FACIES MOSAIC OF THE UPPER YATES AND LOWER TANSILL FORMATIONS (UPPER PERMIAN), RATTLESNAKE CANYON, GUADALUPE MOUNTAINS, NEW MEXICO

BY

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ABSTRACT

This study is focused on 45 m of stratigraphic interval centered around the contact between the upper Permian, Guadalupian age Yates and Tansill Formations. Six stratigraphic sections were measured in Rattlesnake Canyon, which trends perpendicular to the Capitan Escarpment. The objective of this study was to determine, through studying and tracing individual beds, the detailed facies mosaic of a limited stratigraphic interval of carbonate shelf strata adjacent to the Capitan Limestone. The study area encompasses the most basinward 2 1/2 km of shelf strata, an area of abrupt facies changes related to differences in energy, depth, and water salinity variation.

The carbonate shelf can be subdivided into five facies of a marginal mound profile; in a shelfward direction the facies are: 1) shelf edge, 2) outer shelf, 3) shelf crest, 4) inner shelf, and 5) evaporite shelf. The marginal mound profile was first recognized by Dunham (1972). The Capitan Limestone comprises the shelf edge. Rocks of the study area comprise the outer shelf, shelf crest, and inner shelf.

Eight rock types were differentiated using several variables, grain type, grain size, grain abundance, rock fabric, and depositional texture. The rock types are: 1) skeletal grainstone, 2) nonskeletal grainstone, 3) fenestral nonskeletal-skeletal packstone-grainstone, 4) fenestral nonskeletal packstone-grainstone, 5) pisolitic packstone-grainstone, 6) peloid mudstone-wackestone, 7) siliciclastic sandstone,

8) siliciclastic-rich peloid wackestone.

The outer shelf strata are composed of Rock Types 1 and 2. These rocks consist of well sorted, well rounded grains with abundant cross lamination and parallel lamination. Evidence of emergence is absent. The outer shelf strata are interpreted to have been deposited in a subtidal, non-emergent carbonate sand shoal environment.

The shelf crest strata are composed of Rock Types 3, 4, and 5. Fenestral fabric, sheet cracks, tepees and pisolites are the predominant elements of the shelf crest rocks. The shelf crest strata were the result of peritidal deposition on a "tidal flat" on the lee side of the carbonate sand shoals. This supports Esteban and Pray's (1975, 1976, 1977) interpretation of peritidal deposition for shelf crest strata.

The inner shelf strata are composed of Rock Type 6. These rocks may be laminated or burrowed; some lamination may be of algal origin. Locally, crystallotopic or scattered nodular molds of evaporites, gypsum and anhydrite, occur. Evidence of prolonged subaerial exposure is absent. The inner shelf strata were deposited in a restricted, probably hypersaline, shallow lagoon.

Sandstone rocks of the shelf strata are classed as Rock Types 7 and 8. The author suggests that the siliciclastic grains were transported in highly concentrated sediment dispersions from which they were deposited in water depths of 3 to 6 m.

This study supports Esteban and Pray's (1975, 1976, 1977) interpretation of pisoliths as primary particles deposited subaqueously in

a peritidal environment. Pisolitic beds are largely localized in the intertepee depressions. The intertepee depressions served as "mega-splash" cups for pisolite formation.

Tepee formation was syndepositional. It is the result of a two-phase process, an early phase of desiccation contraction and a later phase of expansion from crystallization of carbonate cements.

Two forms of cyclic sedimentation are present, sandstone-carbonate cycles, and cycles within carbonate strata. The changes in sea level responsible for the sandstone-carbonate cycles were probably on the order of 5 to 10 m; cycles within the carbonate strata were the result of sea level changes on the order of 1/2 to 2 m. Changes in sea level resulted from episodic shelf subsidence. This study supports Pray's (1977) suggestion that the sandstone units were deposited at high sea level stages and the carbonate units at low sea level stages.

The lateral extent and the abundance of pisolites and tepees within the shelf crest varied with time. Three informal units within the study interval, the Hairpin Dolomite, the Triplet Dolomite, and the Basal Dolomite, are characterized by different amounts of pisolites and tepees. The Hairpin Dolomite is characterized by a widespread abundance of pisolites and tepees; the Triplet Dolomite has a paucity of pisolites and tepees; the Basal Dolomite contains an amount that is between these two extremes. The author interprets these differences to be the result of different lengths of time shelf crest sediments spent in the peritidal environment, which is a function of the rate and frequency of shelf subsidence.

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INTRODUCTION

Purpose

Numerous geologists have ventured into the Guadalupe Mountains of west Texas and New Mexico to examine exposures of a now-classical carbonate shelf-to-basin transition in the Permian Reef Complex. Despite the considerable amount of study, information is still needed on many problems. The problems related to shelf strata are: 1) genesis of the pisolite facies, 2) depositional environment(s) of the shelf sandstones 3) depositional profile across the shelf and shelf margin, 4) types of cyclic sedimentation, 5) range of sea level fluctuation responsible for cyclic sedimentation, 6) relative importance of vadose meteoric versus submarine diagenetic processes.

Previously, broad Guadalupian age regional facies have been described by Lloyd (1929), King (1948), Newell et al. (1953), Kendall (1969), Tyrrell (1969), Dunham (1972), and Smith (1974b). These studies were concerned with most of the stratigraphic interval present in the Guadalupe Mountains. However, detailed sedimentologic studies of the various Guadalupian age rock types and stratigraphic units composing these regional facies have not been done. The objective of this study was to determine, through study and tracing of individual beds, the detailed facies mosaic of a limited stratigraphic interval of carbonate shelf strata adjacent to the Capitan "reef". This detailed investigation would yield some of the information necessary to solve the problems described above.

Geologic Setting

The Guadalupe Mountains are a triangular block, tilted gently eastward and bounded on the west by a zone of large normal faults (Figure 1). Relief on the southeastern side of the Guadalupe Mountains reflects the original depositional topography from the shelf to the basin. The Capitan or Guadalupe Escarpment, on the southeast side of the block (Figure 1), forms part of the rim of the Delaware Basin, a Permian intracratonic basin (Figure 2). Most of the Guadalupe Mountains are part of the Northwest Shelf. Strata in the Guadalupe Mountains are Leonardian and Guadalupian in age.

During Guadalupian time deposition of the predominantly carbonate rocks of the Northwest Shelf occupied a narrow 15 to 25 km wide belt on the margin of the Delaware Basin. Shelfward the carbonate belt grades into an evaporite belt 50 to 100 km wide. Further shelfward the evaporites grade into continental red shales and sandstones. Basinward the carbonate shelf rocks grade into the rocks of the Delaware Basin, which are predominantly sandstone with minor amounts of interbedded limestone (Figure 3).

This study focuses on approximately 45 m of stratigraphic section, centered about the contact between the Yates and Tansill Formations (Figure 4). The Yates and Tansill Formations are composed of shelf strata deposited on the Northwest Shelf. The Yates Formation is composed of interbedded sandstones and carbonates. The overlying Tansill Formation and underlying Seven Rivers Formation are predominantly carbonates. Basinward the Yates and Tansill Formations

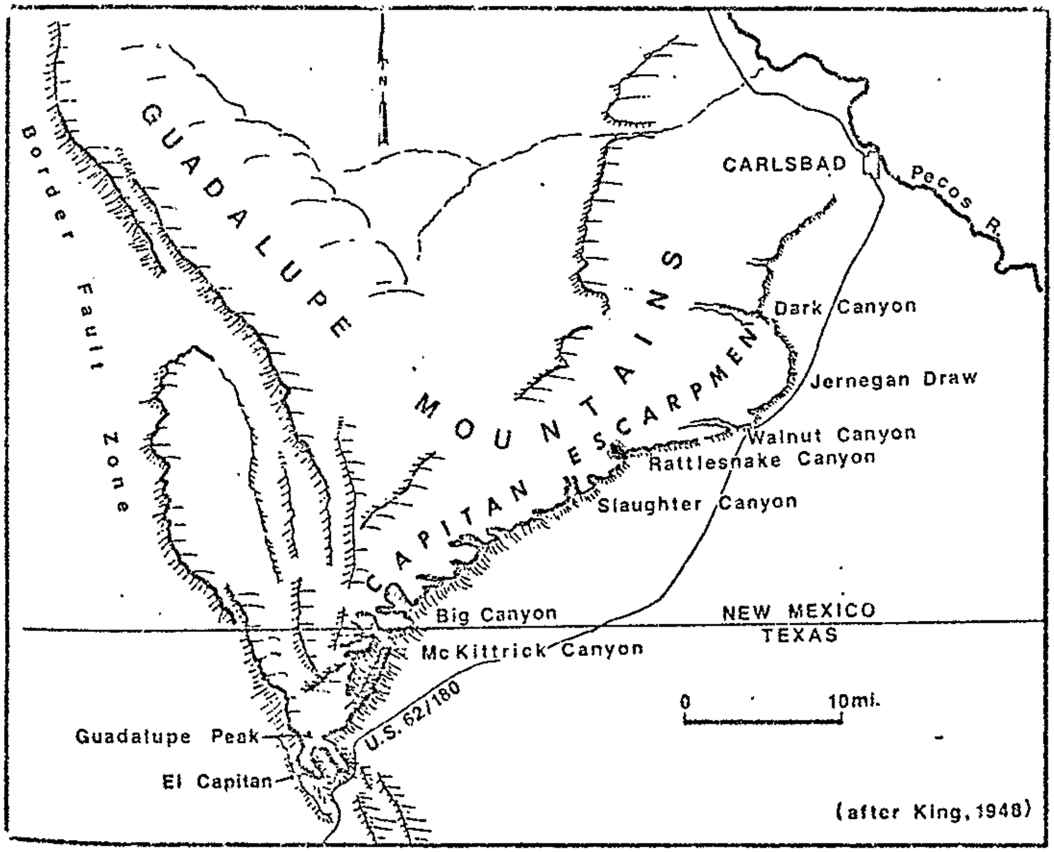

Figure 1. Physiography of the Guadalupe Mountains, Texas and New Mexico. Study area colored green.

