although some crinoid ossicles can be found locally, the lithology is generally non-fossiliferous. The beds are separated by planar to slightly undulatory contacts. In both hand sample and polished section, the rock appears to be non-laminated. Calcite veins are common on outcrop; micro-quartz veins are evident in thin section.

Although lithologically monotonous at outcrop, the rocks appear in thin section to range from dolomitized mudstone to dolomitized pelletal wackestone. The dolomite is diagenetic in origin and is fabric selective. Scattered throughout the mudstone matrix are clear dolomite-filled voids, minor detrital silt-sized quartz, and minor crinoid fragments. The voids are possibly the result of bioturbation. The matrix, cut by a microstylolite in ECIII-22 (Figure 11), consists of micrite neomorphosed to microspar (Folk, 1965). Porosity in both the mudstones and wackestones is absent.

Micro-pellets in the wackestones of ECIII-5 may approach 40%. Many of the micro-pellets are blurred and indistinct yet appear to be well-sorted and are either in point contact with each other or surrounded by an interlocking, cloudy dolomite groundmass. The micro-pellets are elliptical and approach 0.07 mm. along the long axis. Subangular to subrounded, silt sized, detrital quartz forms a minor constituent, and quartz infills minor fractures.

McGovney (1977) described a "dolomite pellet wackestone" (p. 76) as indicative of the Kobeh Member of the McColley Canyon Formation in the Hot Creek Canyon, a canyon nine miles to the north of Empire Canyon. The outcrop and thin section similarities, as well as geographical proximity, suggest that the basal part of section ECIII is also Kobeh. Biostratigraphically, the dehiscens and gronbergi Zones represent, at Empire Canyon, the upper portion of the Kobeh Member (Figure 5).

Overlying the McColley Canyon Formation is the Denay Limestone. The contact between the Coils Creek Member of the McColley Canyon and the Denay may exist in the covered interval between ECIII-60 and ECIII-90. Thin bedded (7-20 cm.), bench and slope forming blocky mudstone (Figure 12) grades upward into rare wackestone beds and platy weathering, very thin bedded mudstone which is especially evident near the top of section ECIII (Figure 13). The beds appear to have undulatory lower surfaces but planar bedding planes on which tentaculites, both Styliolina and Nowakia (Bouček, 1964), are locally abundant (Figure 14). The light yellowish-gray to dull, steel-gray, weathered surfaces of the beds are pockmarked, and this texture gives a false porosity to the rock; one which is not evident in thin section. During a collecting trip to the northern Antelope Range in June, 1982, similar outcrop characteristics were noted for rock units labeled lower Denay by Trojan (1978). Murphy (1977) also characterized the lower part of the Denay as "laminated, very thin and thin-bedded mudstones

..." (p. 192).

Sedimentary structures are either absent or have escaped my attention due to extensive caliche coatings. Aside from rare laminations seen on a few slabs, internal structures are scarce. No noticeable grading or bottom markings are present. Prominent calcite veins cut the beds; discontinuous, iron-stained limestone nodules and lenses parallel to bedding are found locally.

The wackestones contain unsilicified brachiopods, crinoid stem segments, rare tetracorals, and fenestrate bryozoans weathering out on the surface and floating in a mud groundmass. These beds grade into mudstones within a few centimeters and cannot be traced far laterally. The brachiopod shells do not appear to be in life positions but appear, rather, to have been transported. However, a current direction could not be measured from outcrop. The platy, yellowish-gray recessive mudstone beds (7 mm. - 1.5 cm.) in the upper part of the sections, especially section ECIII, contain abundant tentaculites and a rare, small (1 cm. wide) brachiopod valves.

Upon fresh fracture, the well-indurated rocks emit a strong fetid odor and are colored slate gray to black. Stylolites, with an amplitude less than 1 mm., can be found randomly on outcrop.

In thin section, the mudstone contains between 94-99% matrix. Subrounded, finely disseminated silt particles with corroded edges are present in minor amounts along with sponge spicules, silt-size opaque grains, calcispheres, and brachiopod spines. The matrix has undergone coalescive neomorphism (Folk, 1965), and pressure solution has produced micro-stylolites. These processes and results are discussed later, in the section on diagenesis.

The samples are non-porous. Clear calcite spar has grown within some of the brachiopod spine interiors while others indicate an original micrite filling which has been neomorphosed. Several thin spine walls have been replaced by silica. No geopetal texture is evident. Veins in the mudstone have been filled with clear, calcite cement. The crystals are in sharp contact with the vein walls and exhibit enfacial angles with each other typical of carbonate cement (Bathurst, 1975).

Thin sections of micro-pelletal wackestones found at ECI-226, ECI-228, and ECII-152 consist of 40-70% pellets (Figure 15). Except in ECI-228, where the pellets are poorly defined, they are eliptical to very well rounded, fairly well sorted, and are in point to concave-convex contact. Crinoid osscicles are prevalent and are typically broken and abraded. Other fossils seen in thin section include brachiopod spines, ostracode valves, conodont fragments, and a rare trilobite "shepard's hook" (Figure 15). Subhedral pyrite cubes are less than 1% and can reach 0.2 mm. in size. Sparry calcite has infilled the brachiopod spines and articulated ostracodes in ECI-226. The fossils of ECI-226, which are aligned in a subparallel manner, exhibit some geopetal structure. Slightly undulose, corroded, angular to subrounded, disseminated silt particles are present to about 2%. The matrix in these wackestones is microspar containing relict micrite patches as well as dolomite rhombs, about 0.08 mm. along the long diagonal, floating in the groundmass.

ECIII-91, the bed which produced conodonts indicative of the <u>australis</u> Zone and which seems equivalent in outcrop to the Denay of Potter (1975, sample EP 105.9) and Trojan, proved in thin section to be a pelletal grainstone (Figure 16). Silt to very fine sand sized pellets compose about 70-80% of the sample. These are well-rounded and are in point contact as well as long contact. Surrounding the pellets and partially supporting them is a matrix of calcite spar. Minor suturing between pellets indicates an influence from compaction which may have biased the original ratio of pellets to matrix.

Besides pellets, quartz particles, fossil fragments, and chert are present. The fine sand to silt size quartz (0.08 mm.-0.22 mm.) is rounded to angular, only moderately sorted, with slightly undulose extinction except for the larger particles which display a straight extinction. The fossils include brachiopod shell and crinoid (echinoderm) fragments as well as a possible Dasycladacian algae. Chert is only in trace amounts and occurs as a replacement.

Above a 95 foot (29 m.) covered interval in section ECII, a bryozoan-bearing, stromatoporoid, intraclast wackestone crops out which has not been dated (Figure 6, Plate 4). Lithologic similarity to Upper Denay rocks described by Trojan (1978, p. 85) and the superposition of the wackestones above the Lower Denay limestones found at the base of ECII suggest a lithologic correlation with the middle portion of the Upper Denay units in the northern Antelope Range. Lying above the uppermost part of the Middle varcus Subzone to Upper varcus identified from ECII-152, these beds may be correlative with the Lower dengleri Subzone limestones which crop out above a covered interval in the upper part of Trojan's section V (Trojan, 1978; Johnson et al., 1980, Figure 4). Thin to thick bedded, bench forming limestones contain stromatoporoid clasts which weather out of a light gray to yellowish gray mud matrix. Upon fresh fracture, a fetid odor is emitted.

The concentric laminae and radial pillars of the stromatoporoids are visible in hand sample and thin section (Figure 17). The stromatoporoids have not been dolomitized. Zooecia in the bryozoan fronds have been replaced by interlocking blades of calcite. However, angular clasts of dolomitized micrite dominate the wackestones along with 5 mm. clasts of sparry calcite in a radiaxial fibrous mosaic. The matrix has undergone neomorphism to microspar. Fine monocrystalline quartz, brachiopod shell fragments, and replacement chert are present in minor or trace amounts.

Stratigraphically above this wackestone in the eastern end of Empire Canyon, although not found in the measured sections, is a yellowish weathering limy finely sucrosic dolomite containing, locally, deeply weathered, silicified horn corals. These beds may prove equivalent to the Middle Devonian dolomite of Morey Peak (Potter, 1975). Thin sections from the limy dolomite of Empire Canyon reveal dolomite clasts with sutured boundaries set in an argillaceous matrix with abundant dolomite rhombs. Samples collected from Potter's Middle Devonian dolomite revealed, in thin section, a limpid dolomite surrounding micrite patches which resembled micro-pellets (about 0.35 mm. diameter) with a vague outline. Crinoid columns were also present. Unfortunately, the samples have failed to yield any conodonts so the age of the Middle Devonian dolomite of Morey Peak remains an enigma.

Only three of the "numerous beds of fine-grained quartzite" reported by Quinlivan and Rogers (1974) to be present in upper Empire Canyon were found on the northwest flank of Cine Mountain. The quartzite weathers to a pinkish color, but it is a blue-gray color on fresh fracture. The clean, equigranular sedimentary quartzite contains vugs lined with drusy quartz crystals.That the quartzite units can be defined as "beds" is questionable. Although they follow the general north-to-south trend of the enclosing dolomite beds, the quartzite beds are highly broken and fractured and bedding surfaces are not to be trusted. These "beds" may more appropriately be termed "lenses". The "Nevada Formation" of Cine Mountain (Quinlivan and Rogers, 1974) has been altered to light and dark bands of dolomite.

The measured sections ECII and ECIII are terminated by fault breccias which contain quartz grains, chert clasts, and sedimentary rock fragments. In thin section, the quartz grains are clear, well-rounded to subrounded and poorly sorted. Overgrowths are absent and extinction is undulose. Chert clasts are the largest and most abundant constituent. Two generations of chert formation are indicated by the two distinct grain sizes forming the clasts and the micro-veining which cuts the finer grained chert. Silica-cemented sedimentary rock fragments and detrital grains of semicomposite, slightly stretched quartz (Scholle, 1979) are minor components. The breccia is cemented with hematite although calcite crystals have grown in and destroyed original breccia porosity.

Rawhide Mountain

Samples collected from section RM are in the interval kockelianus to ensensis Zone and are lithologically similar to the upper Lower Denay rocks described in Empire Canyon. The first 95 feet (29 m.) consist of alternating units of bench-forming lime mudstone and dolomite and recessive-weathering shaly lime mudstone. Above the highest dolomite bed, the remaining 122 feet (37 m.) grade upward from a mudstone into packstone and grainstone. The section and ridge are then truncated by a major fault (Plate 2).

Tentaculites, deeply weathered, but appearing to be

Figure 10: Thin-bedded, dolomitized McColley Canyon mudstone at Section ECIII-5.

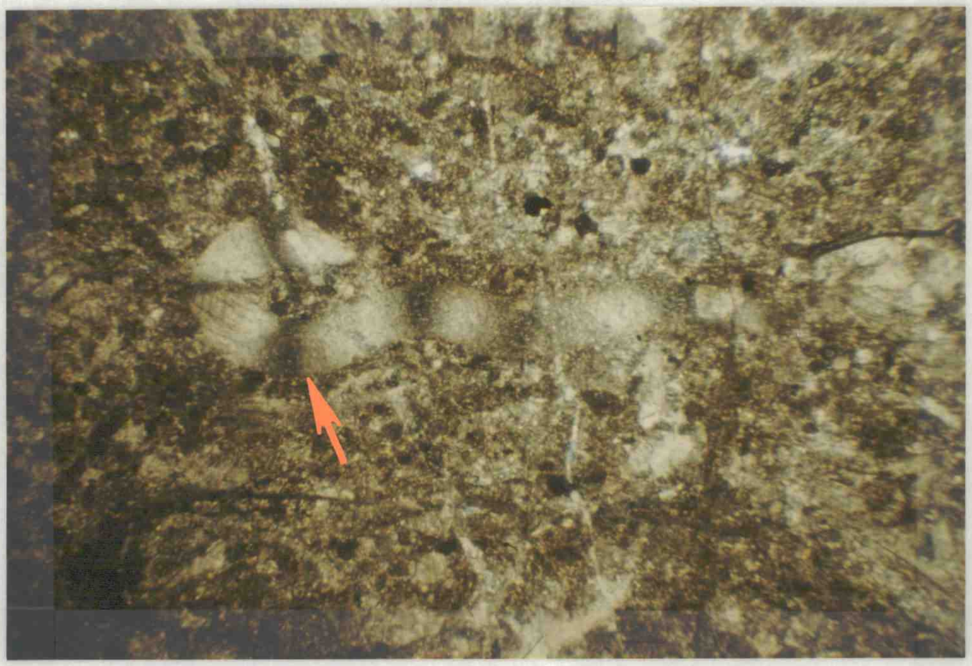
0 1 mm

Figure 11: Photomicrograph of micro-stylolite cutting dolomitized mudstone of ECIII-22. Note absense of porosity, large dolomite crystals next to stylolite, and anhedral dolomite in matrix. Plane light. Sample ECIII-5 is similar but contains an increase in pellets.

Figure 12: Thin-bedded, Denay mudstone ninety feet from the base of section ECI. Scale is two inches long.

Figure 13: Platy weathering, very thin-bedded mudstone near top of section ECIII (ECIII-200), Denay Limestone.

Figure 14: Tentaculites found on bedding surfaces. Locally, rocks range from tentaculitic mudstone to tentaculitic wackestone.



0 1 mm

Figure 15: Photomicrograph of micro-pelletal wackestone found at section ECI-226, Denay Limestone. Note trilobite "shepard's hook" (arrow) in center of photo and point contact of grains without suturing. Matrix is microspar with "dirty" micrite patches and pellets. Crossed nicols.

Figure 16: Photomicrograph of pelletal grainstone found at ECIII-91. Note brachiopod shell fragments in amongst pellets (arrow). Sample ECIII-91 is Denay Limestone. Plane light.

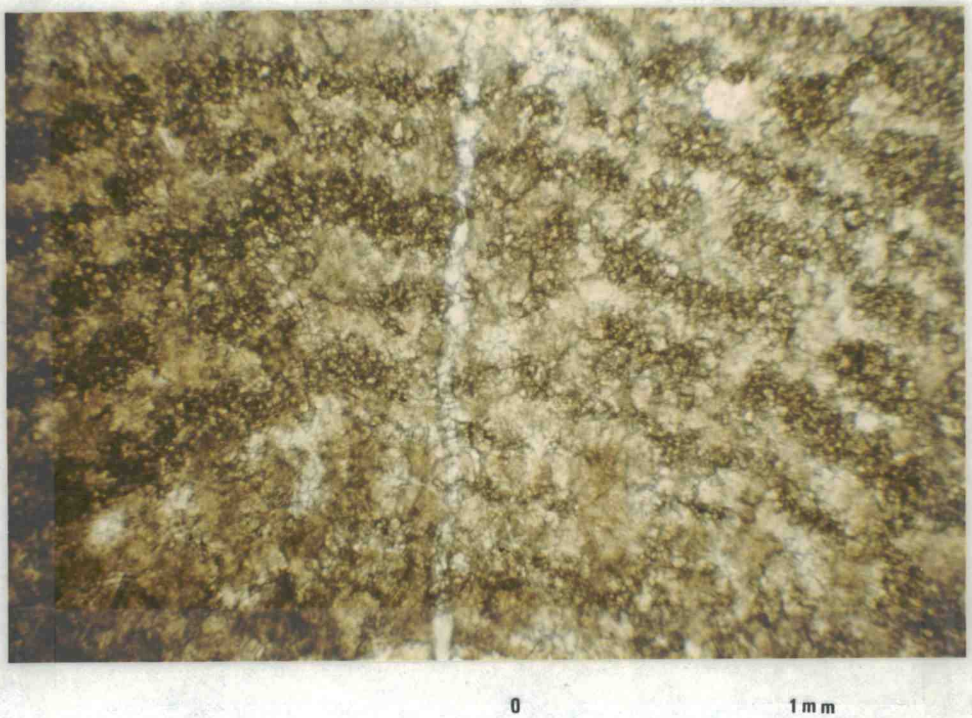


Figure 17: Photomicrograph of stromatoporoid clast from ECII-365 showing typical concentric laminae. Bryozoans from the same sample have been replaced by calcite, and dolomite clasts are common. Plane light.

mostly Styliolina with some Nowakia, litter the bedding surfaces of the mudstone in the lower unit. The mudstone weathers to a yellowish gray, but it is a dark gray on fresh fracture. Bedding is thin (7.6 cm. or 3 in.) to very thin (1.27 cm. or 0.5 in.) with planar contacts and shaly partings on the very thin beds. Thin laminations are present locally as well as micro-stylolites, with amplitudes averaging 1 mm., which trend oblique to bedding. On outcrop, the rock appears porous, but thin sections reveal a tight, non-porous sample. The matrix is 94-99% micrite with the remainder composed of detrital clay films along stylolites, minor amounts of hematite, and minor amounts of fragmented and whole brachiopod spines about 0.28 mm. in diameter.

Two units, 15-20 feet (4.5-6 m.) thick, of dolomite are sandwiched between tentaculitic mudstone. The lower unit is a fossil-bearing, dark gray, sucrosic dolomite, thinly bedded and finely laminated. The beds appear to be finely graded. Graded intervals 2-3 mm. to 1 cm. thick lie within each 9 inch (23 cm.) bed. Crinoid ossicles and small (5 mm.), plicated, disarticulated brachiopod valves occur isolated in the matrix. Besides being disarticulated, pedicle valves were concave up on bedding planes. This orientation suggests little current activity although the paucity of shells and the small size of the valves make this evidence inconclusive.

The laminations are faintly seen on a polished slab of RM-49, but neither the laminations nor the grading can be seen in thin sections. In thin section the lower unit is a dolomitized mudstone. The matrix consists of small (0.062 mm. - 0.25 mm.), equant, interlocking, crystals of dirty dolomite with relict micrite patches. Large (0.58 mm.) clear crystals of secondary dolomite interlock with enfacial angles in veins and fractures.

Some crystals of dolomite spar are zoned. Chert is also a replacement mineral along some vein-matrix contacts although very minor compared to dolomite. Abundant sponge spicules were recovered in sample RM-49 upon dissolution in formic acid.

The upper dolomite unit is non-fossiliferous, finely sucrosic, and light gray on a weathered surface. Grading is not present. A sample was not collected from this unit so I am uncertain as to whether this dolomite is primary or secondary. However, I think it is reasonable to assume a close proximity to the Simonson Dolomite, a primary dolomite during late Eifelian time (Johnson and Sandberg, 1977). This proximity may have contributed to dolomitization at a saltwater/freshwater lens contact as described by Dunham and Olson (1978) for other areas of Nevada (see section on depositional environments).

From 95 feet to 135 feet (29 m.-41 m.) bench forming, tentaculite-bearing mudstone again crops out. Clay seams and micro-stylolites are present in thin sections and may be the result of overburden stress (Wanless, 1979). Patches of calcite spar floating in the matrix do not resemble a skeletal outline and suggest recrystallization by porphyroid neomorphism. Clay stringers resulting from compaction may have given rise to the faint, very fine laminations seen in polished slabs. A few Nowakia were recovered from the residue after dissolving sample RM-98 in formic acid. Sponge spicules and scraps of thin-shelled brachiopods were also recovered from the fine residue.

Occurring at 183 feet (56 m.) are alternating mudstone and crinoidal wackestone beds. The platy weathering, yellowish gray mudstone is laminated (3 mm.) whereas the more resistant, dark gray wackestone is thin-bedded (20 cm.) and contains graded beds spaced about

1.5 cm. apart (Figure 18). Contacts between the mudstone and wackestone beds are always covered. In thin section, this non-porous rock is a dolomitic bioclast-quartz packstone (Figure 19). Subangular to subrounded, fairly well-sorted, silt to very fine sand sized disseminated quartz with undulose extinction makes up about 15% of the constituents. Fossils, including brachiopod shell fragments and crinoid stem segments, compose an additional 10%. Equant dolomite rhombs to 20% float in a micrite matrix and appear to have relict micrite in the center of the crystals. Some rhombs are in point contact with each other.

At 195 feet (59 m.), a noticeable increase in fossil content occurs (Figure 20). The thin-bedded brachiopod-crinoid wackestones are resistant and are fractured because of close relationship to the fault. When sample RM-200 was dissolved in formic acid, brachiopods, bryozoans, and crinoid ossicles remained in the screens. A thin section of RM-200 reveals ostracodes, conodonts, pellets, quartz, chert, and dolomite rhombs in addition to the aforementioned fossils. The thin section also shows a fining upward from a fossiliferous, pelletal grainstone to a dolomitic, fossiliferous, pelletal wackestone.

An articulated brachiopod shell greater than 5.5 mm. occupies most of the lower part of RM-200 along with articulated ostracode valves, crinoid ossicles, pellets, minor silt-sized to fine sand sized quartz, replacement chert and quartz, and a possible conodont fragment. The bivalves display geopetal textures. The matrix is calcite spar. The large shells disappear in the upper portion and pellets, crinoid fragments, and dolomite rhombs increase in abundance. The matrix is no longer spar, but is not readily visible. Rather, pellets of micrite are tightly packed which gives this upper part

Figure 18: Graded bedding and laminations found 185 feet from the base of section RM. Except for the rare beds containing macro fossils, this bed is as coarse as any found in Empire Canyon or Rawhide Mountain. Captain America, used for scale, is two inches tall.

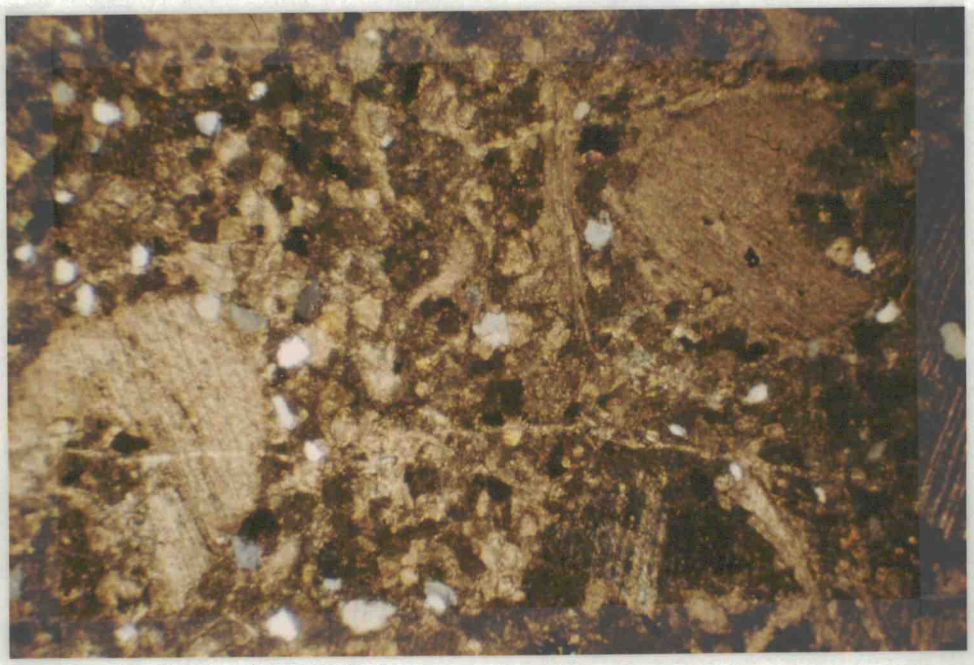


Figure 19: Photomicrograph of the bioclast-quartz packstone at RM-185. Crinoid ossicles, brachiopod shell fragments, quartz particles (clear), and subhedral dolomite crystals are present. Overgrowths are absent. Crossed nicols.

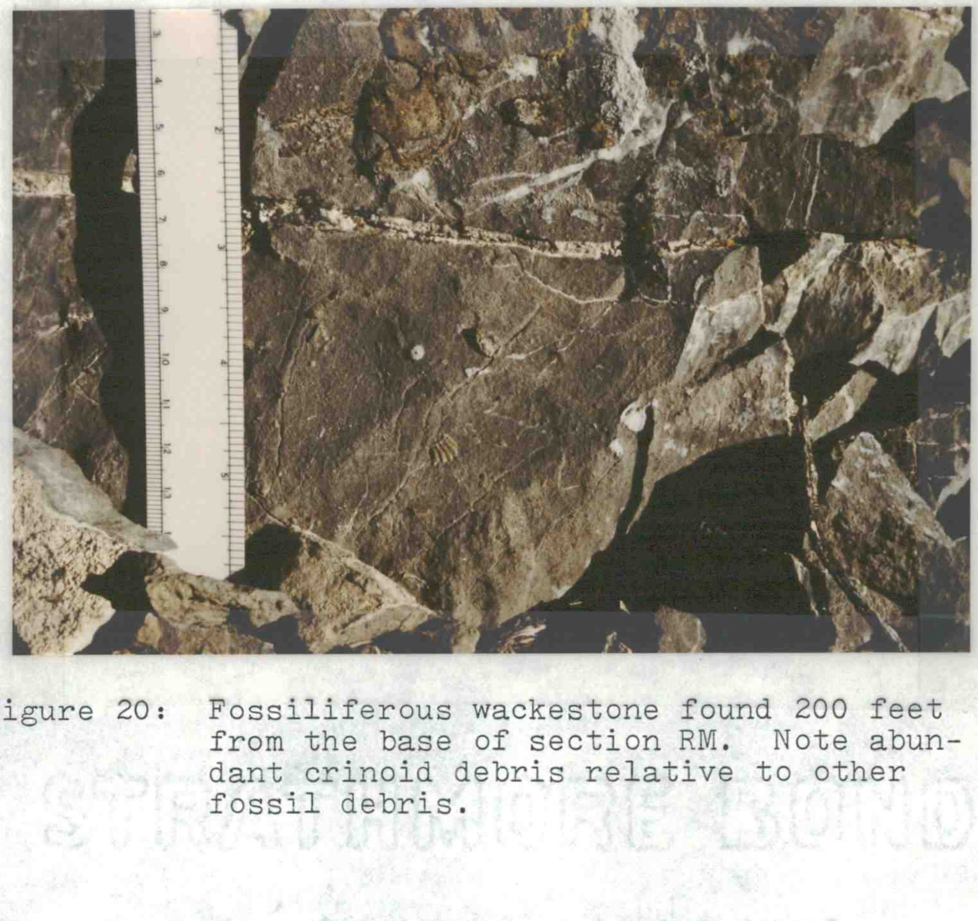


Figure 20: Fossiliferous wackestone found 200 feet from the base of section RM. Note abundant crinoid debris relative to other fossil debris.

a possible packstone texture.

The well-rounded, equant (0.22 mm.) micrite pellets compose about 20% of the overall components while fossil fragments amount to about 10%. The well-sorted, silt-sized quartz particles are angular to subrounded with corroded rims and undulose extinction. They compose about 3% of the total. Dolomite rhombs are scattered and amount to only 1% of the total constituents. Two to three percent is replacement chert and quartz; the rest is matrix, be it spar or micrite.

Warm Springs

Over 730 feet (223 m.) of very fine-grained Mississippian limestone were measured at Warm Springs (Figures 9, 21). Although very thin (4 cm.) chert beds are interbedded with very thin (2.5 cm.) to thin (8 cm.) beds of lime mudstone in the first 50 feet (15 m.) of WSII, 1-4 foot (0.3 m.-1.2 m.) beds of bench-forming, argillaceous, non-laminated mudstone compose most of sections WSI and WSII and the entirety of WSIII where bedding may be as much as 6 feet (1.8 m.) thick (Figure 22). This mudstone is gray on a dry, weathered surface and, when cracked, emits a fetid odor from a dark gray, dense, aphanitic, non-fossiliferous matrix. Laminations and thin, shaly beds are present only in the first 20-30 feet (6 m.-9 m.) of sections WSI and WSII.

The texture of the bedding surfaces is characteristically pock-marked, warty, and ragged. Calcite veining and minor faulting are common. Contacts between beds are undulatory to planar. The apparent porosity seen at outcrop is deceiving as thin sections reveal a non-porous, compact rock.

Aside from minor laminations in WSII which are commonly discontinuous and contorted, sedimentary structures are scarce. Undulatory bedding reflecting possible ripple marks has a wave length from about 3-4 feet (1-1.2 m.) to 3-4.3 inches (8-11 cm.) between symmetrical crests in the mudstone (Figure 23). Contacts become more planar up-section.

Suggestions of graded bedding occur in the shaly limestone interbeds of WSI and WSII. The contact of the shaly beds with the overlying thin mudstone bed is sharp, but the contact is gradational with the underlying mudstone bed. Occasionally, a more coarse textured, thin limestone bed containing ooid-like grains and undefinable skeletal fragments is found in sharp contact with an underlying fine-grained mudstone bed and grades upward into another mudstone bed. The coarse bed is crystalline on fresh fracture. This combination is found, for example, 216 feet (66 m.) from the base of WSII (Figure 24). The coarse layer contains, in hand sample, subrounded to subangular clasts less than 1 mm. consisting of quartz, mudstone ripups, and possible fossil shell valves. The clasts are aligned subparallel to bedding in a slate gray matrix stained by surficial oxidation, but the layer lacks any distinct grading. Perhaps 40% of the hand sample is composed of clasts.

In thin section, this coarse textured layer, which yielded conodonts of the typicus Zone, is a detrital quartz packstone (Figure 25). The angular to well-rounded, inclusion free, quartz grains are silt to very fine-sand size, display both strong undulose (angular grains) to slightly undulose (rounded grains) extinction, and compose about 15% of the rock. The quartz is matrix supported and is not overgrown. Two or three grains of polycrystalline quartz also are present. Other clasts include 5% opaque, hematitic blotches and pore fillings; 2% anhedral grains of hematite-stained chert; 3% broken ostracode valves rimmed by silica and spheres of calcite fragmented in part and filled with pseudospar; and a trace of brachiopod spines and deeply weathered crinoid fragments. An interlocking hodgepodge of microspar and pelletal size micrite patches supply roughly 75% of the total mineralogy. Any porosity that was originally present has since been destroyed by secondary hematite and mosaic calcite pore fillings and neomorphism.

Bordering these local packstones are carbonaceous, argillaceous, spicular wackestones (Figure 26). Monaxial spicules are subparallel to bedding and make up 15 to 20% of the wackestone. Spheres of calcite 0.21 mm. to 0.07 mm. in diameter also are prevalent to 10% and some suggest calcite replacement of radiolarians. However, the original structure of both the problematic calcispheres and spicules has been replaced by calcite, and pseudospar has infilled the center of the calcispheres. In contrast to the packstone, the wackestone contains a dearth of silt-sized quartz, chert, and pelletal micrite patches. The matrix is micrite with some patchy microspar.

Mottled bedding, suggesting bioturbation, is locally present lower in the sections of both WSI and WSII. Sixteen feet (4.8 m.) from the base of WSII grazing trails of the trace fossil, Nereites, were found on the bedding surface of shaly interbeds. The bed above the Nereites bed contains a mottled lower section grading upward into a laminated layer above a surface deformed by small-scale sediment loading and micro-scours. Thin sections reveal the lower portion to be an argillaceous, spicular mudstone. Spicules are 5% of the rock and calcispheres make up another 2%. Secondary hematite rims and subhedral pyrite grains compose another 2% along with silt grains being encroached by carbonate. Helminthoidea grazing trails can be seen scattered in a microspar matrix.

A micro-scour surface (Figure 27) separates the mudstone from a very thin carbonaceous, silty, pelletal grainstone grading upward into a spicular-calcisphere mudstone. The pellets and silt particles are in point contact and do not express any intense compaction. The spheres of calcite are aligned in parallel lineations composed of micrite. Burrows are absent in this upper part.

A similar carbonaceous, biomicropelletal grainstone is also found 17 feet (5 m.) from the base of WSI. Micropellets, in point contact, compose 40% of the rock while crinoid fragments, scattered ostracode valves, conodont fragments, brachiopod spines, spheres of calcite, and calcite-filled sponge spicules constitute 8% to 10%. Although well-sorted, most of the pellets do not have sharp outlines and the close packing may, in fact, represent an initial stage in the formation of "structure grumeleuse", a structure discussed more fully in the section on diagenesis (Folk, 1965; Bathurst, 1975). Subrounded to angular, silt-sized quartz grains are present to 5%. Carbonaceous films are concentrated along microstylolites. Calcite spar has filled in around the pellets and fossil fragments.

This variety of mudstone to grainstone is lost higher in sections WSII and WSIII. Here, thin sections reveal argillaceous, spicular mudstones and wackestones (Figure 28). The spicules are 20 to 50 microns in diameter and aligned parallel to bedding. They are monaxons which have been replaced by calcite. Spicules range from 5% to 15%; spheres of calcite contribute 3% to 7%. Silt particles are minor to absent. Neomorphism has begun to alter the micrite matrix to microspar.

Wise (1976), mapping in the Monitor Range approximately 50 miles (80 km.) N10W from Warm Springs, described Mississippian "Camp Creek equivalent" rock units with a lithology which closely matches the Mississippian rocks at Warm Springs. However, Wise found repeated cycles of "graded, laminated, and non-graded beds two to four feet thick, separated by two to six inch layers of black calcareous mudstone " (1976, p. 86) which are also present farther north in Elko County in the Upper Camp Creek unit of Pendergast (1981). These cycles do not occur at Warm Springs.

Although the sedimentary structures are not well defined at Warm Springs, the Mississippian rocks still suggest a lithologic equivalency to the Camp Creek. They correspond favorably to the "black, silty, argillaceous, carbonaceous, spicular carbonate mudstone" description of the Lower Unit described by Pendergast (1981, p. 48-54). Camp Creek rocks in both Elko County and Nye County have been dated as Kinderhookian although the Warm Springs rocks may range across the Kinderhookian-Osagean boundary.