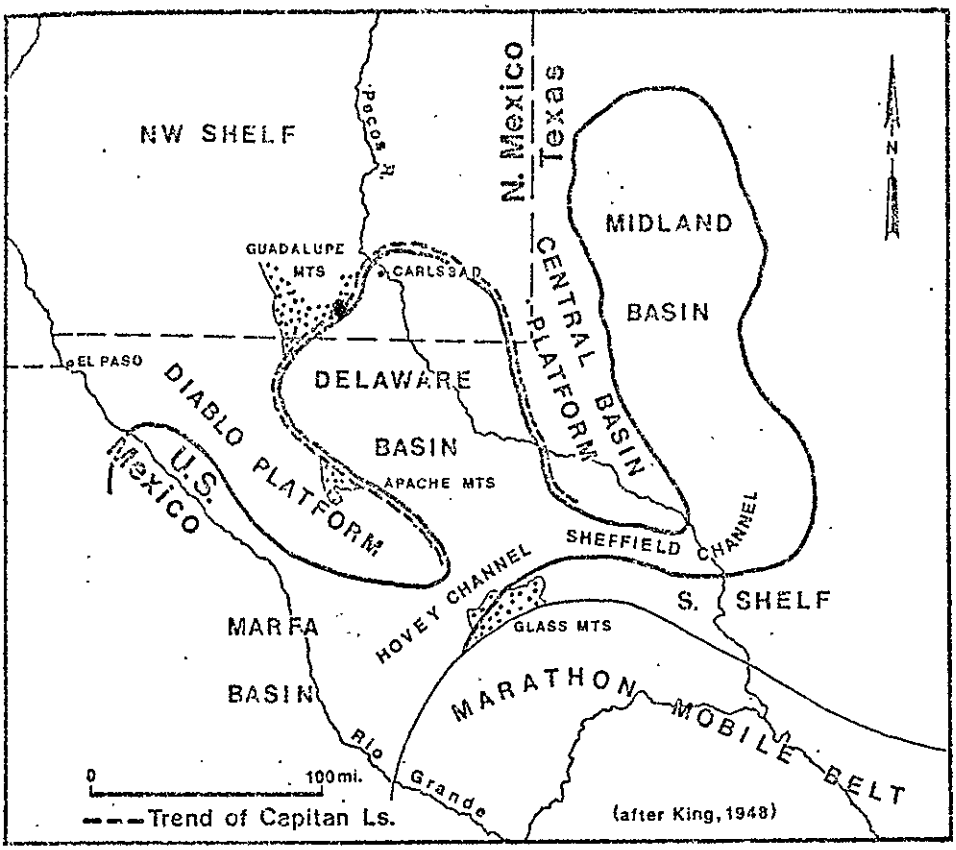

Figure 2. Structural provinces of the Permian basin complex, Texas and New Mexico. Study area colored green.

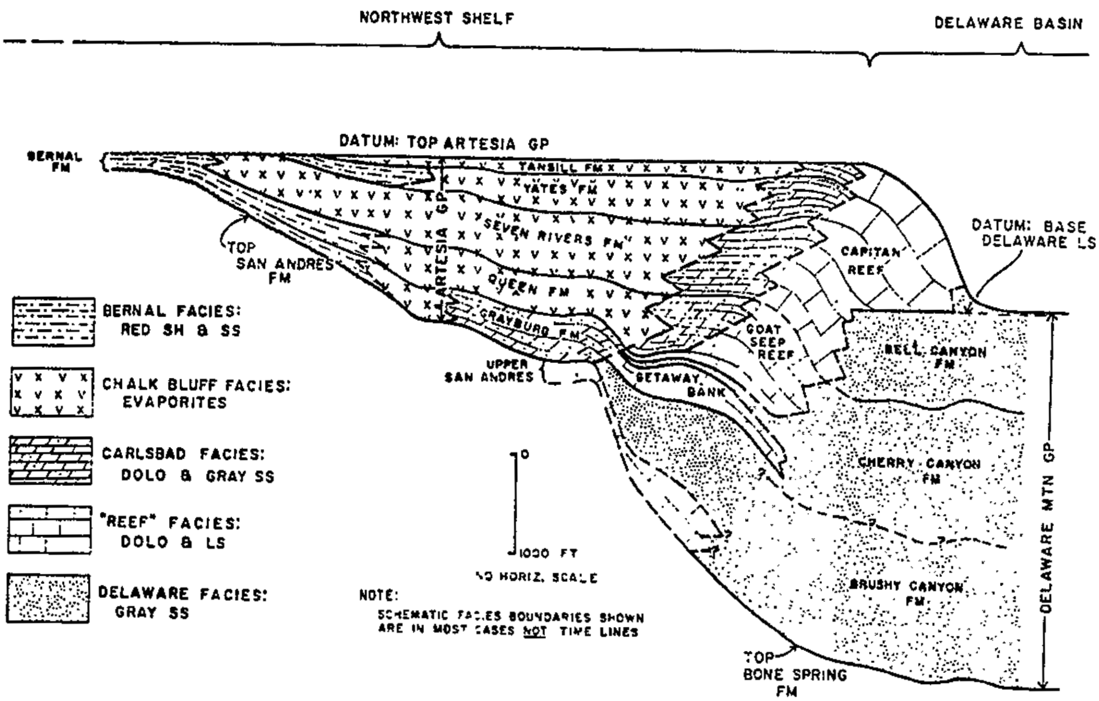

Figure 3. Stratigraphic cross section of the Northwest Shelf and the Delaware Basin showing the regional facies. After Meissner (1972). Study area colored green.

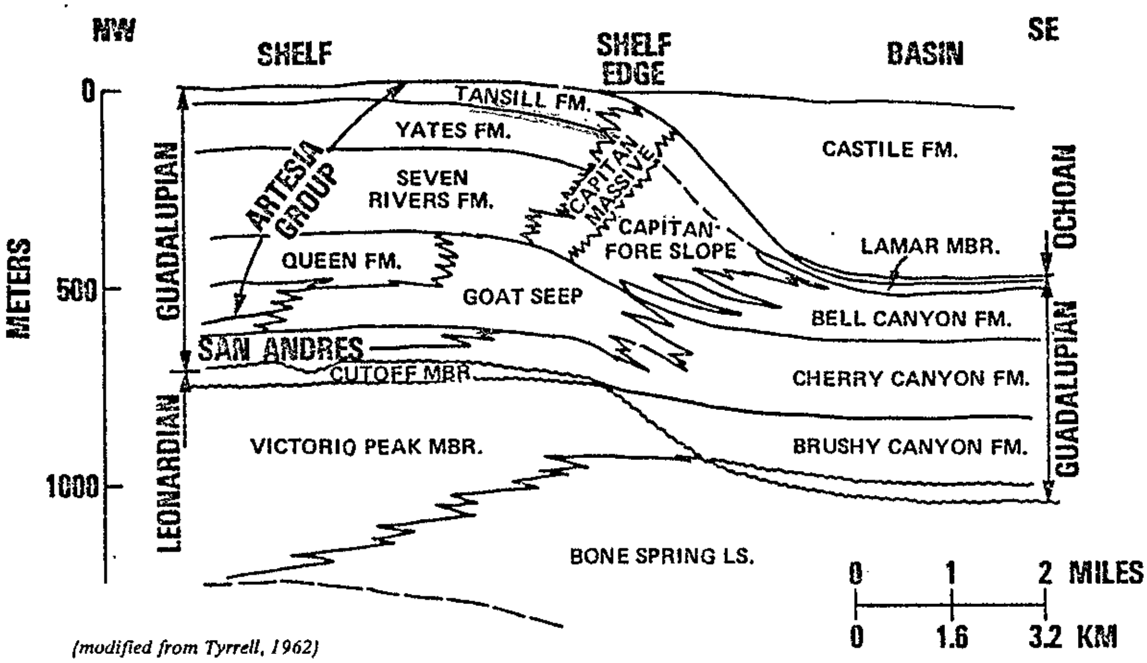

Figure 4. Stratigraphic cross section of the Northwest Shelf and the Delaware Basin. Study area colored green.

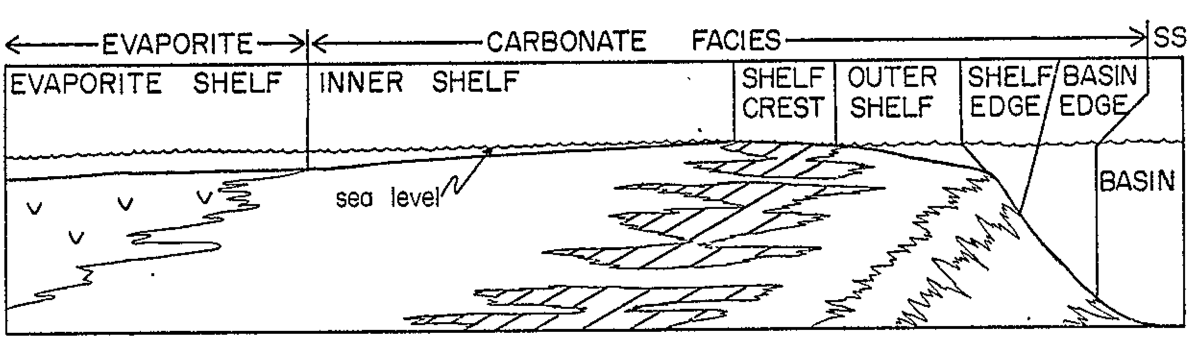

Figure 5. Proposed profile across the Guadalupian carbonate shelf. After Pray (1977).

grade abruptly into the Capitan Limestone. The Capitan Limestone comprises the shelf edge or shelf margin facies. This study is focused on the most basinward 2 1/2 km of shelf strata, an area of rapid facies changes.

This area of rapid facies changes was divided by Pray (1977) into five carbonate facies (Figure 5). These facies listed in a basinward direction are: 1) inner shelf, 2) shelf crest, 3) outer shelf, 4) shelf edge, and 5) basin edge. One main focus of this study is the transition from the inner shelf across the shelf crest to the outer shelf.

Excellent exposures of the shelf and shelf edge rocks occur in the many canyons dissecting the southeast side of the Guadalupe Mountains. Stratigraphic sections were measured in Rattlesnake Canyon and some reconnaisance work was carried out in Walnut Canyon (Figure 6). Both of these canyons are located in the Carlsbad Caverns National Park in southeastern New Mexico. Rattlesnake Canyon was chosen as the study area because of the excellent and continuous exposures of the Yates and Tansill Formations along a line perpendicular to the shelf edge.