The Mississippian limestones conformably overlie undulating beds of well indurated, radiolarian cherts. The beds are 2-4 inches (5-10 cm.) thick (Figure 29), and the chert, dark maroon on a weathered surface, is black when cracked open. The first 7 feet (2 m.) of section WSII is composed of this bedded chert, and thin sections made from samples 4 feet (1.2 m.) above and 14 feet (4.2 m.) below the base of WSII disclose siliceous microspicules and radiolarians to 0.2 mm. diameter (Figure 30). These microfossils are set in a dark, hematite-stained matrix. The larger radiolarians are not fragmented and show little evidence of compaction. Porosity is absent. Secondary fractures, some measuring 0.4 mm. wide, have been filled with chert and very fine quartz crystals. The clear quartz exhibits plane, intercrystalline boundaries and two growth stages in the wider fractures.

Twenty feet (6 m.) of black, well-bedded chert followed by 110 feet (33.5 m.) of shales, sandstones, and interbedded black chert were measured by Wise (1976) south of Dobbin Summit, Monitor Range. The radiolarian chert of Warm Springs resembles the description of the first 20 feet (6 m.) of black chert. This chert is extensive on the southern flank of the Hot Creek Range where Highway 6 bends to the west from Warm Springs en route to Tonopah, Nevada. Outcrops, however, are generally recessive beneath the more prominent Mississippian limestone which caps the peaks. Consequently, the thickness of the Warm Springs chert is unknown. An order of magnitude apparently exists between the 20 feet (6 m.) measured in the Monitor Range and the thickness of the Warm Springs chert.

The chert beds are deformed into tight and similar, isoclinal folds near fault contacts, especially the beds in the draw south of the base of section WSII (Figure 31). These drag folds unfold away from the faults and return to the gentle warping and undulations which are characteristic of the bedding throughout the area.

Aside from this local folding, the chert expresses a regularity in bedding and a conformable contact with both the superjacent Mississippian limestone and the subjacent Devonian limestone.

Recessive-weathering, undifferentiated Devonian limestone wraps around the southern end of the Hot Creek Range in isolated segments. The limestone varies from a weathered, light gray, very thinly bedded tentaculite mudstone to thin to massive interbeds of mudstone and medium gray, coralline-crinoid-brachiopod wackestone. The mudstone, in thin section, contains minute dolomite crystals floating in an argillaceous micrite. The wackestone, heavily veined with calcite, is a pelletal bioclastic wackestone containing broken and disarticulated, silicified fossil fragments. Red laminations of hematite are locally visible in thin mudstone beds.

Behind the town of Warm Springs, an isolated block of fractured Devonian limestone crops out. The limestone is capped by a massive, dark brown, aphanitic jasperoid as defined by Lovering (1972; Plate 1). Jasperoid deposits separate the Paleozoic strata from the younger volcanics and form prominent, blocky cliffs which stretch along the southern, western, and eastern flank of peak 6948. In thin section, the jasperoid has a hematite-stained chert matrix enclosing trace amounts of disseminated silt particles.

The jasperoid bodies in Nevada are considered to be late Mesozoic to early Cenozoic in age (Lovering, 1972). The Manhattan mining district in the southern part of the Toquima Range, about 45 miles (72 km.) N60W from the town of Warm Springs, contains probable Tertiary age jasperoid deposits (Lovering, 1972). The age of the jasperoid at Warm Springs was not determined but probably falls within these age constraints.

N ←

Figure 21: Panorama of Warm Springs, Mississippian limestone along ridge west of the town of Warm Springs, Nevada. View to the east of the radio tower located at 6485 feet. Limestone beds dip in a general northerly direction along section III. The relative positions of Sections I, II, and III are plotted.

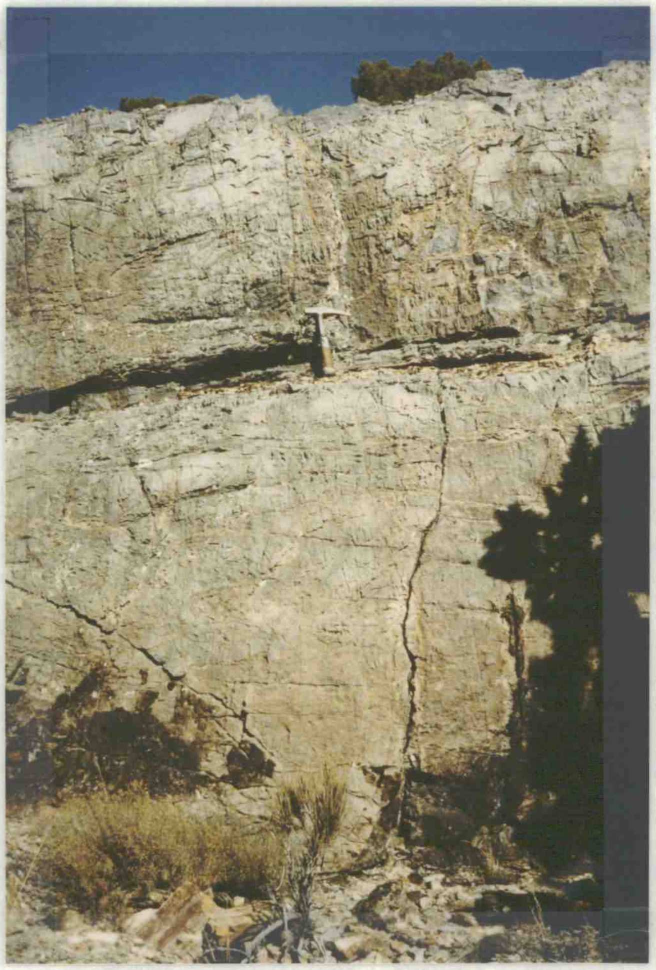
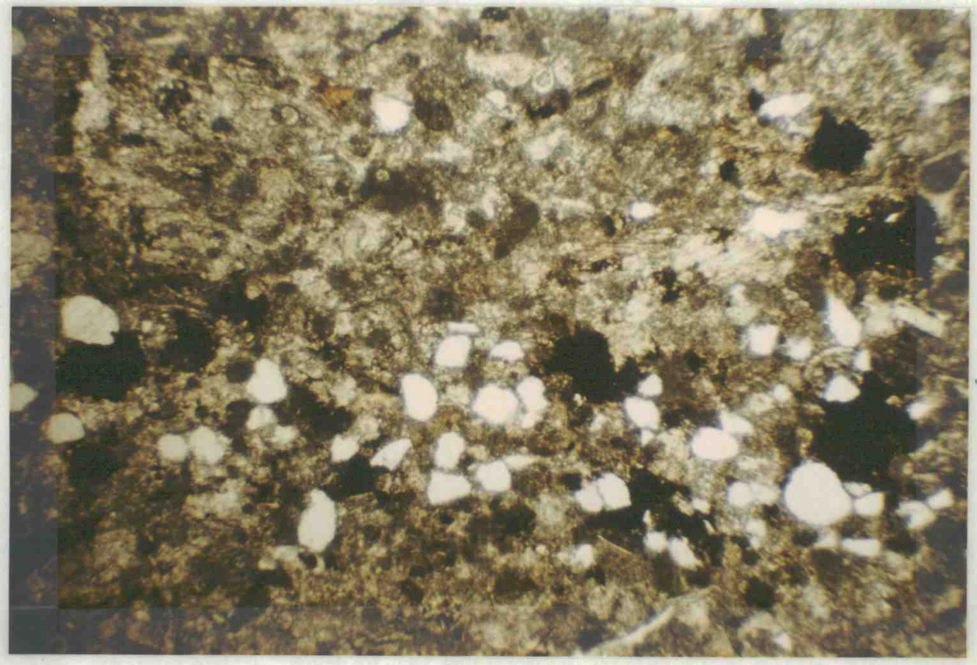


Figure 22: Typical, thick beds found in Section WSIII. Unit is a lime mudstone. Note rough texture to outcrop caused by solution weathering.

Figure 23: Ripple marks (arrow) four feet from the base of WSII. These are not common up-section.

Figure 24: Allodapic limestone bed found 216 feet from base of WSII. Coarse (relatively) bed stained with hematite sits atop Captain America's head and is in a sharp contact with the underlying, gray mudstone.



0 1 mm

Figure 25: Photomicrograph of sample WSII-216B, the coarse bed of Figure 24. Limestone is a detrital quartz (colorless grains) packstone. Plane light.

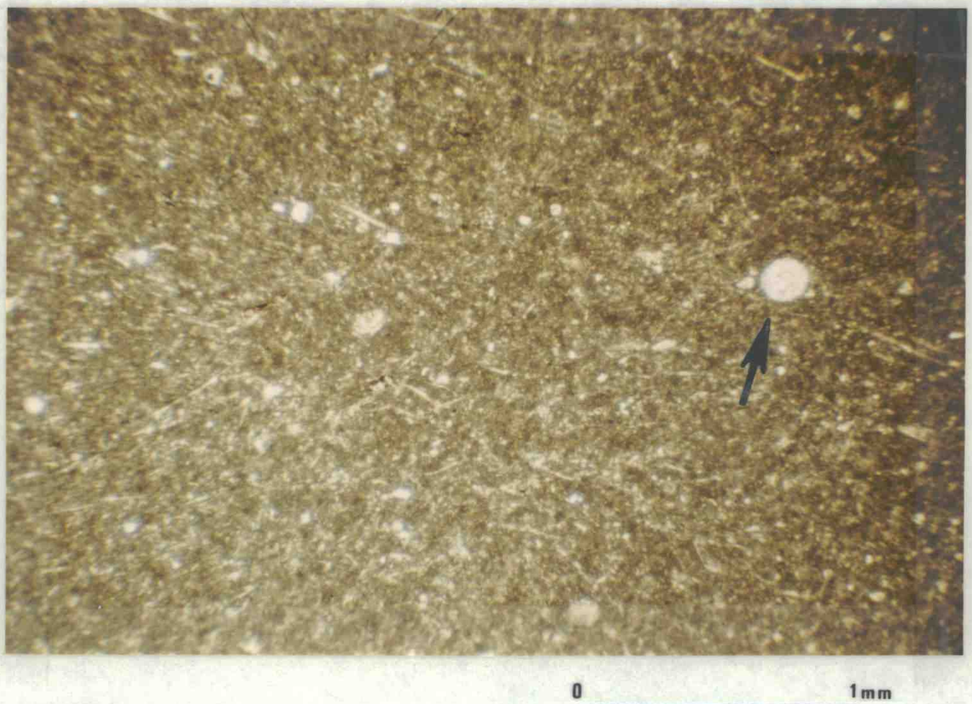


Figure 26: Photomicrograph of sample WSII-216A,- a carbonaceous, argillaceous, spicular wackestone found beneath sample WSII-216B. Note parallel alignment of spicules and possible radiolarian test in right center (arrow). Plane light.

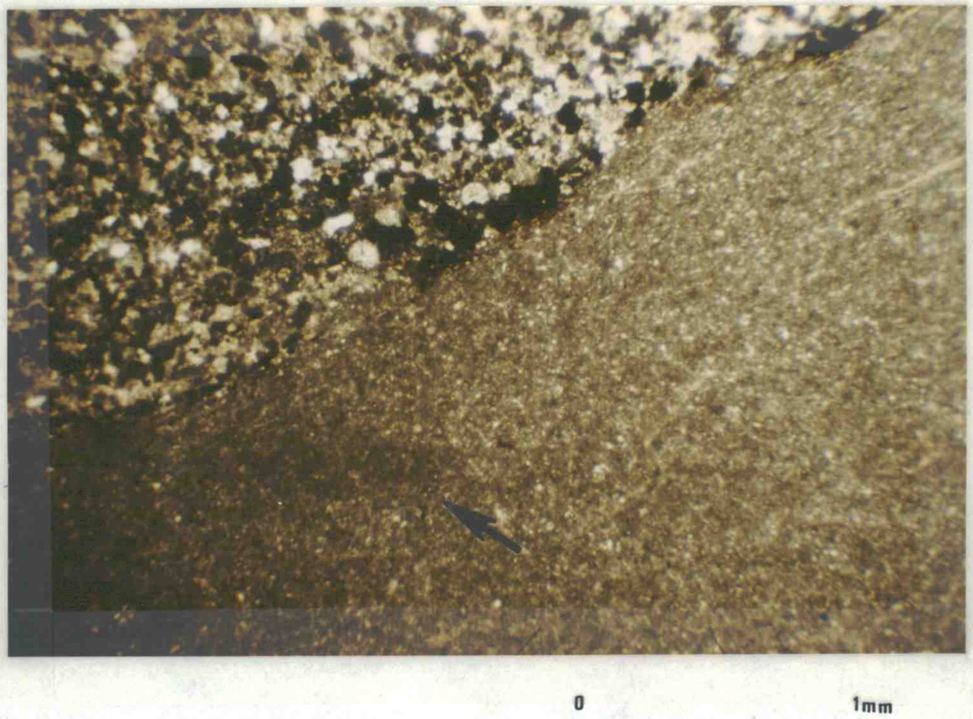


Figure 27: Photomicrograph of micro-scour surface separating the spicular mudstone (lower) from a carbonaceous, silty, pelletal grainstone (upper) in WSII-17.5. Possible Helminthoidea trace in mudstone, left center (arrow). Quartz grains are colorless. Plane light.

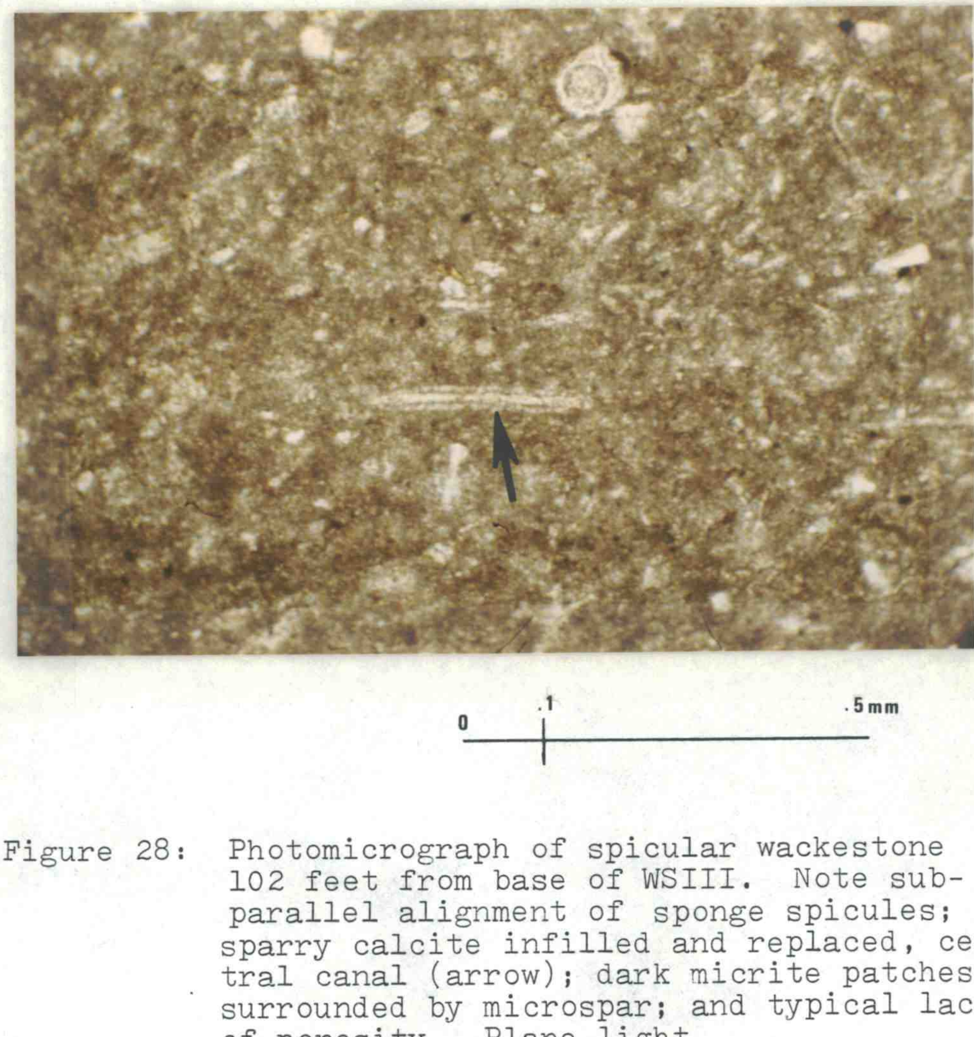


Figure 28: Photomicrograph of spicular wackestone 102 feet from base of WSIII. Note sub-parallel alignment of sponge spicules; sparry calcite infilled and replaced, central canal (arrow); dark micrite patches surrounded by microspar; and typical lack of porosity. Plane light.