The stratigraphic interval studied was divided into three informal rock units; the Hairpin Dolomite, the Triplet Unit, and the Basal Tansill Unit (Figures 7 and 8). These names were used by the author and other members of the Wisconsin carbonate research group working in the Guadalupe Mountains (Esteban and Pray, 1977). The stratigraphic section in Figure 7 was measured in Walnut Canyon,

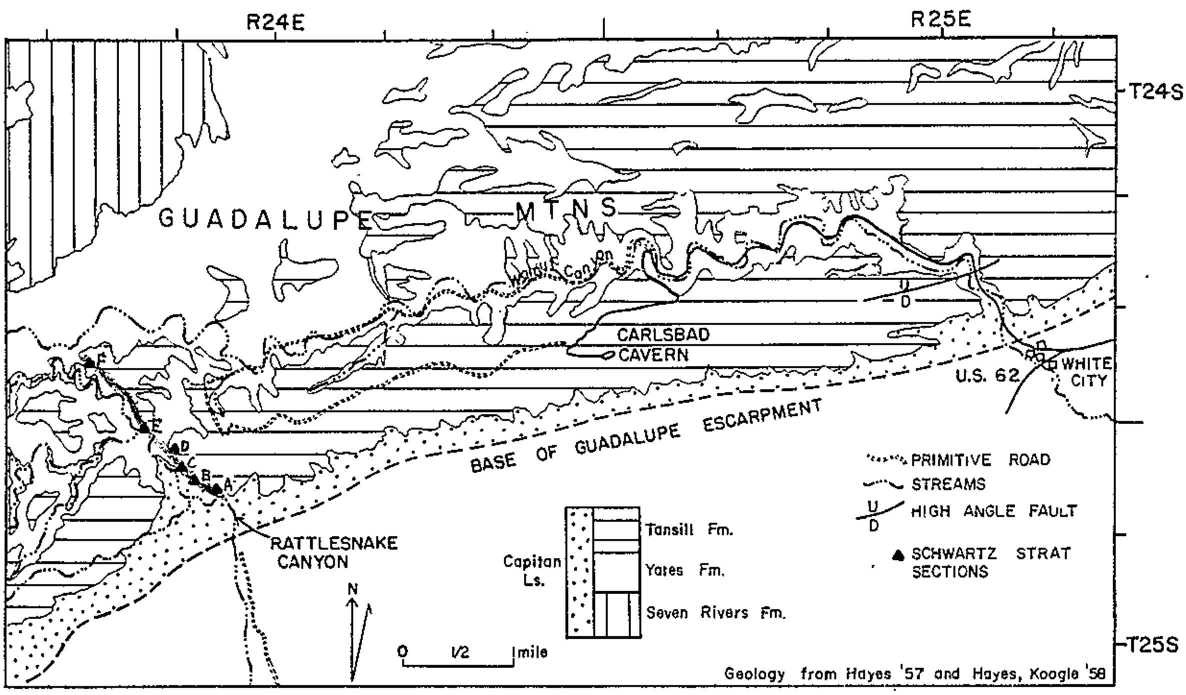

Figure 6. Geologic map of the study area and locations of measured sections. Study area of author colored green.

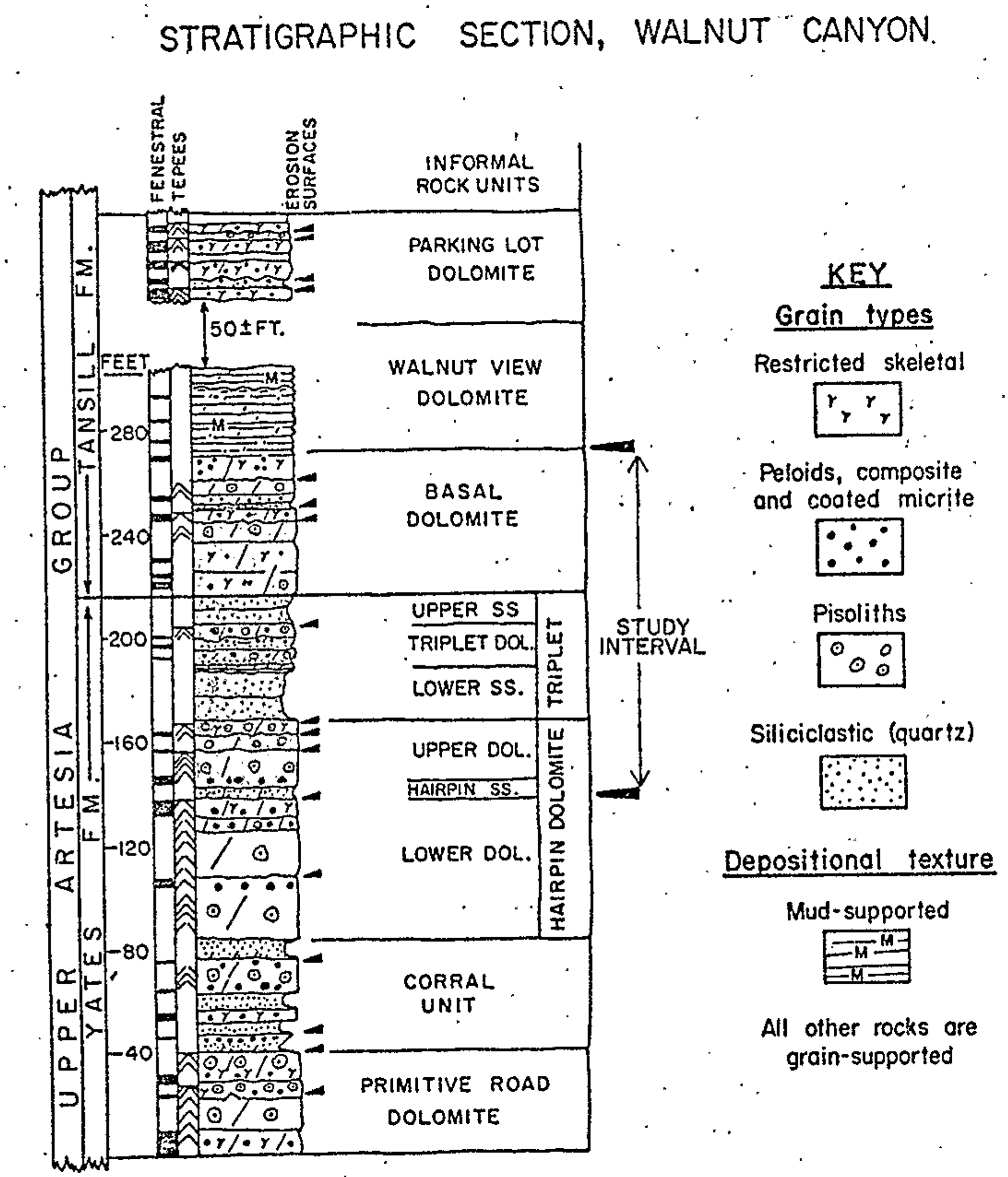

Figure 7. Generalized stratigraphic section of the Yates and Tansill Formations showing the informal division into smaller units. Section measured in center of the pisolite belt in Walnut Canyon. Taken from Esteban and Pray (1977). Study interval colored green.

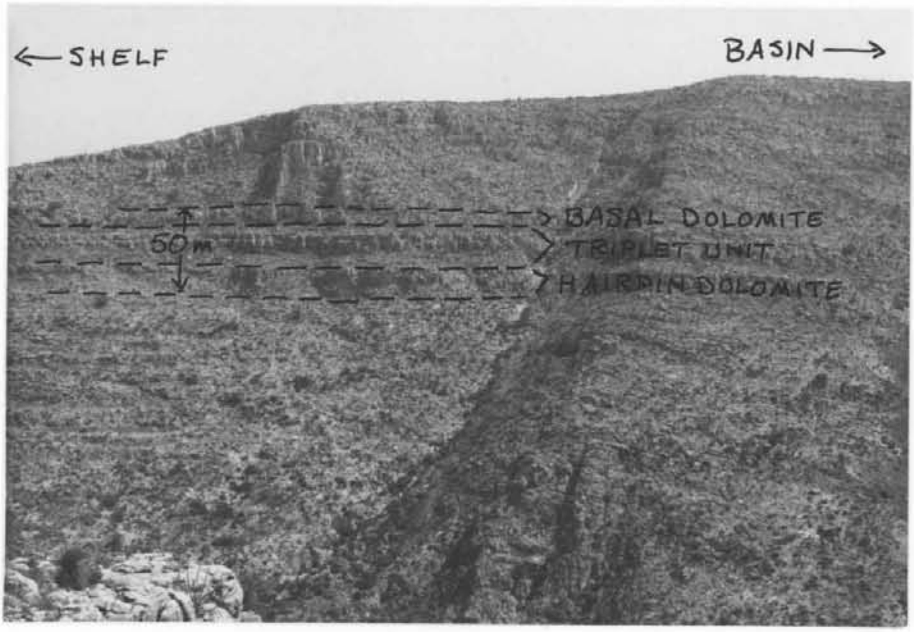

Figure 8. Photograph of the Yates and Tansill Formations showing the informal division into the Hairpin Dolomite, Triplet Unit, and Basal Dolomite. Photograph is of the northeast side of Rattlesnake Canyon about 3/4 km shelfward of the Capitan Escarpment. The basin is to the right and the shelf is to the left.

1.6 km shelfward of the Capitan Limestone and 7.2 km east of Rattlesnake Canyon, along approximate facies strike with measured section D in Rattlesnake Canyon. Boundaries between these three units are at contacts between carbonate and sandstone beds. Two sandstone units and two sandy carbonate units in the study interval provide excellent markers and are continuous over most of the study area, persisting to within 200 m of the Capitan Limestone, where they thin and grade abruptly into the shelf edge carbonates.

Strata in most of the study area are undeformed with dips of a few degrees or less. This made correlation in the study area relatively easy. However, in Rattlesnake Canyon 1.1 and 1.2 km shelfward of the Capitan Escarpment, two previously unmapped faults occur with a down-thrown block between them (Figure 9). The faults trend parallel to the Capitan Escarpment and have stratigraphic displacements of approximately 16 m. King (1948) determined the age of the faulting in the Guadalupe Mountains to be late Pliocene or early Pleistocene. Use of the sandstone markers and the carbonate lithology permits accurate correlation across the faults.