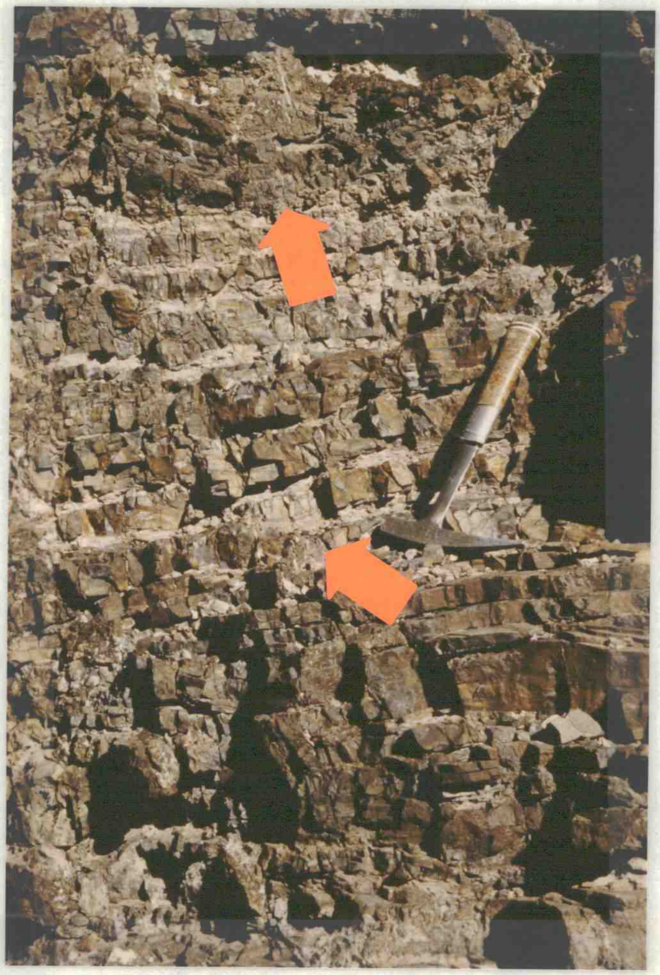


Figure 29: Warm Springs chert beds underlying the Mississippian limestones. Note regularity of bedding with some lens pinch-outs near hammer head and the possible channel in the upper part of the photograph (arrows).

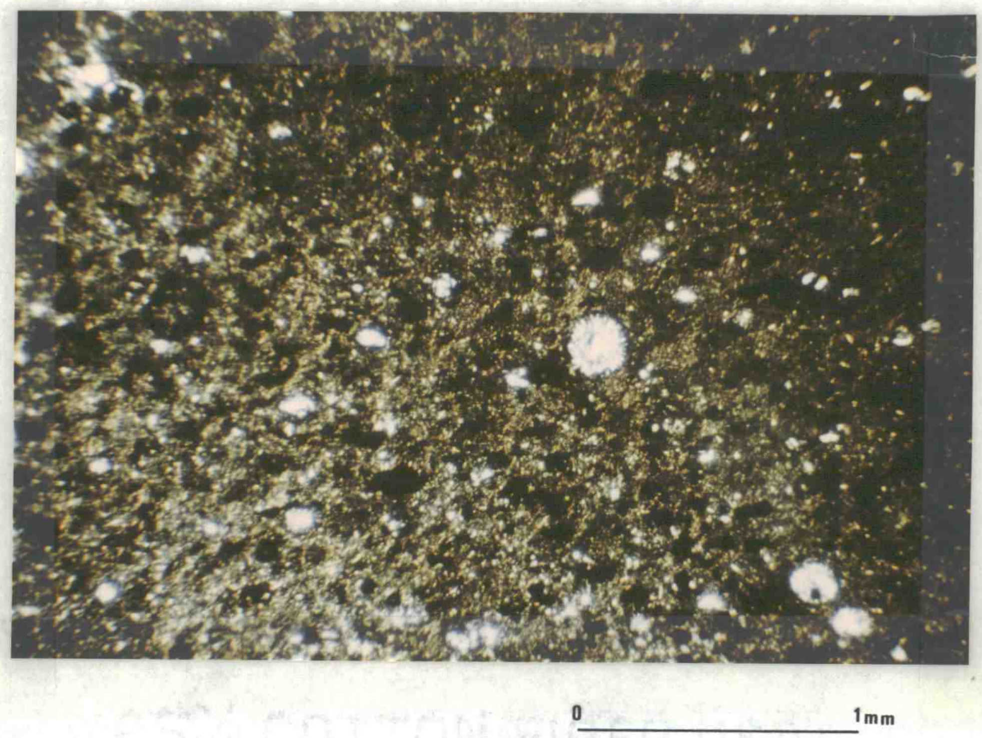


Figure 30: Photomicrograph of radiolarian chert. Radiolarian test shows little sign of compaction. Crossed nicols (overexposed). Matrix is microcrystalline quartz.

Figure 31: Folded chert beds south of the base of section WSII at Warm Springs.

DEVONIAN DEPOSITIONAL ENVIRONMENTS

McColley Canyon Formation

The McColley Canyon Formation, cropping out only at the west end of Empire Canyon, was deposited as a lime mudstone on a gently inclined, relatively featureless, open shelf depositional regime (Figure 32A). Allochems in the fine-grained sediments were mud-supported, and deposition probably took place below wave base. The primary depositional textures characteristic of tidal or supratidal environments are absent. For examples, the autochthonous benthic shelly fauna, scour-and-fill tidal channels and ripple crosslaminae, desiccation features, oolites, laminated dolomite and (or) evaporite suites, fenestrate bird's eye structures, vertical infaunal burrows, and algal evidence: all are missing.

Abundant micro-fecal pellets imply an environment suitable for pellet-making organisms such as mud-ingesting worms. In situ deposition of the pellets in a low energy environment is suggested by the mud-support fabric and the lack of pellets segregated into laminae. The transportation history of pellets, however, is difficult to determine. If the pellets were subject to early cementation, they would not be fragile and could withstand some turbulence. The preservation of the pellet shape indicates a high content of original carbonate (Folk and Robles, 1964), but the smooth ellipsoidal shape and good sorting of pellets does not help with determining either bottom traction or turbulence (Bathurst, 1975). However, recent studies of pellets generated in both shallow and deep marine sediment show that fecal pellets are not transported far from the original site of deposition (Purdy, 1963; Folk and Robles, 1964; Bathurst, 1975; Reineck and Singh, 1980).

Except for crinoid fragments, abundant in some beds, the dolomitized pellet wackestones of the Kobeh Member of the McColley Canyon are void of benthic shelly fauna. This general dearth of calcareous epifauna also was reported in the Kobeh of Hot Creek Canyon (McGovney, 1977). However, abundant epifauna was reported farther north in the Morey Peak quadrangle (Potter, 1975) and the northern Antelope Range (Trojan, 1978).

A sparse shelly fauna is not surprising for a pellet-mud environment. Recent studies of the lithofacies and biofacies surrounding Andros Island on the Great Bahama Bank illustrate a negative or low spatial correlation between pellets and corals or molluscs (Purdy, 1963; Bathurst, 1975). Subtidal sediments extending for 10-20 miles (16-32 km.) seaward of the tidal flat zone are composed almost wholly of micro-pellets averagine 20-50 microns in length (Shinn et al., 1969).

The original mud-rich matrix has been dolomitized since deposition. Regional studies of Devonian dolomites in Nevada suggest that secondary dolomitization in the subsurface is a diagenetic process controlled by paleogeography (Dunham and Olson, 1978). Bordering the McColley Canyon Formation to the east was a broad primary dolomite suite represented by the Beacon Peak Dolomite and its equivalent, the Sevy Dolomite. The McColley Canyon interfingered with this dolomite suite during the Pragian and was, therefore, certainly in a position to receive dilute, magnesium solutions flushed seaward through the subsurface during periods of high rainfall over large subaerial tracts of land.

Prior to deposition of the McColley Canyon, a ridge and basin topography on the outer continental shelf created a system of silled depositional basins (Matti and McKee, 1977). These silled basins lasted throughout late Early Silurian (late Llandoverian) to middle Early Devonian (early Pragian). With the onset of the Pragian transgression, however, these basins began to fill, and the outer shelf was gradually transformed into a gentle depositional ramp (Matti et al., 1975; Matti and McKee, 1977; Murphy, 1977; Johnson and Sandberg, 1977). The McColley Canyon rocks at Empire Canyon represent this deepening event and deposition on a carbonate shelf below wave base.

A second cycle of regression and transgression occurred in the upper Lower Devonian and deeper water shales and limestones were deposited over the dolomite. The deepening event is represented in the compressed and covered section of McColley Canyon from ECIII-60 to ECIII-90. Why this part of the section is compressed remains an enigma. Evidence for a type of sediment by-pass model--hardgrounds, for example--or a starved basin model is lacking, and response to compaction is not significant in these fine-grained limestones.

Denay Limestone

The lower Denay Limestone at Empire Canyon and Rawhide Mountain is interpreted to represent a lower slope or level bottom depositional environment similar to the paleoenvironment described by Murphy (1977) for the lower Denay of Roberts Mountains (Figure 32B,C). Most of the thickness of the measured sections in both Empire Canyon and Rawhide Mountian is of hemipelagic sediments (as defined by McIlreath and James, 1979) in part derived from a carbonate platform to the east (Murphy, 1977; Johnson and Sandberg, 1977). Included with these sediments are thin interbeds of volumetrically minor allodapic limestones (Meischner, 1964).

The fine-grained limestones are dark, evenly bedded with planar or slightly undulatory contacts, and characterized by millimeter-thick laminae. The planar depositional fabric suggests suspension deposition at slow settling rates of very fine-grained carbonate. The lack of bioturbatinn suggests an anoxic environment not conducive to sediment infauna. Sedimentary structures commonly produced by waves or currents are lacking. Similar limestones suggestive of a deep, quiet marine setting have been described by Cook and Taylor (1977), Davies (1977), Matti and McKee (1977), Potter (1975), Trojan (1978), and Wilson (1978).

Common to these hemipelagic limestones are radiolarians, sponge spicules, some conodonts and conodont pearls, and locally abundant dacryoconarids on bedding surfaces. Although the conodont animal's mode of life is still uncertain (Klapper and Barrick, 1978), the cosmopolitan distributional pattern of most conodonts found at Empire Canyon and Rawhide Mountain, combined with the sedimentary evidence, reflects a low energy, offshore, deep water depositional setting. Dacryoconarids, with thin shells, common occurrence on bedding planes of fine-grained limestones, radial symmetry, absence of an operculum to prohibit infiltration of mud and silt, and rapid world-wide dispersal suggest a pelagic mode of life (Bouček, 1964; Fisher, 1962). Small brachiopod shells and spines, crinoid debris, and ostracode valves are minor components of these mudstones as are disseminated silt particles. This lack of abundant shelly benthic fauna again suggests deposition in an offshore, subphotic, deep marine lithotope below the oxygen minimum layer.

Monotonous dark gray, fine-grained limestones, generally thin-bedded with flat planar contacts and internal micro-laminations, are characteristic of carbonate slope, hemipelagic deposits (McIlreath and James, 1979). In addition, soft-sediment slumping which is relatively common on modern continental slopes and at the base of slopes (Cook and Taylor, 1977; McIlreath and James,

1979) occurs in the eastern end of Empire Canyon just west of the base of section ECII. Beds above and to the west of the slump become evenly bedded, indicating a return to a more nearly level bottom or gentle incline.

Six of Wilson's (1978) seven criteria for "deeper water" limestones are satisfied by the mudstones at Empire Canyon and Rawhide Mountain. The limestones are predominantly lime mud, dark, laminated, evenly bedded with planar contacts, and contain a pelagic fauna and slump structures.

Regionally, the Denay Limestone also fits into a slope interpretation. The Denay is exposed in a north to south trend which divides the Middle Devonian of central Nevada into two major lithologic types: 1) shallow water dolomite to the east, and 2) deeper water limestones to the west (Murphy, 1977). The fine-grained limestones of Empire Canyon and Rawhide Mountain are interpreted to be western equivalents of a broad dolomite platform to the east (Johnson and Sandberg, 1977; Murphy, 1977). Limestones described by Murphy (1977) exhibit a similar facies relationship (Murphy, 1977, Figure 3).

The laterally continuous laminations, dark color of the limestones, and "rotten egg" aroma produced by hydrogen sulfide in a reducing environment often reflect stagnant basin conditions (Matti and McKee, 1977; McIlreath and James, 1979). Murphy (1977) proposed that the Denay basin was sheltered from circulating, normal marine waters by a carbonate barrier or series of carbonate buildups in the general area of the antecedent Toiyabe Ridge.

Allodapic Limestones

Interbedded with these mudstones are widely spaced, pelletal-bioclastic wackestones, packstones, and grainstones. The brachiopod valves commonly are disarticulated and broken and are not in life position. Crinoids, abundant in most of the beds, are fragmented. Most clasts are pellets in point and concave-convex contact in section ECII-152 or tightly packed as in RM-200. Subparallel alignment of fossils along with the close packing of pellets, thick shelled brachiopod valves at Rawhide Mountain, and the fragmented nature of the brachiopods, crinoids, and bryozoans indicate an allochthonous nature to these constituents.

One bed, RM-200, shows minor size grading from fossiliferous, pelletal grainstone to a pelletal wackestone. Where visible, most contacts with underlying mudstone beds are planar and sharp yet the bed grades imperceptibly into the superjacent mudstone. This grading is common in turbidite beds.

These allodapic limestones do not exhibit the vertical sequence of sedimentary structures typical of the Bouma cycle in detrital siliciclastic sediments. However, this is not uncommon in limestone turbidites (Davies, 1977; McIlreath and James, 1979). The bioclastic sediments are similar to those described by Davies (1977), Cook and Taylor (1977), Murphy (1977), and McIlreath and James (1979).

Allodapic beds are more numerous and contain a wider assortment of fossils in sections ECI, ECII, and RM than in section ECIII. This may reflect the transition, east to west, of the proximal lower slope into a more level bottom, distal basinal setting. Stratigraphic sections ECI, ECII, and RM are paleogeographically closer to the shallow carbonate platform than is ECIII. By the time any turbidity currents reached ECIII, only a few small crinoid ossicles and pellets remained in suspension.

Except for the bioclastic beds in the upper 75 feet (23 m.) of ECII, the allodapic beds show a high percentage of crinoid debris. Some branching bryozoans were moved downslope. Shallow water features such as primary or penecontemporaneous dolomite and algal clasts are not represented. Ramose bryozoans and crinoid meadows flourished in deeper water slope environments or basinward of carbonate shelves (Davies, 1977). Either crinoid colonies grew preferentially in deep water slope environments susceptible to turbidity currents, or crinoid bioclasts, because of their high internal skeletal porosity and thus low bulk density, were preferentially remobilized and transported downslope by turbidity currents (Mailkem, 1968; Davies, 1977). Periodic storms or minor earthquakes could readily trigger downslope movements of unstable, unconsolidated slope sediments.

The thin, volumetrically minor allodapic beds which crop out between ECII-85 to ECII-290 (Plate 4) represent deposition during the Taghanic onlap. This transgression is reported to be "the most extensive transgression of North America" (Klapper and Johnson, 1980, p. 432), and its inception at Empire Canyon is probably represented by the covered slope above ECII-85. The Taghanic onlap begins in the Middle varcus Subzone, at the base of Interval 21 (Figure 5), in the northern Antelope Range (Johnson et al., 1980, Figure 4; J. G. Johnson, pers. comm., 1982). Sample ECII-152, the first allodapic bed found in the covered interval above ECII-85, has been dated as upper Middle or Upper varcus Subzones, and, therefore, represents deopsition during the Taghanic onlap.

Stromatoporid clasts in the allodapic beds of ECII-295 to ECII-365 suggest a progradation to the west of the shelf edge. Stromatoporoids were a major group that contributed significantly to carbonate buildups during the Devonian (Heckel, 1974). Crinoids and brachiopods are absent from these limestone turbidites. The same shallowing probably took place at Rawhide Mountain although the record above the Lower varcus Subzone has been erased by faulting.

A correlative shallowing event is represented in the Technocyrtina-rich beds of Interval 27, Lower dengleri Subzone, collected and dated from the northern Antelope Range (Johnson et al., 1980, Figure 4; Trojan, 1978). Above these beds in the northern Antelope Range, a Frasnian transgression ensued (Johnson et al., 1980). Post-Devonian faulting in Empire Canyon, however, destroyed this part of the succession.

Light colored dolomite which contains local, poorly preserved, silicified horn corals along with stromatoporoids lies along the ridge to the east of the allodapic beds of ECII-365 and suggests shallowing. This dolomite is not extensive and is restricted to eastern Empire Canyon.

In summary, the Denay Limestone at Empire Canyon and Rawhide Mountain is interpreted as hemipelagic deposition at the toe of a slope grading laterally to the west of Empire Canyon into more distal, level bottom conditions (Figures 32B,C). Hemipelagic deposition was periodically interrupted by thin-bedded, allodapic limestones. These limestones represent turbidity currents which emanated from slumping of upper slope crinoid meadows through early Middle varcus Subzone time. The ratio of hemipelagic deposits to turbidite deposits suggests a similarity with lower fan, low energy turbidites found in siliciclastic submarine fan complexes (Nelson and Kulm, 1973; Reineck and Singh, 1980). However, without the mid and upper fan facies present, such a comparison can only be speculative. Distal, mud-rich turbidity deposits have been described by Davies (1977) in a lower slope or basin lithotope and by Middleton and Hampton (1976) as the end of a continuum of gravity slide mechanisms.