Field and Laboratory Methods

Field work was conducted during the months of June, July, and August of 1976 and during a 2 week period in January of 1977. Six stratigraphic sections were described and measured in the field in the Rattlesnake Canyon area (Figure 6). Reconnaissance work was conducted in the Walnut Canyon area. The reconnaissance work was assisted by Douglas Neese, Mateu Esteban, Lloyd Pray and other mem-

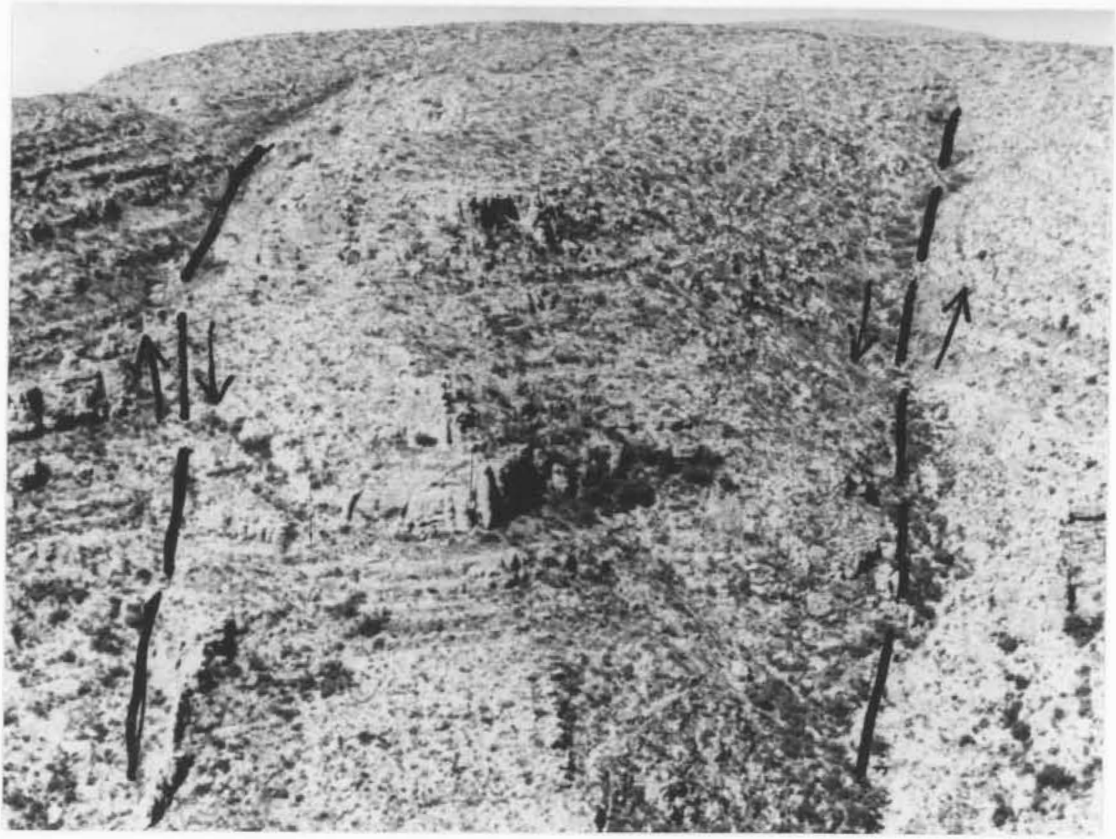

Figure 9. Photograph of two faults in Rattlesnake Canyon. The faults strike parallel to the Guadalupe Escarpment. The faults are located between measured sections B and C. Arrows indicate relative motion along the faults. In this photo the basin is to the left and the shelf is to the right.

bers of the Wisconsin carbonate research group. Douglas Neese conducted field work in Walnut Canyon on a complementary Masters thesis during the same time period as the author (Neese, 1979).

Approximately 300 hand samples were collected. These were slabbed, polished, and examined with respect to their depositional and diagenetic fabrics. Carbonate rocks were classified using the system proposed by Dunham (1962). Sandstones were classified using the system proposed by Dott (1964). Thin sections, approximately 100, were examined using the petrographic microscope. Some thin sections of carbonates were stained with a double stain of alizarin red-S and potassium ferricyanide, to differentiate calcite from dolomite and to detect the presence of ferrous iron (Dickson, 1965). Some thin sections of sandstones were etched with hydrofluoric acid and stained with sodium cobaltinitrite solution to detect the presence of potassium feldspar (Bailey and Stevens, 1960). Figured specimens are on file, thesis number UW 1666, at the Department of Geology, University of Wisconsin, Madison, Wisconsin.

Previous Work

The amount of published literature concerned with the Permian Basin complex is extensive. Philip B. King (1948) wrote a professional paper that dealt with the stratigraphy, tectonics, mapping and many other geologic aspects of the southern Guadalupe Mountains. This now-classic work, served to spark the interests of many geologists in the geology of the Guadalupe Mountains. Hayes and Koogle (1958), Motts (1962), and Hayes (1964) extended the mapping of the Guadalupe Mountains that Philip King had studied. Tyrell (1969), using fusulinid zonation, focused on the difficult problem of correlation from the shelf to the basin.

The book by Newell _et al_. (1953) discussed in detail the sedimentologic and paleoecologic aspects of the Permian reef complex. The Capitan Limestone was interpreted to be a barrier reef by Newell _et al_. (1953). The Newell _et al_. (1953) book further stimulated work on the facies and interpretation of the depositional profile across the shelf and shelf edge.

Studies by Boyd (1958), Kendall (1969), Silver and Todd (1969), Tyrrell (1969), Dunham (1972), Meissner (1972), and Smith (1974b) investigated the depositional and diagenetic environments of broad regional facies. These authors also described the sandstone-carbonate cycles in the shelf strata. Smith (1974b) described cycles within the carbonate strata.

The work by Kendall (1969), Dunham (1972), and Smith (1974b) also dealt with the diagenetic environments of the facies. Dunham's work stressed the importance of vadose meteoric processes during the early diagenesis of these rocks. He also proposed that the paleotopographic crest of the Capitan Reef Complex was in the pisolite belt of the shelf strata and not in the organic "reef" rocks of the shelf edge.

Recently, the University of Wisconsin carbonate research group under the direction of Lloyd Pray has been directing its attention to the sedimentology problems of the Permian Reef Complex exposed in the Guadalupe Mountains. Babcock's (1974) and Yurewicz's (1976) doctoral theses focused on the Capitan Massive. Both theses emphasized the interpretation that the Capitan Limestone was not a barrier reef but formed at a submerged shelf edge, and also stressed the importance of submarine processes during the early diagenesis of the Capitan Massive. My thesis was conducted concurrently with that of Douglas Neese (Masters thesis, 1979). The author and Douglas Neese (1979) were concerned with depositional and diagenetic environments of shelf strata. Early results from these theses were published in a joint article (see appendix for a copy of the publication, Neese and Schwartz, 1977). Sarg's (1976) doctoral thesis was a study of the transition from the evaporite belt to the carbonate belt. At the present time the Goat Seep Formation is under study by Allan Crawford. Neil Hurley (1978) recently finished a study on shelf strata in a position on the depositional profile equivalent to the study area of the author but in an older part of the stratigraphic section.

For a more extensive listing of the literature, a selected bibliography of Guadalupian literature is presented in the 1977 Field conference guidebook of the Permian Basin section of the Society of Economic Paleontologists and Mineralogists.

Acknowledgements

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DEPOSITIONAL PARTICLES

Introduction

Depositional particles can be classified as either nonskeletal or skeletal. Dunham (1972), Smith (1974b), and Esteban and Pray (1977) have documented that in a shelfward direction, the proportion of carbonate nonskeletal to skeletal particles increases. These same authors also documented that particle size decreases in a shelfward direction. Particle types are here defined and described to clarify the descriptions of rock fabrics and rock types given later.

Nonskeletal Particles

Nonskeletal particles are the most abundant type in the shelf strata. Definitions of nonskeletal particles are troublesome because different authors will use different names for the same particle. Another problem with nonskeletal particles is that authors will differentiate two particle types on the basis of a size limitation. Different authors will use different size limitations. The author will attempt to give the best definition and will note synonyms where appropriate.

Peloids - Peloids are grains composed of an aggregate of cryptocrystalline carbonate. This term can be used irrespective of the size, shape, or origin of the grain (McKee and Gutschick, 1969). In my study area, I use the term peloids for cryptocrystalline grains that range from 0.2 to 5 mm in size.

Pellets - "A grain composed of micritic (cryptocrystalline) material, lacking significant internal structure and generally ovoid in shape. Most pellets are very fine sand to coarse silt size grains", (Leighton and Pendexter, 1962). Folk (1962) describes pellets as, "..... rounded, spherical to elliptical or ovoid aggregates of microcrystalline calcite ooze, devoid of any internal structure". This term will be used here to describe the smaller cryptocrystalline grains that range in size from 0.03 to 0.2 mm. Pellets are always rounded spherical to elliptical grains. Peloids can be any shape.

Lumps; Aggregates; Aggregate Lumps; Grapestones - These terms are probably the most troublesome to define. Milliman (1974) defines "aggregate" as a grain that contains two or more fragments joined together by a cryptocrystalline matrix which constitutes less than 50% of the grain. Illing (1954) first recognized these particles in Holocene sediments and called them "lumps". He defined several types. The term grapestone was reserved for those lumps from which round grains protrude, resembling a bunch of grapes. Milliman (1974) proposed that "lump" be used to refer to particles in which more than half the grain is matrix. None of these terms will be used in this paper, instead the term intraclast defined below will be used.

Intraclasts - Folk (1962) defined this term as "fragments of penecontemporaneous, generally weakly consolidated carbonate sediment that has been eroded from adjoining parts of the sea bottom and redeposited to form a new sediment." Folk included Illing's "grapestone" as intraclasts.

Lithoclasts - Folk (1962) described these as fragments of consolidated limestone eroded from ancient limestone outcrops on an emergent land area. The author believes that this definition should be altered by omitting the word "ancient". Evidence now indicates that lithification of carbonate sediment can occur very rapidly and can be considered as penecontemporaneous.

Ooid - Newell et al. (1960) defined an ooid as a particle that has one or more regular lamellae formed as successive coatings around a nucleus and in which, the crystals of the lamellae must show a systematic crystallographic orientation with respect to the grain surface. Ooids are defined to be less than 2 mm in diameter.

Pisolites - Pisolites are spherical to irregular shaped particles with a multiple laminated coating and a diameter greater than 2 mm. Pisolith nuclei can be skeletal or nonskeletal and can be composed of single or multiple particles.

Coated Grains - Coated grains are irregularly shaped particles with an accretionary coating.

Micrite - Micrite refers to clay sized carbonate particles. The upper size limit of micrite is 4 microns in diameter.

Siliciclastics - This term refers to all the noncarbonate particles. Within the carbonate rocks quartz is the predominant siliciclastic mineral and feldspar occurs in small quantities. Within the sandstone rocks there are several siliciclastic minerals. These are described in the section on rock types.

Skeletal Grains

Skeletal grains are most abundant in the outermost or basinward portions of the shelf. Although they are abundant in some beds the diversity of the biota is low. This is an indication that salinity and water circulation on the shelf were variable, probably due to deposition at or near sea level.

Dasycladacean Algae - Grains of dasycladacean algae, a green algae, are the most abundant skeletal particles in the rocks of the study area. The two major genera of dasycladacean algae recognized in the study area are *Mizzia* and *Macroporella* (Johnson, 1942). Ecologic and environmental studies of modern dasycladacean algae indicate that they occur in shallow protected marine lagoons (Ginsburg *et al.*, 1971). See Figure 10.

Calcispheres - Calcispheres are single-chambered, hollow calcareous spheres with a cryptocrystalline wall structure and an average diameter of 70 to 100 microns (Bathurst, 1971). Both spinose and non-spinose forms are present. Currently, calcispheres are interpreted to be the reproductive spores of dasycladacean algae (Wray, 1977). Most fossil calcispheres are reported from rocks interpreted as deposited in very shallow marine water with restricted circulation (Wilson, 1975). See Figure 11.

Foraminifera - Two general types of forams are recognized, encrusting and fusulinids. These foraminifera are discussed in detail because of their unusual morphology and their importance in correlation. Two genera of encrusting forams have been identified, *Tuberitina* and *Hemigordius* (Figures 12 and 13). Generally, these forams are found unattached. *Tuberitina* can often be mistakenly identified as a calcisphere. However, *Tuberitina* has a larger

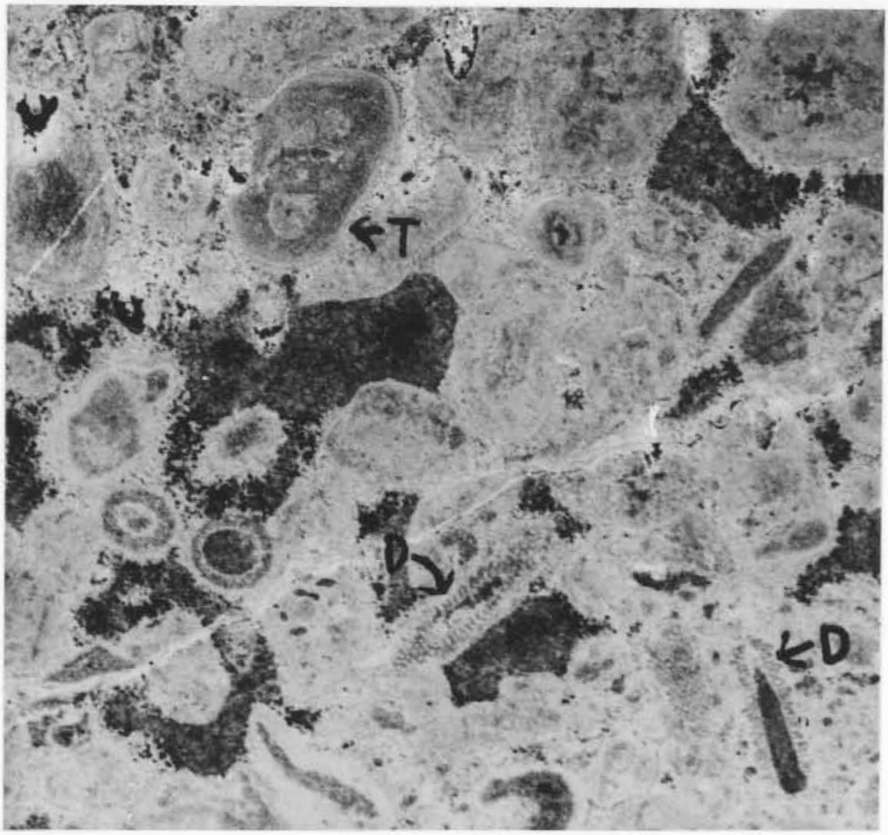

Figure 10. Negative print of a thin section of a skeletal grainstone, Rock Type 1, showing dasycladacean algae (D) and <u>Tubiphytes?</u> (T). (UW 1666/1).

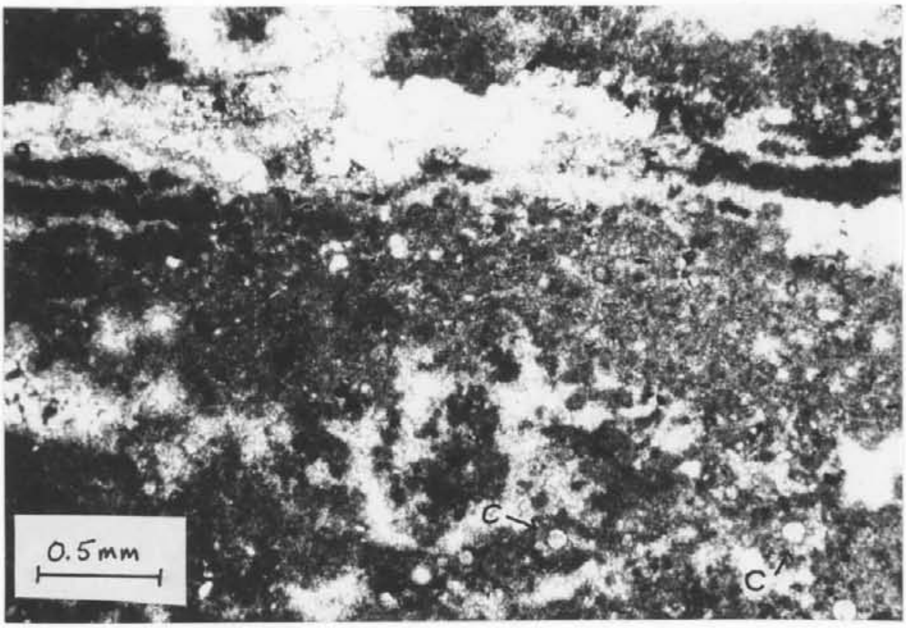

Figure 11. Thin section photomicrograph of a fenestral nonskeletal packstone, Rock Type 4. This particular sample is composed of pellets predominantly. Spinose calcispheres (C) are also present in a small amount. Fenestrae are partially filled with microspar cement. (UW 1666/2).

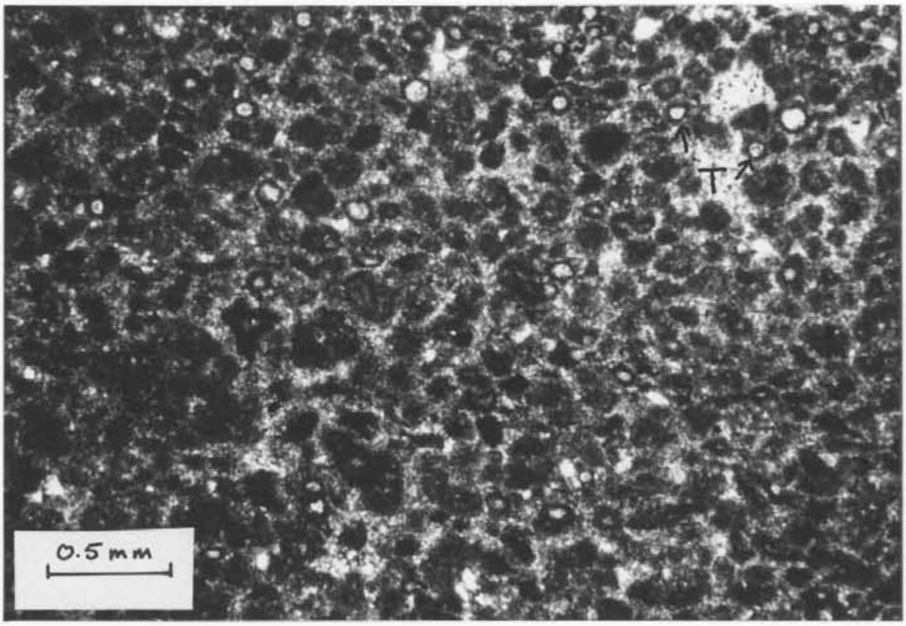

Figure 12. Thin section photomicrograph of a peloid packstone, Rock Type 4. Rock is cemented with microspar. Note the Tuberitina (T). (UW 1666/3).

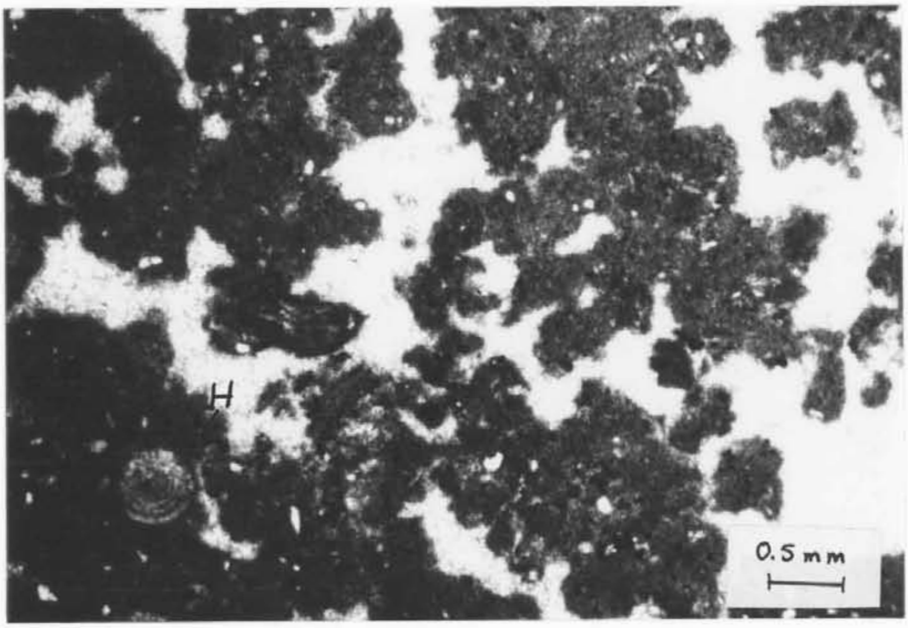

Figure 13. Thin section photomicrograph of a porous fenestral nonskeletal packstone, Rock Type 4. Porosity is fenestral, some solution enlargement of fenestrae is also present. Note the <u>Hemigordius</u> (H) foram. (UW 1666/4).