The easternmost exposures of the Denay are present in the Hot Creek Range and the mountain ranges directly to the north (Murphy, 1977; Trojan, 1978) and in the Reveille Range to the southeast (Ekren et al., 1973). The shallow water carbonate facies was, therefore, nearby to the east and gradually grograded to the west. This progradation ended shortly below the Middle _varcus_ Subzone to Upper _varcus_ Subzone identified at Empire Canyon.

The Denay basin geometry is conjectural. The anomalous clean limestone combined with the lack of current-influenced sedimentary features, general lack of indigenous benthic fauna, the presence of fine laminations, dark color, and fetid odor, all suggest a deep marine, quiet water, anoxic environment below the minimum oxygen level but above the carbonate compensation depth. The lack of circulation may have been influenced by carbonate buildups to the west (Murphy, 1977).

Undifferentiated Devonian Limestone

The undifferentiated, Middle Devonian limestone that crops out in isolated pods at Warm Springs does not differ significantly from the Denay Limestone. The thin-bedded tentaculitic mudstone represents a quiet, deep marine environment. The lack of dolomitization is an indication of formation well offshore from the shallow water platform. The mudstone is laminated and non-burrowed.

The interbedded wackestones are similar to the

Denay allodapic beds. The increase of silicified brachiopods and tabulate corals suggests that these fossils, originally growing on the shallow shelf edge to the east, were dumped higher on the slope than those of Empire Canyon. Nevertheless, the paleoenvironment did not differ significantly from that described for the Denay of Empire Canyon and Rawhide Mountain.

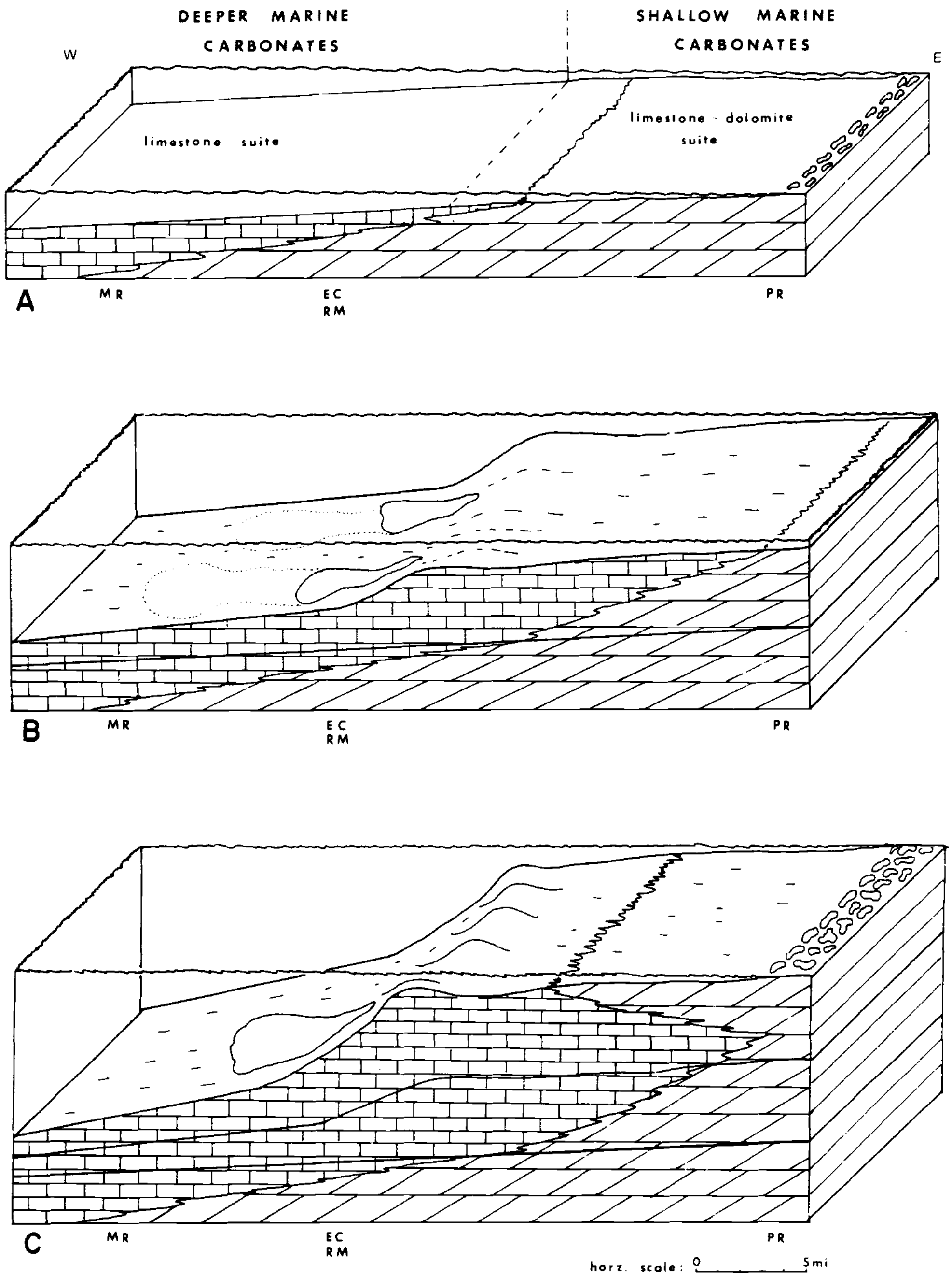


Figure 32: Conceptual diagrams of Devonian deposial environments. A) Depositional ramp of ECIII-22, B) Slope formation and deposition of allodapic beds ECI-226, C) Shallowing during ECII-365. Letters indicate localities as follows: MR, Monitor Range; EC, Empire Canyon, Hot Creek Range; RM, Rawhide Mountain, Hot Creek Range; PR, Pancake Range. No vertical scale.

MISSISSIPPIAN DEPOSITIONAL ENVIRONMENTS

Mississippian Chert

The well-bedded, dark cherts at Warm Springs are analogues of today's deep sea radiolarian ooze. Pelagic fauna, mainly radiolarians and sponge spicules, accumulated slowly in a deep marine, subphotic setting with anoxic bottom conditions. A deep water depositional mode is also supported by the association with overlying limestone turbidites and pelagic limestones. Radiolarites with similar characterisics have been described from the Alpine Jurassic by Winterer and Bosellini (1981) and Garrison and Fischer (1978).

The total lack of benthic fauna and bioturbated carbonate sediments suggests restricted circulation. The lack of any carbonate indicates deposition beneath the carbonate compensation depth (CCD).

Paleobathymetry of the CCD is difficult to access. The CCD represents, today, a dynamic equilibrium between the rate of supply and the rate of solution of calcium carbonate and has deepened dramatically since the advent of calcareous nannoplankton and microplankton in the Jurassic (from about 2000 m. to 4500 m.) (Garrison and Fischer, 1978; Berger and Winterer, 1974). Fluctuations in the CCD are caused by variables which remain a conundrum for the Paleozoic. Variables such as the amount of calcareous organisms in the surface waters, bottom circulation, vertical mixing, effect of sea level rise on the CCD, rates of supply, rates of dissolution, and temperature effects on the CCD cannot be determined. Nevertheless, Gutschick and others (1980) postulated a paleobathymetry of 100-200 meters for the foreland trough (see below) based on sedimentary structures. Poole (1974, Figure 25) depicts a depth of approximately

1200 meters for the trough in the early Mississippian.

Regionally, the Warm Springs cherts accumulated in a trough forming to the east of the submerged Antler allochthon (Poole, 1974; Poole and Sandberg, 1977a; Gutschick et al., 1980; Johnson and Pendergast, 1981). Paleogeographic reconstructions by Gutschick and others (1980) and Johnson and Pendergast (1981) indicate a trough forming during the time of the Lower *crenulata* Zone during the Antler orogeny and while transgression was widespread on the craton. Although the chert is not dated, overlying Mississippian limestones have been dated as no older than *isosticha*-Upper *crenulata* Zone (Appendix I).

Mississippian Limestone

Deep water sedimentation in the foreland trough continued with the deposition of the Lower Mississippian "Camp Creek" equivalent. However, this fine-grained limestone contains limestone turbidite deposits as well as hemipelagic sediments.

The aphanitic and argillaceous nature of the limestone coupled with the dark color and fetid odor indicate a reduced oxygen content. Meandering indigenous grazing trails of the *Nereites* community found in the hemipelagic shaly beds near the base of WSII also imply a deep water, bathyal marine environment (Seilacher, 1978). These trails and mottled bedding suggest dysaerobic zone deposition (Byers, 1977).

Benthic shelly faunas are noticeably missing from these deposits. On the other hand, abundant pelagic sponge spicules oriented parallel to bedding and calcispheres resembling radiolarian tests are prevalent in wackestones and mudstones in WSII and WSIII (Figures 26, 28). The *Nereites* community, sponge spicules, and calcispheres together with the fabric, color, and fetid odor of the limestone suggest a bathyal marine environment for the Warm Springs Mississippian limestone.

Graded beds are minor, but some do occur in WSI and WSII. Contact with underlying mudstone beds is sharp. The graded beds show internal gradation from fine grained limestone to aphanitic lime mudstone. Silt-size quartz is a major component in these beds along with ripups and pellets. Rarely, a brachiopod spine and crinoid fragment occur. Carbonate clasts range from mud-supported to grain-supported and are oriented parallel to bedding. Although the classic sedimentary structures of the Bouma sequence are absent, the graded bedding, orientation of clasts, and bedding contacts suggest transportation into basinal environments by turbidity currents. Small current ripple marks such as those near the base of section WSII (Figure 23) are common to turbidity currents (Reineck and Singh, 1980). As previously mentioned, limestone turbidites do not necessarily mimic the typical Bouma sequence of sedimentary structures found in siliciclastic sediments (Davies, 1977).

A mixed conodont fauna, especially in WSI and WSII, is perhaps the most significant criterion for recognizing limestone turbidites at Warm Springs. Several stratigraphic levels yielded <u>Gnathodus</u> <u>punctatus</u>, <u>Siphonodella</u> <u>isosticha</u>, and other conodonts of possible <u>isosticha</u>-Upper <u>crenulata</u> Zone age mixed with reworked Kinderhookian <u>Siphonodella</u> species and Late Devonian <u>Palmatolepis</u> species. Similar collections, all identified by G. Klapper, have been used to interpret limestone turbidites in the Camp Creek sequence or its equivalent at Swales Mountain (Johnson and Pendergast, 1981), the central Monitor Range (Wise, 1976), the northern Antelope Range (Trojan, 1978), and near Morey Peak in the northern Hot Creek Range (Potter, 1975).

The occurrence of mixed Devonian and Mississippian biotas, particularly conodonts, also has been used by Murphy (1977) and Davies (1977) to identify fine-grained limestone turbidites. As Davies points out (p. 242), distal fine-grained carbonate deposits resulting from turbidity currents may easily be overlooked because they lack the characteristic grain-size sorting and clastic textures common to proximal deposits. Most turbidite deposits at Warm Springs fall into this category of mixed fauna, but without sedimentary structures indicative of gravity flow mechanisms (Middleton and Hampton, 1976; Reineck and Singh, 1980).

Bedding thickens up section until four and six foot beds are common (Figure 22). Such thickness is unusual for distal, fine-grained turbidites (Scholle, 1971a; McIlreath and James, 1979). Assuming these calcareous sediments were deposited in the foreland trough, sediment "ponding" as described by Scholle (1971a) seems to be a valid way to produce the observed thickness of turbidites.

Paleogeographically, the turbidites may represent the distal equivalents of the proximal turbidites of Wise's (1977) Camp Creek equivalent in the Dobbin Summit area of the Monitor Range, one mountain range west of the Hot Creek Range. The Kinderhookian-lower Osagean represents Phase II of Johnson and Pendergast (1981), a time when compressive horizontal forces caused the formation of a foreland trough (Poole, 1974; Poole and Sandberg, 1977) and the emplacement of the Antler allochthon above the Roberts Mountains thrust. Erosion into Upper Devonian carbonates of the allochthon resulted in conodonts being reworked into the Mississippian trough.

As the toe of the thrust advanced, a broad submarine fan complex developed (Johnson and Pendergast, 1981;

Harbaugh and Dickinson, 1981). The turbidites of Warm Springs identify with the outer or lower submarine fan facies (Harbaugh and Dickinson, 1981) of Mississippian clastic deposits in the Antler foreland basin. Turbidity currents that debouch from channels bifurcating across the midfan area would spread across the lower fan causing progressive thickening of turbidites as the sediments ponded in the basin.

Significantly, the source of the Mississippian turbidites was from the west rather than from the east as it was for the Devonian allodapic beds. Concrete evidence for an eastern source, the Joana carbonate bank for example, is lacking at or near Warm Springs. The Joana carbonate bank has been considered to be a contributing source of foreland trough sediments (Poole, 1974; Poole and Sandberg, 1977; Gutschick et al., 1980; Johnson and Pendergast, 1981) and may indeed be contributing to the Tripon Pass Limestone in northeast Nevada (Poole, 1974). However, an eastern, shallow water carbonate shelf or bank influence was not felt in the Warm Springs area during the Early Mississippian. At this time, the carbonate bank was probably separated from the foreland trough by a starved basin (Gutschick et al., 1980).

CARBONATE DIAGENESIS

McColley Canyon Formation

Dolomitization of fine-grained limestone is the principal diagenetic feature of the McColley Canyon Formation in Empire Canyon. The relationship of the dolomite crystals to other diagenetic features indicates replacement dolomite formed before compaction could produce strained carbonate grains, close packing, or stylolites. Following the terminology of Choquette and Pray (1970), dolomite formed in the "eogenetic" zone before sediments were deeply buried in the "mesogenetic" zone, beyond the reach of surface processes.

The fabric selective, eogenetic dolomite consists of large (to 0.4 mm.), clear, unstrained, interlocking crystals with compromise boundaries and few inclusions (Figure 33). Early dolomitization preserved these burrow-like structures from being compressed by later compaction. Fabric selectivity also implies original porosity, permeability and dolomitization before neomorphism altered the surrounding calcium carbonate. In sample ECIII-5, micro-pellets have retained their shape and are loosely packed in a matrix of microspar and dolomite. The peloids are in tangential or point contact or are otherwise supported by the dolomite matrix. The sutured grain boundaries and close packing associated with compaction are missing. Clearly, the dolomite formed in the original pore space before compaction and neomorphism eliminated the porosity. In addition, dolomite crystals are not found concentrated along stylolites as is commonly the case with mesogenetic dolomite (Wanless, 1979; Mattes and Mountejoy, 1980).

The limpid form of the McColley Canyon dolomite, along with the lack of evaporites or their traces, precludes formation of penecontemporaneous protodolomite by hypersaline brines of tidal or supratidal origin (Folk and Land, 1975). On the other hand, one possible model for eogenetic dolomite of the McColley Canyon Formation of Empire Canyon involves mixture of normal seawater or its connate equivalent with meteoric water (Folk and Land, 1975; Dunham and Olson, 1978, 1980). Interstitial seawater diluted by mete ric ground water would allow for slow crystallization rates and a low concentration of foreign ions competing for dolomite lattice sites (Folk and Land, 1975), the two principal requirements for the precise Ca-Mg cation-ordering essential for eogenetic dolomite formation. Dolomite formed at a saltwater/freshwater lens contact zone is characteristically clear limpid dolomite (Folk and Land, 1975).

Dunham and Olson (1978, 1980) apply this model in their discussion concerning paleogeographic control on diagenetic dolomite formation in Nevada. The freshwater groundwater lens, derived from large subaerially exposed tracts of the shallow water inner platform, would migrate towards the east during times of onlap and to the west during times of offlap. Thus, the limestone-dolomite boundary should reflect regional transgressions and regressions of the coastline (Dunham and Olson, 1978, Figure 1).

In the present model, the Pragian-Dalejan onlapprogressively moved the limestone-dolomite boundary far to the east of Empire Canyon and effectively eliminated the possibility of eogenetic dolomitization for the overlying deep water Denay Limestone. Dolomitization by fresh water/seawater mixing requires the early interaction of carbonate sediment with both meteoric-derived ground water and interstitial seawater acting as the Mg source. With increased onlap, the lateral proximity to an area of freshwater recharge would diminish and eventually terminate early diagenetic dolomitization (Dunham and

Figure 33: Photomicrograph of ECIII-22 showing fabric selective, eogenetic dolomite filling burrow-like structures. Crystals are large, clear, and unstrained. Influence of compaction is slight. Plane light.

Olson, 1980).

The absence of evaporites was probably determined as much by the paleoclimate as by shallow subtidal, as opposed to peritidal, diagenesis. Central Nevada may have been between 0° South to 15° South latitude during the upper Lower Devonian (Heckel and Witzke, 1979). These latitudes correspond with today's equatorial doldrums rain belt. Such a location would provide a humid, high rainfall Bahama-type paleoclimate in contrast to the persistently arid-subtropical climate found around the recent Persian Gulf (Dunham and Olson, 1978, 1980; Heckel and Witzke, 1979). If the catchment area was large, as supposed for the Devonian, a dolomitizing freshwater lens in the subsurface could extend for miles out to sea. Evaporites, if they existed at all, would be ephemeral and would not be preserved.