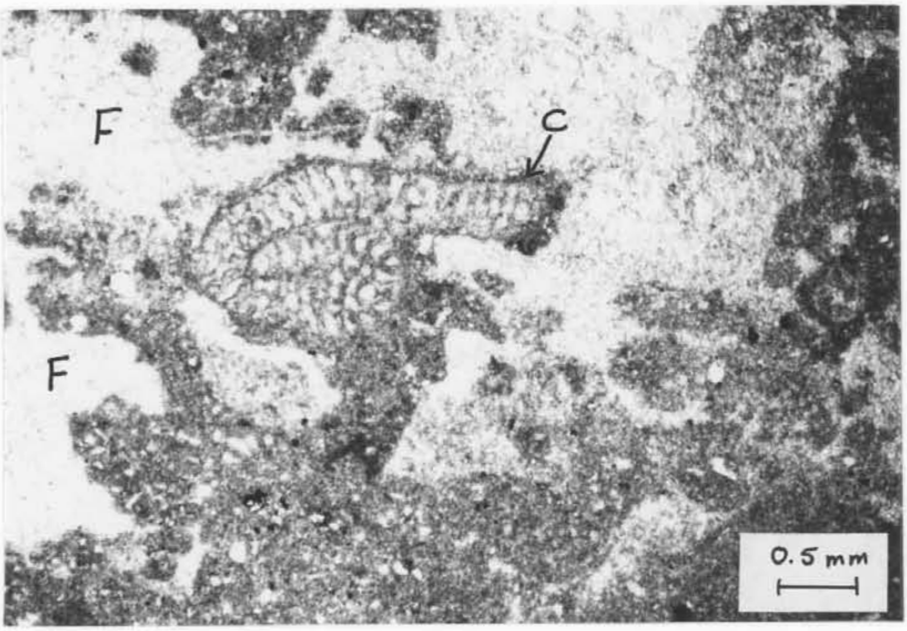

Figure 14. Thin section photomicrograph of a fenestral nonskeletal-skeletal packstone, Rock Type 3. Fenestrae (F) are filled with microspar cement. Note the Codonofusiella fusulinid (C). (UW 1666/5).

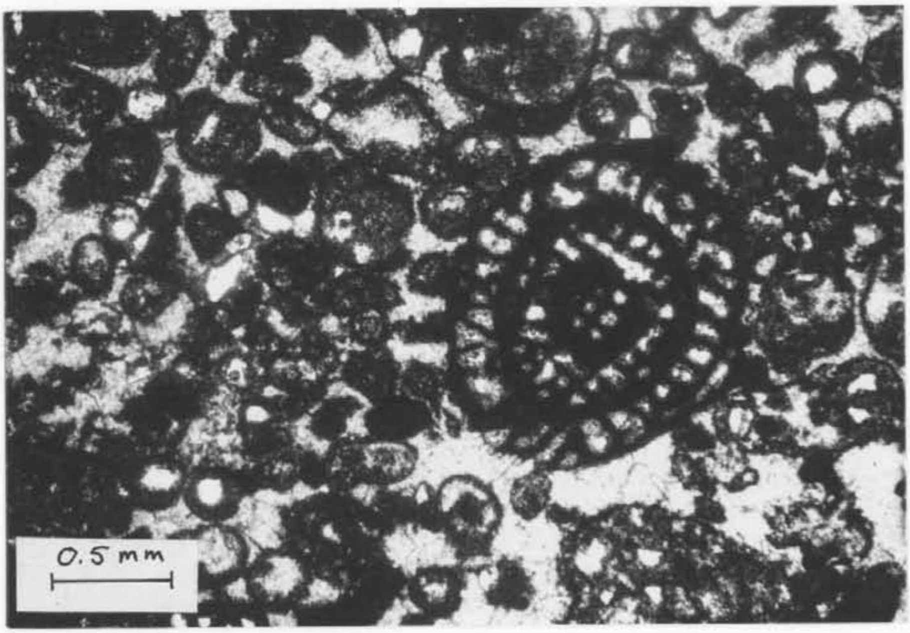

Figure 15. Thin section photomicrograph of a nonskeletal grainstone, Rock Type 2. Most particles are peloids. Some appear to be aggregate grains. Note the Yabeina fusulinid. Cement filling pores is equant spar. Within the fusulinid the pores have an isopachous cement lining. (UW 1666/6).

diameter, 170 to 400 microns, is non-spinose with perforate test walls, and is often multichambered with 2 to 4 bulbous chambers arranged uniserially (Toomey, 1972).

Two genera of fusulinids have been identified, Codonofusiella and Yabeina (Figures 14 and 15). For a systematic description of these fusulinids the reader is referred to Skinner and Wilde (1955). For a discussion of fusulinid zonation in the shelf strata the reader is referred to Tyrrell (1969).

Miscellaneous Skeletals - Other skeletal grains are present in minor or trace amounts. They include gastropods, brachiopods, pelecypods, green algae, ostracods, echinoderms, sponges, bryozoans, and Tubiphytes. These miscellaneous skeletal grains most commonly occur as fragmented particles, although whole gastropods are common.

ROCK FABRICS

Introduction

Rock fabric can be defined as the orientation and arrangement in space of various elements that comprise a rock. In carbonates the major fabric elements are the grains, matrix, cements, porosity, and internal sediment. It is the rock fabric which yields the best information on the depositional and diagenetic environments. Rock fabrics present in strata of the study area are described below. Discussion of their occurence within the "back reef" or shelf strata is reserved for later in the report.

Fenestral Fabric - Fenestral fabric is the term applied to a rock in which fenestrae are present. Tebbutt et al. (1965) defined fenestrae as, "primary or penecontemporaneous gaps in a rock framework, that are larger than grain supported interstices". Choquette and Pray (1970) modified this definition to make it also applicable to mud-supported rocks by defining fenestrae as any primary opening larger than the normal interparticle openings. A fenestrae may be an open space or it may be partially or completely filled with cement and/or internal sediment. (Figures 16 and 17).

Currently, the accepted interpretation is that fenestrae can form by several processes: 1) molds of algal mats, 2) alternate wetting and drying of sediment producing planar shrinkage cavities, 3) trapped air or gas bubbles producing spherical or irregular cavities (Boyd, 1975).

Figure 16. Outcrop photograph showing fenestral fabric in a fenestral nonskeletal packstone. Outcrop located in measured Section C. White circle is 2 cm in diameter.

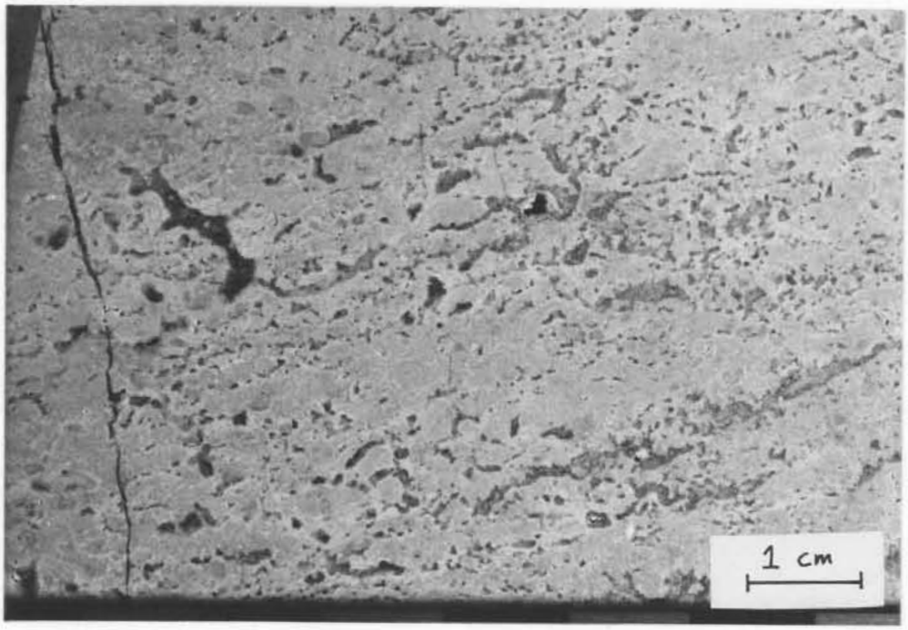

Figure 17. Polished slab showing fenestral fabric. (UW 1666/7).

Examination of modern carbonate environments show that fenestral fabric occurs more commonly in supratidal sediment, less commonly in intertidal sediment, and is apparently absent in subtidal sediment (Shinn, 1968). However, Pray (personal communication, 1978) has informed the author that Logan has found fenestral fabric in subtidal sediments in Shark Bay, Australia. The author still interprets abundant fenestral fabric to be a useful criterion for recognition of periodic subaerial emergence.

Algal Lamination - In the field some rocks locally exhibit a crinkly lamination that I interpret as evidence of blue green algal mats. Thin slit-like pores are locally present parallel to the lamination. The laminae are approximately 1 mm thick (Figures 18 and 19). The algal mats range from 10-30 cm thick. Evidence of algal filaments can be seen in thin section. In Holocene sediments blue green algal filaments comprise most algal stromatolites. Using the classification of Logan et al. (1964) based on the morphology of the algal mat these algal laminated rocks would be classed as LLH, laterally linked hemispheroid stromatolites. Much attention has been given to stromatolite morphology and its relationship to its depositional environment. The Guadalupian stromatolites have a morphology similar to Holocene stromatolites of the Persian Gulf described by Kendall and Skipwith (1968), and Kinsman and Park (1976). This stromatolite morphology has a depth range from shallow subtidal to upper intertidal.

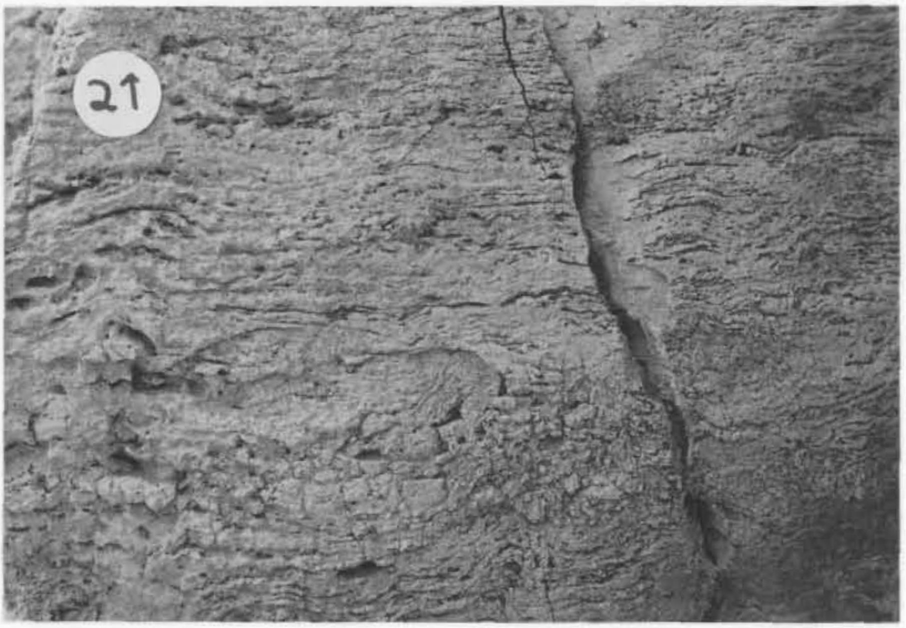

Figure 18. Outcrop in Section C showing algal mat lamination. White circle is 2 cm in diameter.

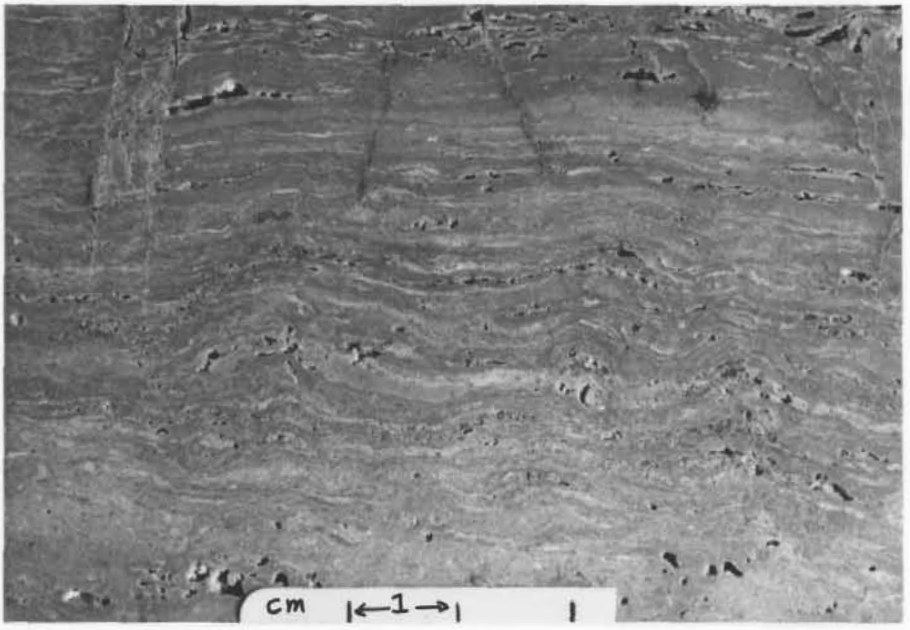

Figure 19. Polished slab showing algal mat lamination. (UW 1666/8).

Burrows - Burrows occur within some of the non-pisolitic rocks. In the field, burrows appear as unfilled, branched or unbranched curved tubes. Smith (1974b) interpreted most of the burrows to be U-shaped. However, the author did not see any evidence of either U-shaped or J-shaped burrows. The burrows are 2 to 5 mm in diameter, up to 10 cm in length, and most are neither vertical nor horizontal but are irregularly inclined. (See Figure 20). Esteban and Pray (1977) suggested the possibility that some tubular fenestrae, 1 to 2 mm in diameter might be burrows. In polished slabs, a few spreite-filled burrows were found.

Garrett (1977) described similar unfilled burrows in Holocene intertidal pond sediments on Northwest Andros Island, Bahamas. These burrows were made by a polychaete belonging to the genera Dasybranchus. Seilacher (1967) and Rhoads (1975) have noted that vertical or inclined burrows are dominant in strata deposited in the intertidal zone. Horizontal burrows along bedding surfaces are present in neritic environments. Horizontal burrows have not been noted in the study area. Comparison of the Guadalupian age burrows with Holocene burrows and burrows of other ages indicate that the Guadalupian burrows probably originated in the intertidal or shallow subtidal environment.

Evaporite Fabrics - Two types of evaporite molds are present, crystallotopic and nodular. Crystallotopic molds show the crystal form of the original evaporite mineral, gypsum or anhydrite. Nodular molds are irregular shaped equidimensional molds. (Figure 21).

Figure 20. Outcrop of mixed nonskeletal-skeletal packstone showing burrowing. Outcrop occurs in Section D. White circle is 2 cm in diameter.

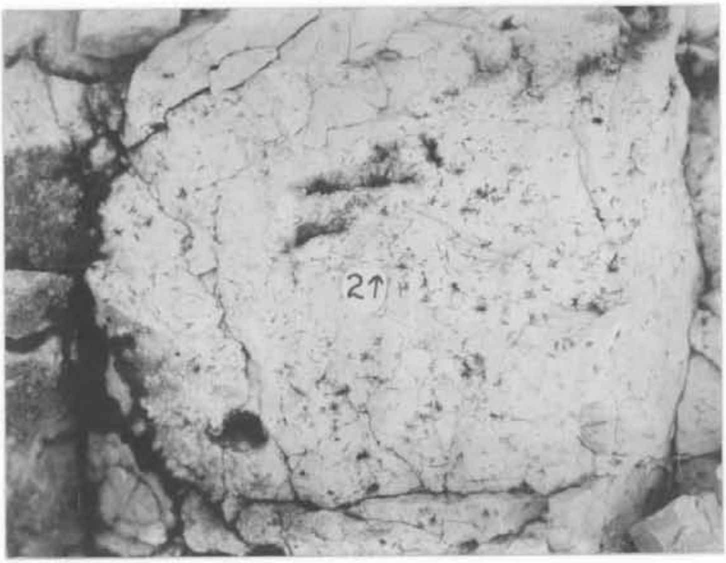

Figure 21. Outcrop of peloid dolomite wackestone showing crystallotopic and nodular evaporite molds. Photograph taken by Douglas Neese in Walnut Canyon. White circle is 2 cm in diameter.

In the study area the evaporite fabrics occur as either open pores or as molds filled with sparry equant calcite cement. The evaporite fabrics have been recognized only in mudstones and wackestones of the study area.

Sheet Cracks - Fischer (1964) defined sheet cracks as horizontal shrinkage cracks or openings. Esteban and Pray (1977) redefined the term so that it would encompass all non-tectonic fractures (Dunham's term, 1968) of shrinkage or expansion origin. Most sheet cracks are horizontal or parallel to bedding. High angle or vertical sheet cracks are much less frequent than those parallel to bedding and more commonly cut across the grains. (See Figures 22 and 29). Fischer (1964) interpreted sheet cracks as shrinkage structures that resulted from desiccation of sediments deposited in the intertidal environment. Smith (1974a) interpreted some of the sheet cracks in the Permian strata as the result of expansion due to forces exerted by growing aragonite and evaporite crystals trapped in intertidal sediments. Both Smith and Fischer consider sheet cracks to be an intertidal phenomenon. Esteban and Pray (1975, 1976, 1977) have suggested that sheet cracks may also have formed in a shallow subtidal environment. In the Permian strata most sheet cracks are complexly filled with eogenetic cements and/or internal sediment.

Tepees - Tepees are nontectonic anticlinal structures. A typical tepee has a characteristic symmetrical chevronlike structure with bedding inclined upward toward a subvertical axial fracture. There is some interdigitation of truncated beds from opposite sides of

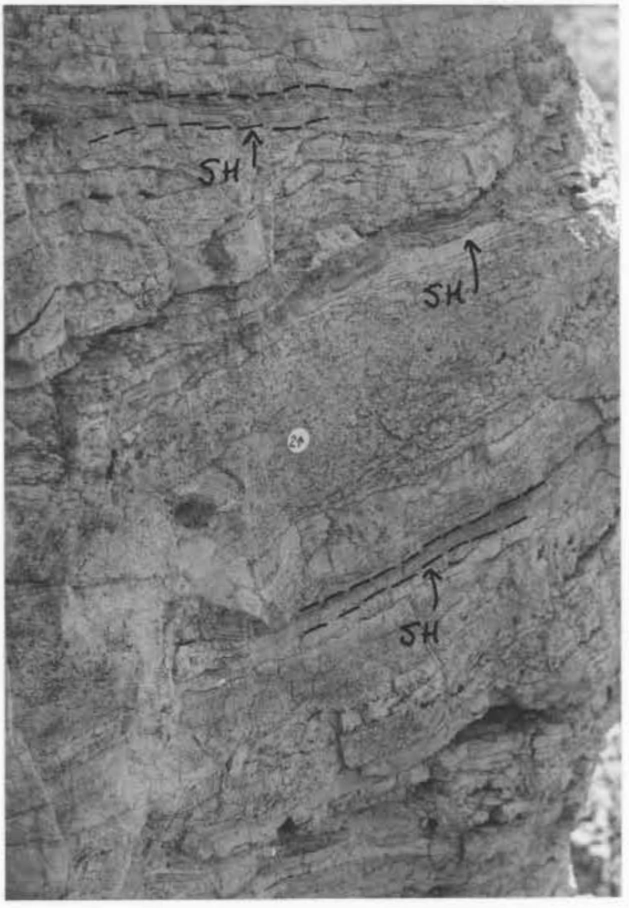

Figure 22. Outcrop showing left flank of a small tepee with sheet cracks (SH). Outcrop occurs in section C. White circle is 2 cm in diameter.

the axial fracture. Dips of beds within tepees are generally less than 50 degrees, but some beds do approach vertical. Tepees range from 1/4 to 2 m in height. (Figures 22, 23, and 24.) Esteban and Pray (1977) reported that commonly the spacing of tepees is related to the size of the tepees. The smaller tepees are closely spaced and the larger tepees are widely spaced. Tepees also have a tendency to be stacked.