Denay Limestone

As the limestone-dolomite platform shifted to the east, a different diagenetic response was felt in the fine-grained slop-derived limestones of the Middle Devonian. Authigenic, euhedral to subhedral cubes and clusters of pyrite formed penecontemporaneously with carbonate deposition in a reducing environment only a few centimeters beneath the surface/seawater interface. Pyrite cubes approach 2 mm. in size and are aligned parallel to laminae. Sedimentary pyrite can form when sulfate-reducing bacteria decompose organic compounds (Scholle, 1971b).

Compactional dewatering of the carbonate ooze within the first few feet of burial probably resulted in the mud being neomorphosed to micrite (Folk, 1965). The solution of the smaller grains and the reprecipitation onto larger original grains would decrease the already limited pore space in the fine-grained micrite. The growth of micrite crystals was restricted to 1.5 to 2 micron diameters by a magnesium "cage" which formed around the calcite polyhedron and, thus, prevented further growth (Folk, 1974).

Except in the allodapic beds, matrix cement is absent in the Denay Limestone. Because the limestone was deposited in relatively deep waters offshore from the platform, the sediments were not subjected to a fresh water "flush". Marine pore waters could be retained longer and cementation delayed (Scholle, 1971b). Retention of Mg ions would further prohibit crystal growth past 2 microns.

Large brachiopod shells, while somewhat disarticulated, are not significantly crushed in the allodapic beds. The pellets have also avoided the onslaught of compaction and, thus, suggest early calcite cementation. Geopetal calcite, crinoid overgrowths, and sparry calcite and silica cement surrounding allochems (Figure 34) suggest a permeability to the allodapic beds which permitted early calcite cementation soon after deposition of the allodapic limestones (Folk, 1965).

Remobilization of silica occurred early in the diagenetic history of the authochthonous limestones. Opaline sponge spicules provided the silica which was selectively precipitated in the walls of brachiopod spines such as those in sample ECIII-91. Apparently, replacement of calcite by silica proceeded without the formation of a void space. If a void space had developed, the cell walls would have been displaced and silica would have replaced the calcite within the interior of the spines.

With increased burial, the "micrite curtain" (Folk, 1965, 1974) was rent asunder and micrite recrystallized by coalescive neomorphism into microspar. Cloudy patches of microspar may be seen floating like "nuts in nut-bread" within the micrite (Folk, 1965). Folk (1974) states that any of the following four processes may be responsible for removing the Mg cage:

1. leaching of Mg ions by brackish water from sediments originally deposited in a brackish environment with initially low Mg;
2. injection of fresh water into underlying carbonate strata;
3. flushing of Mg ions by meteoric waters upon outcrop weathering;
4. seizure of Mg ions by interbedded or intercarbonate-crystal clay minerals as waters percolate through the rock.

Of these four, the last seems the most viable for the depositional setting of the Denay Limestone. Theoretically, clays change to montmorillonite or chlorite through the uptake of magnesium (Folk, 1974). However, clay mineralogy was not pursued in this report.

The pervasive dolomite rhombs floating in the matrix of ECII-149 and the few, relatively minor dolomite rhombs seen in the microspar of other samples may also be responsible for removing Mg ions released during pressure solution. With increased burial, dolomite reaction rates would increase, the Mg hydration barrier would be surpassed, and the permeability barriers would be counteracted by rate considerations thus allowing dolomite to form (Mattes and Mountjoy, 1980).

With deep burial in the mesogenetic zone, neomorphism effectively blocked the remaining pore space and pressure solution responses took on a different mode. Micro-stylolite "swarms" similar to those described by Wanless (1979) became locally abundant in the hemipelagic limestones. They are thin, discontinuous, and marked by detrital clay seams (Figure 35). Scholle (1971b) suggests a depth of 500 meters as a depth of nearly complete cementation which marks the transition from grain-to-

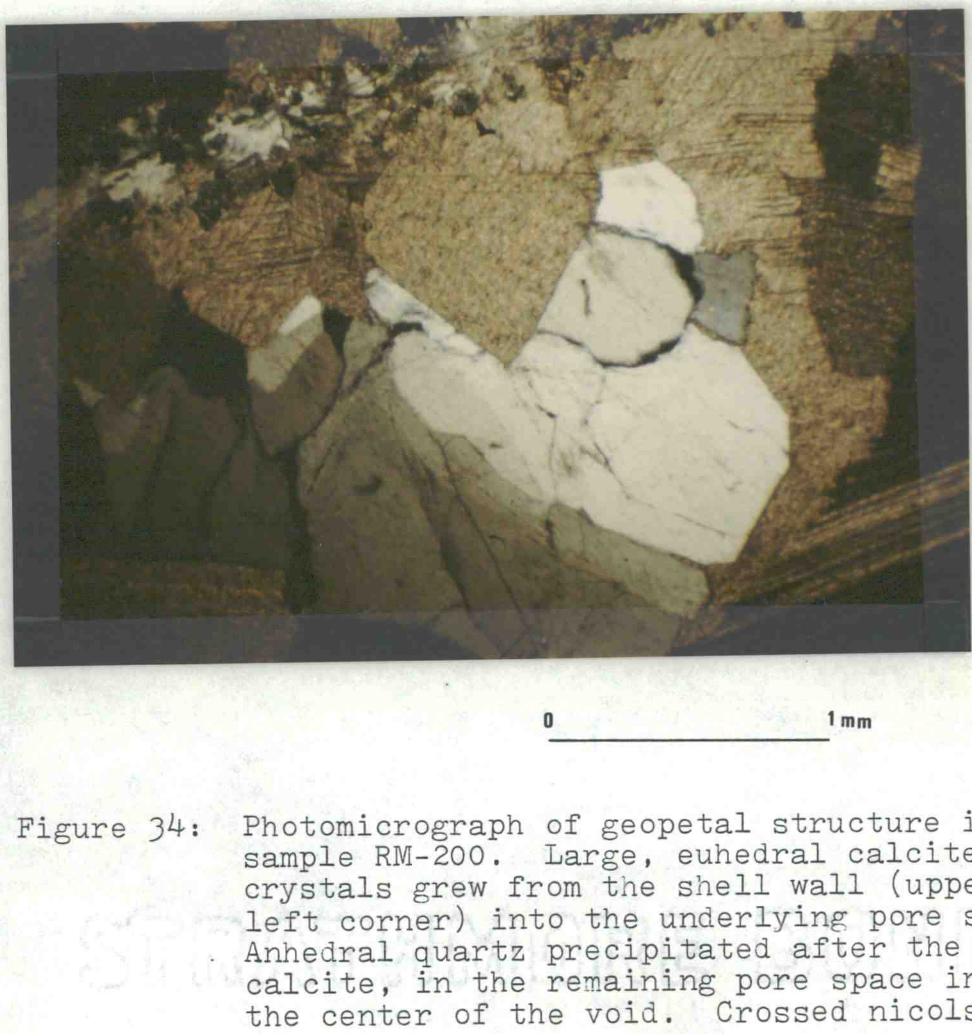


Figure 34: Photomicrograph of geopetal structure in sample RM-200. Large, euhedral calcite crystals grew from the shell wall (upper left corner) into the underlying pore space. Anhedral quartz precipitated after the calcite, in the remaining pore space in the center of the void. Crossed nicols.

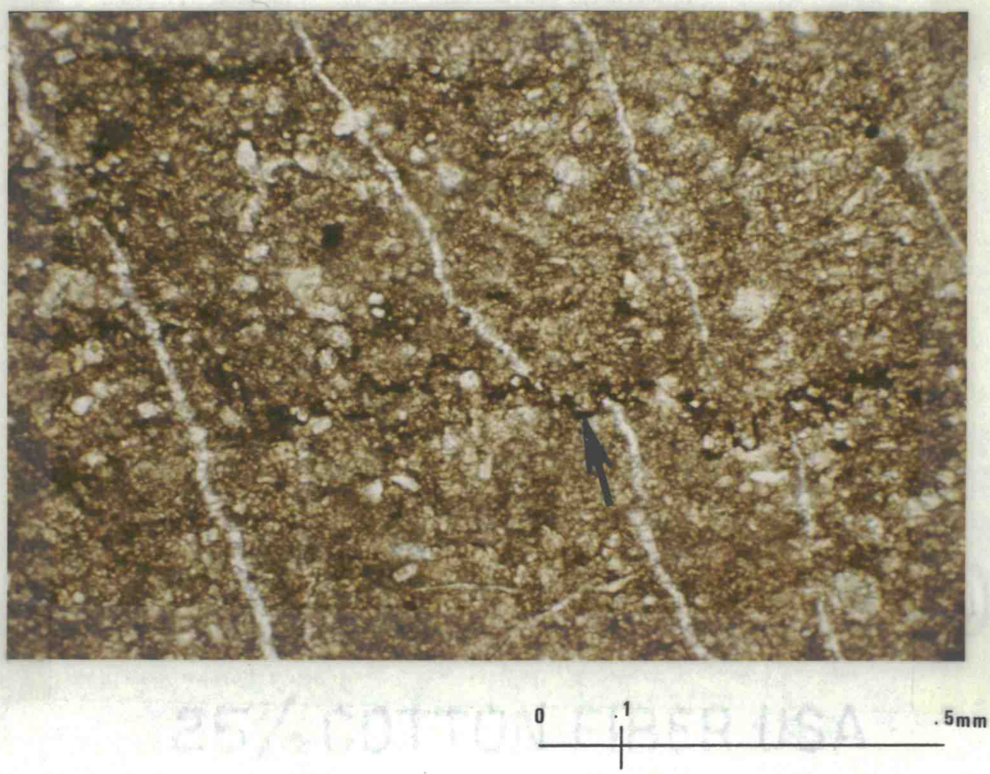


Figure 35: Photomicrograph of micro-stylolite "swarms" found in sample ECI-90 , Denay Limestone. Stylolites are outlines by dark, detrital clay seams (arrow). Secondary calcite veins offset by solution of the limestone during stylolitization (Bathurst, 1975). Plane light.

grain pressure solution of stylolite formation. Microfractures, also found after the limestones were tightly cemented, are offset by the stylolites.

Iron minerals were affected by outcrop weathering. Oxidation of these minerals to hematite and limonite can be seen in thin section as well as on outcrop.

Mississippian Limestone

The deep marine Mississippian limestone at Warm Springs responded to diagenetic processes which similarily affected the Denay Limestone. The low permeability associated with the original argillaceous carbonate ooze was further reduced by early compaction. Sponge spicules and other allochems were aligned parallel to bedding and pellets were closely packed as a result of overburden. Partial merging of fecal pellets may be the cause of the clotted or "grumous" texture (Folk, 1965) mentioned on page 50 with regards to sample WSI-17. This interpretation seems reasonable for WSI-17 where some pellets have retained their well-rounded form while others have lost their shape as a result of compaction. However, "structure grumeleuse" may also form by recrystallization of a homogeneous micrite or upon conversion of a pelsparite to microspar, which tends to blur the pellet coutline (Folk, 1965; Bathurst, 1975).

In shallow marine carbonate sediments, compactional dewatering of marine pore waters permits the influx of fresh waters and the consequential growth of carbonate cement (Scholle, 1971b). However, carbonate cement is scarce in the limestones at Warm Springs. As Scholle points out (1971b, p. 244), deep marine sediments are sheltered from any fresh water influence. Thus, calcite cementation is delayed.

While cementation is delayed, compaction and pressure solution continue to decrease available pore space. This reduced pore space and absence of carbonate cement is further suggested by the lack of overgrowths on echinoderm fragments and the compression of argillaceous material into organic, discontinuous laminae.

Sponge spicules, some with visible central canals (Figure 28), and radiolarian tests are common to these limestones. During early diagenesis the original opaline silica was partially remobilized and replaced with neomorphic calcite (Scholle, 1971b). Some of this silica precipitated in ostracode shell walls and as minor chert replacement in the matrix, but the total quantity of silica which must have been remobilized cannot be justifiably accounted for in these thin sections.

With increased burial, neomorphism of micrite to microspar significantly reduced any remaining porosity. As previously discussed, clay minerals would seize the magnesium ions thus allowing the carbonate crystals to break through the micrite barrier and grow to microspar size (Folk, 1965, 1974). Folk (1965, p. 55) believes the breakthrough is the result of porphyroid neomorphism in which a swarm of 5-6 micron crystals grow in scattered intervals throughout the matrix of 2 micron material. Continued neomorphism from 5 micron to the 20 or 30 micron stage, however, is coalescive with all crystals involved in a dynamic, uniform growth. The microspar in these Mississippian limestones is generally equant and uniform in size although more difficult to detect than the microspar of the Denay Limestone. Only under the highest microscopic power (x400) can microspar crystals be distinguished floating in the micrite matrix. The cloudy nature typical of the microspar is due to impurities trapped during the transformation.

Microspar is more prominent in these limestones than is pseudospar, but some dirty pseudospar patches have apparently formed in the matrix and in the centers of calcispheres. Neomorphism effectively destroyed any remaining porosity.

Microstylolites and microstylolite swarms in the fine grained limestone of WSI-17, WSII-273, and WSIII-730 reflect diagenetic effects resulting from pressure solution after neomorphism. Grain welding and abnormal grain penetration, common to solution transfer under pressure solution (Folk, 1965), are rare in the allodapic beds of Warm Springs. This is probably the result of the fine-grained limestone texture and the minor effects of compaction in relation to cementation resulting from pressure solution (Wanless, 1979; Mattes and Mountjoy, 1980).

Corrosion of the minor quartz silt fraction by calcite probably occurred in the deep burial, mesogenetic regime. With increased temperature, calcite would precipitate while silica would become more soluble. Microfractures also resulted from overburden while the limestone was deeply buried.

Upon exposure in the telogenetic zone, outcrop weathering oxidized the iron minerals present in minor amounts in these limestones. Subhedral pyrite, initially deposited in a reducing environment similar to the environment described for the pyrite formation in the Denay, was oxidized to anhedral hematite and limonite stains and blotches.

REGIONAL SIGNIFICANCE

While the geometry of the easternmost trace of the Roberts Mountains thrust in Eureka County has undergone extensive fine tuning since 1958 (Roberts et al., 1958; Stewart and Poole, 1974; Johnson and Pendergast, 1981), the southern projection of this trace into Nye County remains in dispute. Some authors (Johnson and Sandberg, 1977; Johnson and Pendergast, 1981) extend the thrust southward into northern Nye County through the Monitor Range and on the western side of the Hot Creek Range but do not hazard a more southerly projection. Others (Poole, 1974; Stewart and Poole, 1974; Poole and Sandberg, 1977a; Poole et al., 1977b; Stewart, 1980) extend the toe of the thrust southward on the eastern side of the Hot Creek Range, wrap it around the town of Warm Springs, and project it westward along Highway 6 towards Tonopah, Nevada. Lithostratigraphic and biostratigraphic evidence from the southern Hot Creek Range, especially the Warm Springs area, suggests that the toe of the Roberts Mountains thrust does not include the Hot Creek Range in its anatomy.

Stewart (1980) and Stewart and Poole (1974) considered the Devonian rocks at Warm Springs to belong to the shale and chert province and to have been thrusted over the limestone and shale province. They recognize four depositional provinces of Devonian strata in Nevada. From east to west these provinces are: 1) a carbonate and quartzite province which would include the nearshore dolomites such as the Sevy and Simonson dolomites as well as offshore limestones such as the Denay, 2) a limestone and shale province, 3) a shale and chert province, and 4) a chert province. Stewart (1980, p. 40) admits that assignment of the "Devonian siliceous rocks" near Warm Springs to the Roberts Mountains allochthon is tenuous considering the next closest outcrop of allochthonous rocks is approximately 50 miles (80 km.) to the northwest (Stewart, 1980, Figures 21, 22). South of these outcrops in northern Nye County the distribution of the Roberts Mountains allochthon is not well defined. The road log for the 1979 Basin and Range Symposium also refers to the Kinderhookian limestone behind the Warm Springs Cafe (which, incidently, is also the town of Warm Springs) as klippen resting on "deformed allochthonous western facies rocks that are probably middle Paleozoic" (French et al., 1979, p. 651).

To my knowledge, the Mississippian rocks at Warm Springs are not part of a klippe, and the Devonian limestone is not a part of the shale and chert facies. Rather, they belong to the same carbonate and quartzite province as those at Empire Canyon and Rawhide Mountain (Stewart, 1980, Figure 20) and represent a progressive deepening as the carbonate shelf edge migrated to the east during the Devonian (Figure 36). The Devonian rocks behind the cafe are indeed siliceous, but the silica represents Mesozoic-Cenozoicddeposits of jasperoid.

Unequivocal support for a thrust sheet is non-existent in the Hot Creek Range. The Mississippian limestone which caps the ridge behind Warm Springs conformably overlies the deep marine, bedded cherts deposited in the subsiding foreland trough. Conodonts collected from the measured sections indicate a Kinderhookian-Osagean age for these deposits. The facies represents deep water sedimentation typical of basal Mississippian units in Nevada (Johnson and Pendergast, 1981). As the incipient Antler highlands formed to the west, limestone turbidites were shed eastward into the subsiding trough. The Warm Springs Mississippian limestones represent the distal facies of a complex submarine fan complex. As such, they are part of the foreland sequence and were not emplaced by thrusting. The deep water lithotope at Warm Springs reflects the progressive onlap as orogeny continued and the carbonate shelf edge was pushed to the east (Figure 36). It is not surprising, therefore, that influence from the Joana bank, as previously supposed (Poole, 1974; Gutschick et al., 1980; Johnson and Pendergast, 1981), is missing.

A thrust contact is also missing between the Devonian limestone and bedded cherts at Warm Springs. On the contrary, sparse outcrops southwest of peak 6948 indicate similar attitudes to both Devonian limestone and chert beds. Where the integrity of the Devonian limestones has not been violated by jasperoid deposits an autochthonous Devonian to Mississippian sequence is apparent.

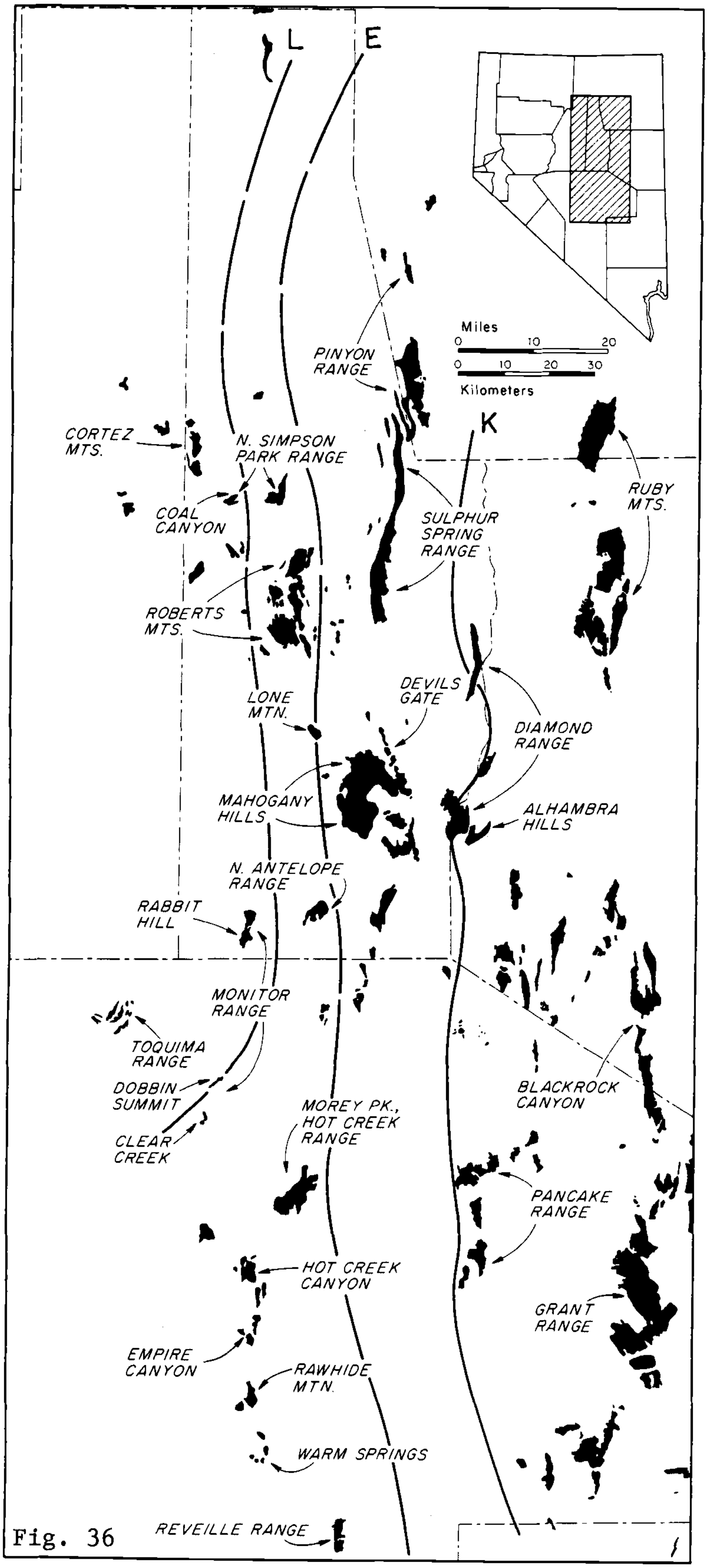
Sections measured in Empire Canyon and Rawhide Mountain also failed to provide evidence supportive of a thrust sheet. Much of the Upper Devonian to Lower Mississippian succession has been lost due to normal faulting and subsequent erosion, or possible non-deposition, but good exposures on the eastern flank of the southern Hot Creek Range show an uninterrupted superposition of strata (Quinlivan and Rogers, 1974). Interleaved rock units and low angle thrusting typical of the Roberts Mountains thrust are not present.

In the northern Hot Creek Range, Potter (1975) mapped an early Famennian rock unit correlative with the Woodruff Formation in a dubious thrust contact with both the underlying "Middle Devonian dolomite" and the Devonian Devil's Gate Formation. Brachiopods collected from the lower part of the Middle Devonian dolomite established the unit as late Middle Devonian in age, but the age of the Devil's Gate was not determined (Potter, 1975). Consequently, the relationship among the three formations remained an enigma. Unfortunately, a collecting trip to resolve these questionable relationships has so far proved unsuccessful. Samples systematically collected from the 2575 feet (785 m.) added to Potter's Summit 8782 measured section have yet to yield conodonts. Dating of these units is imperative for determining the autochthonous or allochthonous nature of the overlying Woodruff Formation.

Undisputed allochthonous rocks occur northwest of Morey Peak in the Monitor Range. Based partially on Ordovician conodonts recovered from location S509, Wise (1976) mapped the Ordovician Vinini Formation in thrust contact with Denay Limestone at Dobbin Summit. Bedded chert in the southern half of Wise's map area near Clear Creek was also considered to be Vinini. However, comparable lithology and the stratigraphic relationship with the overlying siliceous Webb equivalent (Wise, 1976, Plate 3) suggest this assemblage is autochthonous and, thus, represents Mississippian foreland deposits shed from the Antler highland (J. G. Johnson, pers. comm., 1982).

Johnson and Pendergast (1981, Figure 2) swept the trace of the Roberts Mountains thrust gently from the western edge of the northern Antelope Range southwestward to Dobbin Summit. This reconstruction represents a salient shift from previous Roberts Mountains thrust geometry (Poole, 1974; Stewart and Poole, 1974; Poole and Sandberg, 1977a; Poole et al., 1977b; Stewart, 1980). The paucity of thrust-related rocks at Empire Canyon, Rawhide Mountain, and Warm Springs supports this interpretation. Rather than swinging in an erratic aberration to include the Hot Creek Range, the toe of the Roberts Mountains thrust probably continues in a southwestward direction from Dobbin Summit (Figure 37, dotted line).

Figure 36: Migration of the carbonate shelf edge in Nevada through time. Letters refer to the following: L, Lochkovian (Lone Mountain Dolomite); E, Eifelian (Sentinel Mountain Dolomite); K, Kinderhookian (Joana Limestone). Note progressive deepening in the areas of Empire Canyon, Rawhide Mountain, and Warm Springs.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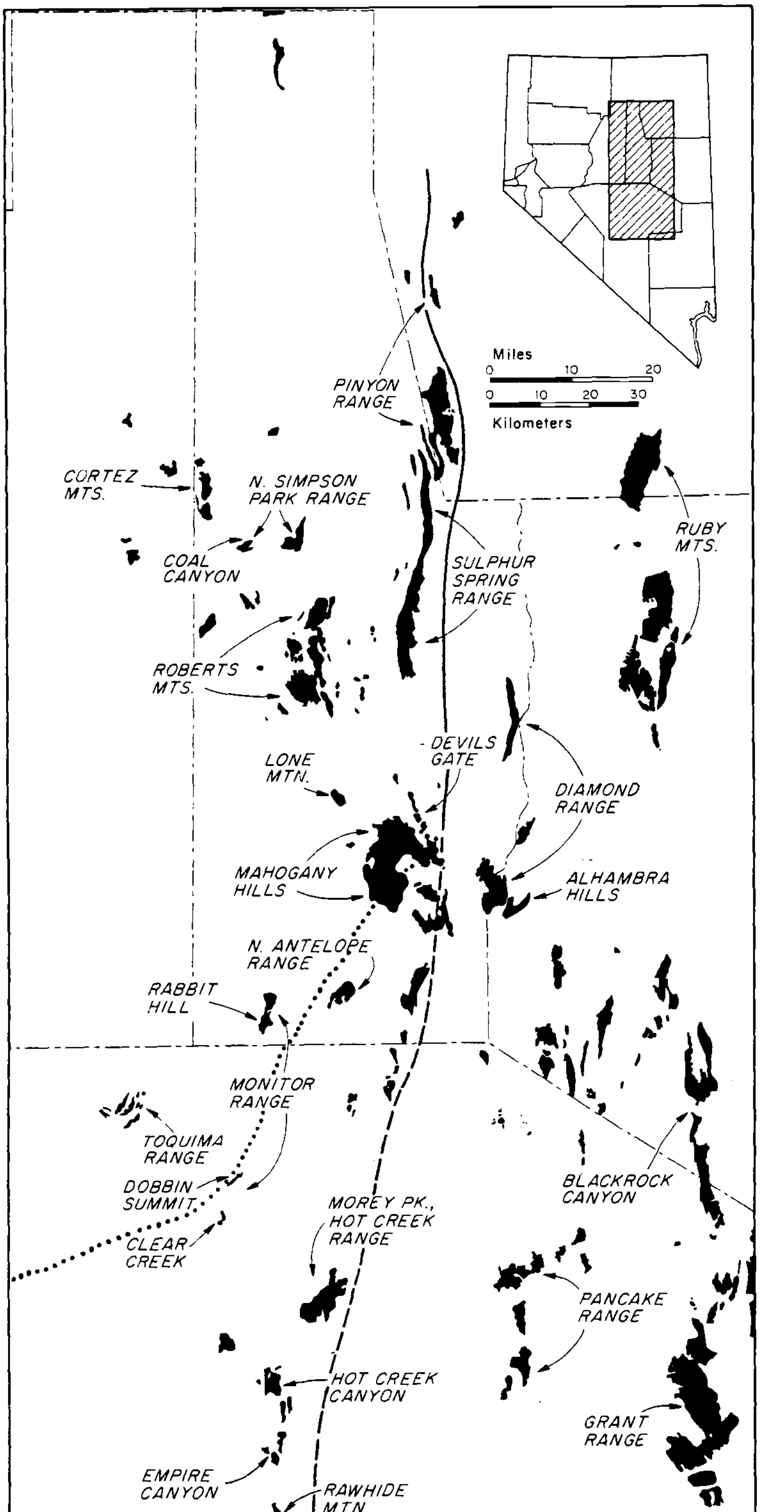
RUBY MTS.
ROBERTS MTS.
LONE MTN.
DEVILS GATE
DIAMOND RANGE
MAHOGANY HILLS
ALHAMBRA HILLS
N. ANTELOPE RANGE
RABBIT HILL
MONITOR RANGE
TOQUIMA RANGE
DOBBIN SUMMIT
CLEAR CREEK
MOREY PK., HOT CREEK RANGE
BLACKROCK CANYON
PANCAKE RANGE
HOT CREEK CANYON
GRANT RANGE
EMPIRE CANYON
RAWHIDE MTN.
WARM SPRINGS
REVEILLE RANGE

Fig. 36

Figure 37: Two interpretations of the geometry of the trace of the Roberts Mountains thrust. Dotted line represents interpretation of this report. Dashed line represents alternative interpretation (see text).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ROBERTS MTS.
DEVILS GATE
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DIAMOND RANGE
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N. ANTELOPE RANGE
RABBIT HILL
MONITOR RANGE
TOQUIMA RANGE
DOBBIN SUMMIT
CLEAR CREEK
MOREY PK., HOT CREEK RANGE
BLACKROCK CANYON
PANCAKE RANGE
HOT CREEK CANYON
GRANT RANGE
EMPIRE CANYON
RAWHIDE MTN.
WARM SPRINGS
REVEILLE RANGE
Fig. 37

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APPENDICES

APPENDIX I

CONODONT COLLECTIONS AND LOCALITIES

Conodonts reported by Gilbert Klapper May 4 and July 11, 1982. Sample numbers indicate footages from base of the section.

Empire Canyon

Section ECI:

Location: 116°26'E, 38°18'15"N; 1.1 miles S55E from Cine Mountain and 2.1 miles W. of E. boundary of T6N, R49E, northern Nye County, Nevada. Base of section at elevation 6960 feet; north side of creek bed.

Sample: ECI-5

Tortodus kockelianus kockelianus (Bischoff and Ziegler)
Polygnathus sp. indet.
indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECI-45

Polygnathus l. linguiformis indet. morph.
indet. ramiform elements

Age: indeterminate

Sample: ECI-90

Polygnathus parawebbi Chatterton
P. sp. indet.
P. intermedius (Bultynck)?
Tortodus k. kockelianus
indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECI-165

Tortodus k. kockelianus

Polygnathus parawebbi
P. l. linguiformis gamma morph.
indet. ramiform elements
Belodella sp.
Panderodus sp.

Age: *kockelianus* Zone

Sample: ECI-226

Tortodus k. kockelianus
Polygnathus trigonicus Bischoff and Ziegler
P. eiflius Bischoff and Ziegler
P. parawebbi
P. l. linguiformis gamma morph.
P. l. linguiformis indet. morph.
Icriodus sp.
indet. ramiform elements
Belodella sp.
Panderodus sp.
Neopanderodus sp.

Age: *kockelianus* Zone

Sample: ECI-228

Polygnathus parawebbi
P. eiflius
P. l. linguiformis indet. morph.
P. sp.
Icriodus sp.
indet. ramiform elements
Ozarkodina sp. aff. *O. brevis* (Bischoff and Ziegler)
Belodella sp.
Neopanderodus sp.
Panderodus sp.

Age: indeterminate

Section ECII:

Location: 116°26'E, 38°18'15"N; 528 feet N55W from ECI, across draw, base of slope.

Sample: ECII-2

Tortodus k. kockelianus
Polygnathus parawebbi
Icriodus sp. indet.
indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECII-150

<u>Polygnathus</u> <u>l</u>. <u>linguiformis</u> gamma morph.
indet. ramiform elements

Age: indeterminate

Sample: ECII-152

<u>Polygnathus</u> <u>l</u>. <u>linguiformis</u> gamma morph.
<u>P</u>. <u>ansatus</u> Ziegler and Klapper? (1 specimen)
<u>P</u>. cf. <u>P</u>. <u>beckmanni</u> Bischoff and Ziegler
<u>Ozarkodina</u> <u>semialternans</u> (Wirth)
<u>O</u>. <u>semialternans</u> ⟶ <u>P</u>. <u>latifossatus</u> Wirth
(note: this is not the same as the "<u>O</u>. sp. aff. <u>O</u>. <u>semialternans</u>" of Trojan X13)
<u>Icriodus</u> <u>difficilis</u> Ziegler and Klapper
<u>I</u>. sp. indet.
indet. ramiform elements

Age: probable zonal range is from the uppermost part of Middle <u>varcus</u> Subzone to Upper <u>varcus</u> Subzone.

Sample: ECII-365

<u>Polygnathus</u> sp. indet.
indet. ramiform elements

Age: indeterminate

<u>Section</u> ECIII:

Location: 116°25'30"E, 38°18'30"N; 950 feet N26W of Cine Mountain, Empire Canyon, and 3.3 miles W. of E. boundary of T7N, R49E, northern Nye County, Nevada. Base of section at top of first outcrop, across saddle west of Cine Mountain.

Sample: ECIII-5

<u>Icriodus</u> sp. indet.

Age: indeterminate

Sample: ECIII-22

<u>Icriodus</u> <u>nevadensis</u> Johnson and Klapper
<u>Pandorinellina</u> <u>exigua</u> <u>philipi</u> Klapper

Age: probably <u>dehiscens</u> Zone

Sample: ECIII-60

Icriodus nevadensis
Polygnathus gronbergi Klapper and Johnson

Age: *gronbergi* Zone

Sample: ECIII-91

Polygnathus parawebbi
P. parawebbi?
P. costatus costatus Klapper
P. sp. indet.
indet. ramiform elements
Belodella sp.

Age: *australis* Zone

Sample: ECIII-144

Polygnathus eiflius
P. sp. indet.
indet. ramiform elements

Age: probably *kockelianus* Zone

Sample: ECIII-184

Polygnathus robusticostatus Bischoff and Ziegler
P. angustipennatus Bischoff and Ziegler
indet. ramiform elements

Age: indeterminate

Sample: ECIII-200

Tortodus k. kockelianus
P. parawebbi
P. robusticostatus
P. angustipennatus
P. eiflius
P. angusticostatus Wittekindt
P. trigonicus
P. l. linguiformis indet. morph.
indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECIII-212

Tortodus k. kockelianus
Polygnathus parawebbi
P. sp. indet.

indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECIII-220

Tortodus *k.* *kockelianus*
T. *k.* *australis* (Jackson)?
Polygnathus *eiflius*
indet. ramiform elements

Age: *kockelianus* Zone

Sample: ECIII-245

Polygnathus *parawebbi*
P. *l.* *linguiformis* indet. morph.
Belodella sp.
indet. ramiform elements

Age: probably *kockelianus* Zone to Lower *varcus* Subzone

Rawhide Mountain

Section RM:

Location: 1320 feet S72W of peak 9145 and 5280 feet W. of E. boundary of T5N, R49E, northern Nye County, Nevada. First ridge north of steep valley cut into west flank of Rawhide Mountain.

Sample: RM-0

Polygnathus *eiflius*
P. cf. *P.* n. sp. A Savage
indet. ramiform elements

Age: *kockelianus* Zone, by superposition in this sequence

Sample: RM-49

Tortodus *k.* *kockelianus*
P. sp. indet.
indet. ramiform elements

Age: *kockelianus* Zone

Sample: RM-98

indet. ramiform elements
Belodella sp.

Age: indeterminate

Sample: RM-185

Polygnathus parawebbi
P. l. linguiformis indet. morph.
P. sp. indet.
Icriodus sp. indet.
indet. ramiform elements
"Coelocerodontus" sp.
Belodella sp.

Age: indeterminate

Sample: RM-200

Polygnathus xylus ensensis Ziegler and Klapper
P. parawebbi
Icriodus sp.
indet. ramiform elements
"Coelocerodontus" sp.
Belodella sp.

Age: probably ensensis Zone

Sample: RM-210

Polygnathus parawebbi
P. pseudofoliatus Wittekindt
P. l. linguiformis gamma morph.
P. sp.
P. sp. indet.
Icriodus sp. indet.
indet. ramiform elements
Belodella sp.

Age: possible zonal span from ensensis Zone to the Lower varcus Subzone

Warm Springs

Section WSI:

Location: 845 feet S52E of radio tower at 6485 feet elevation and 3696 feet W. of E. boundary of T4N, R49 1/2E, northern Nye County, Nevada.

Sample: WSI-1

Gnathodus punctatus (Cooper)
G. delicatus Branson and Mehl sensu lato
G. sp. indet.
Siphonodella crenulata (Cooper)
S. obsoleta Hass
S. isosticha (Cooper)
S. cf. S. isosticha
S. cooperi Hass
S. quadruplicata (Branson and Mehl)
S. duplicata (Branson and Mehl)
S. lobata (Branson and Mehl)
Polygnathus longiposticus Branson and Mehl
P. c. communis Branson and Mehl
P. sp.
Pseudopolygnathus marginatus Branson and Mehl
Pseudopolygnathus triangulus Voges
Branmehla inornata Branson and Mehl
"Spathognathudus" macer (Branson and Mehl)
"S." sp.
Elictognathus laceratus (Branson and Mehl)
E. bialatus (Branson and Mehl)
Icriodus sp.
Palmatolepis gracilis gracilis Branson and Mehl
Pa. rugosa Branson and Mehl
Pa. subperlobata Branson and Mehl
Pa. stoppeli Sandberg and Ziegler?
Pa. glabra lepta Ziegler and Huddle
Pa. glabra prima Ziegler and Huddle
Pa. glabra pectinata Ziegler
Pa. distorta Branson and Mehl
Pa. marginifera Helms s.l.
Pa. quadrantinodosa inflexoidea Ziegler
Pa. perlobata schindewolfi Müller
Pa. sp.
indet. ramiform elements

Age: Kinderhookian-Osagean (see text). Klapper notes that this is the best preserved reworked Palmatolepis he has seen from Nevada, and although the list appears complete, it is highly provisional.

Sample: WSI-11

Siphonodella isosticha
S. obsoleta
S. sp. indet.
Gnathodus sp. indet.
Polygnathus c. communis
"Spathognathodus" sp.
Elictognathus laceratus

indet. ramiform elements

Age: indeterminate

Sample: WSI-17

Siphonodella obsoleta
S. sp. indet.
P. *c*. *communis*
Elictognathus laceratus
Bispathodus sp.
Pa. sp. indet.
indet. ramiform elements

Age: indeterminate

Sample: WSI-45A

Gnathodus punctatus
G. *delicatus* s.l.
G. sp. indet.
Siphonodella isosticha
S. cf. *S*. *isosticha*
S. *obsoleta*
S. *crenulata*
S. *cooperi*
S. sp. indet.
Polygnathus c. *communis*
P. *symmetricus* Branson
Pseudopolygnathus marginatus
Pseudopolygnathus t. *triangulus*
P. sp.
"*Spathognathodus*" sp.
Bispathodus sp.
Elictognathus laceratus
E. *bialatus*
Palmatolepis perlobata schindewolfi
Pa. *marginifera* s.l.
Pa. *glabra lepta*
indet. ramiform elements
aff. *Eotaphrus bultyncki* (Groessens)

Age: Essentially the same fauna, in terms of the youngest elements, as WSI-1.

Sample: WSI-45B

Gnathodus punctatus
G. sp.
Siphonodella obsoleta
S. *isosticha*
S. *cooperi*

S. crenulata
S. quadruplicata
Polygnathus c. communis
P. inornatus
P. distortus Branson and Mehl
P. longiposticus
P. symmetricus?
P. sp.
Pseudopolygnathus marginatus
Ps. multistriatus Mehl and Thomas
Ps. marburgensis trigonicus Ziegler
Ps. sp.
"Spathognathodus" sp.
Protognathodus? sp.
Eotaphrus? sp.
Elictognathus laceratus
E. bialatus
Bispathodus sp.
Branmehla inornata
Palmatolepis marginifera s.l.
Pa. minuta minuta Branson and Mehl
Pa. distorta Branson and Mehl
Pa. glabra lepta
Pa. glabra pectinata
Pa. perlobata schindewolfi Müller
Pa. rugosa ampla Müller
Pa. rugosa rugosa
Pa. gracilis expansa Sandberg and Ziegler
indet. ramiform elements

Age: zonal range from the upper part of isosticha-Upper crenulata Zone to typicus Zone

Section WSII:

Location: 2323 feet S33E of radio tower at 6485 feet elevation and 3062 feet W. of E. boundary of T4N, R49 1/2E, northern Nye County, Nevada.

Sample: WSII-16

Siphonodella sp. indet.
Gnathodus sp. indet.

Age: indeterminate

Sample: WSII-17.5

Gnathodus punctatus
G. sp. indet.
Siphonodella isosticha
S. cf. isosticha

Polygnathus _c_. _communis_
_P_. sp. indet.
"_Spathognathodus_" sp.
Elictognathus _laceratus_
Palmatolepis _gracilis_ _signoidalis_ Ziegler
_Pa_. spp. indet.
indet. ramiform elements

Age: probably same age as WSI-1

Sample: WSII-180

Siphonodella _crenulata_
_S_. sp.
Polygnathus _c_. _communis_
"_Spathognathodus_" sp.
indet. ramiform elements

Age: indeterminate

Sample: WSII-216A

Gnathodus _punctatus_
_G_. sp.
Siphonodella _obsoleta_
_S_. _isosticha_
_S_. _crenulata_
Polygnathus _c_. _communis_
_P_. _longiposticus_
Elictognathus _laceratus_
indet. ramiform elements

Age: probably same as WSI-1

Sample: WSII-216B

Gnathodus _cuneiformis_ Mehl and Thomas
_G_. _punctatus_
_S_. _obsoleta_
_S_. _cooperi_
_S_. _isosticha_
Polygnathus _c_. _communis_
_P_. _inornatus_
_P_. sp.
Eotaphrus _bultyncki_ (Groessens)
Protognathodus sp.
Elictognathus _laceratus_
_E_. _bialatus_
"_Spathognathodus_" sp.
Palmatolepis _glabra_ _lepta_
_Pa_. _glabra_ _pectinata_
_Pa_. _perlobata_ _schindewolfi_

<u>Pa</u>. <u>quadrantinodosa</u> Branson and Mehl
<u>Pa</u>. <u>marginifera</u> s.l.
<u>Pa</u>. <u>perlobata</u> <u>postera</u> Ziegler
indet. ramiform elements

Age: <u>typicus</u> Zone based on the overlap of <u>G</u>. <u>punctatus</u> and <u>G</u>. <u>cuneiformis</u>. Klapper notes that the range of <u>Eotaphrus</u> <u>bultyncki</u> given by Lane and others (1980, Table 2) is discounted here because the only specimen they illustrate is not closely similar to the species.

Sample: WSII-273

<u>Gnathodus</u> sp. indet.
indet. ramiform elements

Age: indeterminate

Sample: WSII-325

<u>Gnathodus</u> sp. indet.
<u>Siphonodella</u> <u>obsoleta</u>
<u>S</u>. sp. indet.
<u>Polygnathus</u> <u>c</u>. <u>communis</u>
"<u>Spathognathodus</u>" sp.
<u>Palmatolepis</u> <u>glabra</u> Ulrich and Bassler
<u>Pa</u>. <u>gracilis</u> <u>sigmoidalis</u>
indet. ramiform elements

Age: indeterminate

Sample: WSII-332

<u>Gnathodus</u> <u>punctatus</u>
<u>G</u>. sp. indet.
<u>Siphonodella</u> <u>obsoleta</u>
<u>S</u>. <u>isosticha</u>
<u>S</u>. cf. <u>S</u>. <u>isosticha</u>
<u>S</u>. <u>cooperi</u>
<u>S</u>. <u>lobata</u>
<u>Polygnathus</u> <u>spicatus</u> Branson
<u>P</u>. <u>c</u>. <u>communis</u>
<u>P</u>. <u>inornatus</u> Branson
<u>P</u>. sp.
<u>Pseudopolygnathus</u> <u>marginatus</u>
<u>Ps</u>. <u>t</u>. <u>triangulus</u>
<u>Elictognathus</u> <u>laceratus</u>
"<u>Spathognathodus</u>" sp.
<u>Bispathodus</u> <u>costatus</u> (Branson)
<u>Palmatolepis</u> <u>marginifera</u>
<u>Pa</u>. <u>rugosa</u>
<u>Pa</u>. <u>perlobata</u> Ulrich and Bassler

Pa. glabra lepta
Pa. glabra pectinata
Pa. sp. indet.
indet. ramiform elements

Age: Kinderhookian-Osagean. Klapper notes that the sample could range from the isosticha-Upper crenulata Zone to the Lower typicus Zone based on the range of G. punctatus. The range of G. punctatus is outlined in Lane and others (1980, Table 2) but, unfortunately, it is not documented.

Sample: WSII-370

Gnathodus punctatus
G. sp. indet.
Siphonodella obsoleta
S. isosticha
S. cf. S. isosticha
S. crenulata
S. sp. indet.
Polygnathus c. communis
Elictognathus bialatus
E. laceratus
"Spathognathodus" sp.
Palmatolepis perlobata
Pa. rugosa
Pa. subperlobata s.l.
Pa. gracilis sigmoidalis
Pa. glabra lepta
Pa. glabra pectinata
Pa. glabra subsp. indet.
Pa. marginifera
Pa. sp. indet.
indet. ramiform elements

Age: same as WSII-332

Section WSIII:

Location: 3168 feet S77E of radio tower at 6485 feet elevation and 1373 feet W. of E. boundary of T4N, R49 1/2E, northern Nye County, Nevada.

Sample: WSIII-4

Gnathodus sp. indet.
Siphonodella crenulata
S. sp. indet.
Polygnathus c. communis

Elictognathus _laceratus_
"_Spathognathodus_" sp.
Palmatolepis _glabra_ _pectinata_
_Pa_. _minuta_ Branson and Mehl
_Pa_. _marginifera_
_Pa_. sp. indet.
indet. ramiform elements

Age: sample is no lower than the _isosticha_-Upper _crenulata_ Zone. If _S_. _crenulata_ is not reworked, the sample would fall in this zone.

Sample: WSIII-50

indet. platform and ramiform elements

Age: indeterminate

Sample: WSIII-102

indet. ramiform and platform elements

Age: indeterminate

Sample: WSIII-203

Gnathodus sp. indet.
Siphonodella sp. indet.
Polygnathus _c_. _communis_
"_Spathognathodus_" sp.
indet. ramiform elements

Age: indeterminate

Sample: WSIII-350

indet. ramiform element

Age: indeterminate

Sample: WSIII-400

Polygnathus _c_. _communis_
Palmatolepis sp. indet.
indet. ramiform elements

Age: indeterminate

Sample: WSIII-450

Siphonodella sp. indet.
Polygnathus _c_. _communis_
Palmatolepis sp. indet.

indet. ramiform elements

Age: indeterminate

Sample: WSIII-515

Gnathodus sp. indet.
Siphonodella sp. indet.
Polygnathus _c._ _communis_
Palmatolepis sp. indet.

Age: indeterminate

Sample: WSIII-550

indet. ramiform elements

Age: indeterminate

Sample: WSIII-650

indet. ramiform and platform elements

Age: indeterminate

Sample: WSIII-730

indet. ramiform elements

Age: indeterminate

The following samples are random samples collected at Warm Springs in order to help date the Devonian limestone. Localities are indicated on Plate 1.

Sample: WS1-82

Icriodus _claudiae_?
Icriodus sp. indet.

Age: Devonian

Sample: WS4-82

Polygnathus _linguiformis_ _linguiformis_ gamma
_P._ _alatus_ Huddle
_P._ _ansatus_
_P._ _timorensis_
_P._ _parawebbi_?
_P._ _xylus_ _ensensis_?
_P._ _varcus_ group

Icriodus sp. (Middle Devonian)
Icriodus sp. (_I_. _claudiae_? and _I_. _steinachensis_ beta?-- reworked from Lower Devonian)
indet. ramiform elements

Age: Givetian, Middle _varcus_ Subzone

Sample: WS5-82

Polygnathus sp. indet.
Icriodus sp. indet.
indet. ramiform elements

Age: Devonian

The following sample was collected from an isolated block of dolomite 3 miles N17W of the town of Warm Springs and 0.35 miles (1848 feet) E. of W. boundary of T4N, R50E. The sample is located on a general trend between Warm Springs, Nevada, and Rawhide Mountain.

Sample: WSD

Tortodus _k_. _kockelianus_? (1 fragmentary specimen)
indet. platform elements
Belodella sp.
Panderodus sp.

Age: possibly _kockelianus_ Zone

APPENDIX II

BRACHIOPOD COLLECTIONS

Brachiopods identified by J. G. Johnson, Spring, 1982. Locations for samples collected are given in Appendix I. Sample numbers refer to footages from the base of the section.

Empire Canyon

Section ECI:

Sample: ECI-226

Cassidirostrum eurekaensis? (2 specimens)
Leiorhynchus sp. (3 specimens)
Warrenella sp. (4 specimens)

Age: Eifelian, probably *circula* Zone

Sample: ECI-228

Cassidirostrum? sp. (1 specimen)
Warrenella sp. (3 specimens)

Age: probably Eifelian

Section ECII:

Sample: ECII-2

indet. smooth spiriferid sp. (1 specimen - discarded)

Age: indeterminate

Sample: ECII-150

Emanuella sp. (12 specimens)
Rensselandia? sp. (2 specimens)
Stringocephalus sp. (2 specimens)
indet. spiriferid sp. (3 specimens - multiplicate, lamellose)

Age: Givetian

Rawhide Mountain

Section RM:

Sample: RM-200

Leiorhynchus sp. (1 specimen)
Nucleospira sp. (2 specimens)
Warrenella sp. (4 specimens)
Echniocoelia sp. (1 specimen)
Emanuella sp. (5 specimens)

Age: Middle Devonian - Interval 20 or older

Sample: RM-210

Cassidirostrum? sp. (1 specimen)
Spinatrypa sp. (1 specimen - fine costae)
Warrenella sp. (11 specimens)
Echinocoelia (1 specimen)
Emanuella sp. (2 specimens)
indet. rensselandid sp. (1 specimen)

Age: Middle Devonian

Warm Springs

Sample: WS 4-82

Location: Approximately 1.6 miles SW of Warm Springs and 2000 feet N. of Highway 6, 350 feet W. of E. boundary of T4N, R 49 1/2E, northern Nye County, Nevada.

Schizophoria sp. (3 specimens)
Eoschuchertella? sp. (1 specimen)
Gypidula? sp. (4 specimens)
Hadrorhynchia sp. (2 specimens)
Spinatrypa sp. (2 specimens)
Variatrypa? sp. (2 specimens)
Desatrypa? sp. (5 specimens)
Mimatrypa? sp. (2 specimens)
Warrenella sp. (8 specimens)
Echinocoelia septata (8 specimens)
Emanuella sp. (39 specimens)
Cyrtina sp. (1 specimen)
Subrensselandia sp. (8 specimens)
Stringocephalus sp. (2 specimens)
indet. lamellose spiriferid sp. (1 specimen)

Age: Givetian. Upper part of Interval 19 to Interval 21.