The larger tepees occur in a belt 1/2 to 3/4 km shelfward of the Capitan Massive Limestone. Their size decreases away from this belt. In the Guadalupe Mountains the tepee crests cannot be traced in plan view but random strikes suggest polygonal patterns. In Jurassic strata in Morocco, Assereto and Kendall (1977) documented tepee crests forming megapolygons, 1 to 6 m in diameter. Sheet crack fracturing of beds in tepees is common. These sheet cracks are usually filled in a complex manner with eogenetic cement and/or internal sediment.

The core of most tepees is composed of fenestral laminated rocks and sheet cracks filled with cement and internal sediment. Sediments in intertepee depressions generally wedge out against the flanks of the tepees. This indicates that tepee formation was syndepositional. Esteban and Pray (1977) reported that the pisolite-rich strata are generally localized in the intertepee depressions.

Holocene tepees have been documented in three different environments. Shinn (1969) documented subtidal tepees in the sediments of the Persian Gulf. Price (1925), Reeves (1970) and other authors have documented tepees in subaerial terrestrial calcrete soils, caliche.

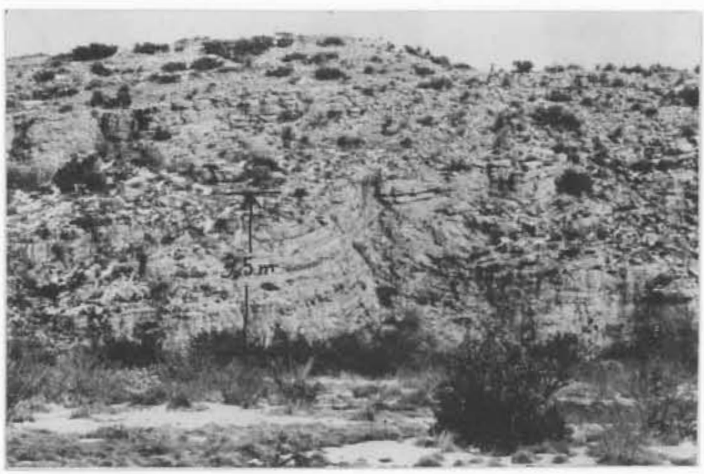

Figure 23. Outcrop of large tepee in Yates Formation. Outcrop occurs 1/4 km east of Rattlesnake Canyon and along strike with Section F.

Figure 24. Outcrop showing a small tepee. Outcrop occurs at Section C. White circle is 2 cm in diameter.

Evamy (1973) reported peritidal tepees in the sediments of the Persian Gulf.

Comparison of the fabric of the different types of Holocene tepees to the tepees in the Permian strata suggested to Assereto and Kendall (1977) that the Permian tepees are of a peritidal origin. Subtidal tepees are generally composed of skeletal calcarenites with a complex history of submarine cementation and boring. Fenestral fabric was not observed in subtidal tepees. Caliche tepees also lack fenestral fabric.

Smith (1974a) and Dunham (1972) interpret tepee structures to be the result of expansion. Smith (1974a) and Esteban and Pray (1977) interpret the cause of expansion as the result of forces exerted by penecomtemporaneous growth of aragonite or possibly magnesian calcite cement during lithification. Assereto and Kendall (1977) interpreted the formation of tepees as a two-phase process, an early desiccation contraction phase and a later expansion phase. In an early phase desiccation contraction, in the peritidal zone, produced the polygonal pattern of tepee crests. The author interprets the desiccation to have resulted from periodic subaerial exposure. I suggest that the exposure was longer in duration than that needed to produce fenestral fabric. Fractures, unfilled sheet cracks, formed by this desiccation contraction are then enlarged during the later expansion phase of tepee formation. They are enlarged by moisture swelling and by the force exerted during crystallization of carbonate cements. The desiccation phase of tepee formation occurs in the intertidal and/or supratidal environments. The expansion phase of tepee formation can occur anywhere in a peritidal environment. The author suggests that the large amounts of cements associated with tepees indicates a subtidal dominance during the expansion phase.

DIAGENESIS

Introduction

Diagenesis within shelf strata is extensive. It was not the purpose of the writer to deal with all of the aspects of diagenesis or to treat them in detail, but my observations permit some diagenetic interpretations. The numerous articles on the diagenesis of the Permian strata testify to its complexity. In the section on rock fabrics, the diagenetic fabrics, tepees, sheet cracks, and evaporite molds were briefly discussed. In this section, major aspects of the cement, porosity, and internal sediments associated with these features and with the strata will be discussed.

Cements

The most spectacular cements are developed within the large (5-10 cm) sheet cracks in tepee cores. Cement types present in sheet cracks are 1) radial fibrous hemispheroids, 2) mammillary fibrous crust, 3) laminated crusts and 4) sparry equant calcite. Assereto and Kendall (1977) described the fibrous cements as pseudofibrous. They described pseudofibrous cement as, "...sparry calcite pseudomorphing elongate fibrous crystals, which are hexagonal in cross section and have square or feathery terminations". Folk and Assereto (1974) interpreted the fibres' square terminations as an indication that the fibrous crystals were originally aragonite that has inverted to calcite.

Cements within voids other than sheet cracks form smaller masses and are not as spectacular but are quantitatively more important in the rock. The cements within interparticle, intraparticle, and fenestral pores are generally dolomite. The dolomitization has destroyed most of the original crystal fabric. However, ghosts of the original fabric are present, four cement textures occur: 1) isopachous bladed to fibrous, 2) microcrystalline, 3) microspar, 4) sparry. In the more basinward strata the cement is calcite.

Sheet Crack Cements

Radial Fibrous Hemispheroids - The radial fibrous hemispheroids are composed of calcite fibrous crystals or dolomite ghosts of fibrous crystals, 10 to 30 mm long that form a radiating fan. This is the "druse fan" cement of Dunham (1972). These hemispheroids are found on the floors and roofs of sheet crack cavities. I observed that these hemispheroids are more common in the basinward portions of the shelf. Some hemispheroids occur on a floor of internal sediment and are subsequently overlain by a layer of internal sediment. The penecontemporaneous nature of the internal sediment, as shown by the presence of Guadalupian age biota (see discussion on internal sediment), indicates that the radial fibrous hemispheroids are of penecontemporaneous origin. Where present, this cement type is usually the earliest cement precipitated in a sheet crack. However, it is not present in all sheet cracks. Dunham (1972, Figures II-11, II-15, and II-16) and Esteban and Pray (1977, Figure II-12c) have excellent photos of this cement type.

Ginsburg and James (1976) reported the occurrence of radial fibrous hemispheroids of aragonite, ("botryoidal aragonite", their term), within cavities in Holocene reef wall limestones in Belize, British Honduras. These radial fibrous hemispheroids line the floors, sides, and roofs, of cavities within the reef wall limestone.

Radial fibrous hemispheroids are abundant within the Capitan Massive Limestone. Babcock, Pray, and Yurewicz (1977) and Schmidt (1977) interpret these hemispheroids to be of a subtidal marine origin. Dunham (1972) compared the morphology of this cement to stalactite and stalagmite growths formed in fresh water vadose conditions in caves. The writer considers the evidence of a submarine origin more acceptable. The radial fibrous hemispheroids in sheet cracks were probably precipitated during a rise in sea level that flooded the tepees.

Mammillary Fibrous Crusts - Mammillary fibrous crust cement consists of calcite laminae and dolomite ghosts of laminae of fibrous crystals oriented perpendicular to the surface of deposition. Individual laminae range in thickness from 0.1 to 5 mm. The crust may be as thick as 4 cm. This cement type commonly, but not everywhere, forms an isopachous lining on the walls of a sheet crack. Dunham (1972) noted that some of these fillings are asymmetric, the floor deposits are thicker than the roof deposits and have more internal sediment in them. However, close examination in the field reveals that the reverse situation can also occur, roof deposits thicker than floor deposits (cf. Esteban and Pray, 1977; Figure II-12a). Internal sediment commonly overlies or is incorporated in the crusts on the floors of sheet cracks. The penecontemporaneous nature of the internal sediment indicates that this is an eogenetic cement.

Dunham (1972) compared the morphology of this fibrous cement to growths lining the walls of caves. On the basis of their similarity he interpreted the fibrous crust cements to have precipitated in fresh water vadose conditions. Assereto and Kendall (1977) noted that this cement was similar to the aragonite cements that encrust grains in a beach rock. The writer feels that the isopachous nature of most of these fibrous crusts indicates a phreatic subaqueous origin. Folk (1974) proposed that the formation of aragonite fibres generally requires water with a Mg/Ca ratio of 2/1 or more. This condition most commonly occurs in marine to hypersaline waters.

Laminated Crust Cement - Laminated crust cement consists of alternations of calcite or dolomite micritic crinkled laminae and microspar layers. The crinkled laminae are 0.02 to 0.5 mm thick and the microspar layers are 0.1 to 3 mm thick (cf. Assereto and Kendall, 1977). Crusts may be as thick as 3 cm. In the field this type of cement could easily be mistaken for a laminated sediment. However, the laminated crust cement can be found lining the roofs of cavities. Dunham (1972, Figures II-15 and II-16) has two photographs of this cement type.

In some localities this type of cement does not line sheet cracks, but forms a coating over several particles. The laminated crusts may form a mammillary or botryoidal coating that is immediately overlain by sediments. Examination of geometric relations (cf. Esteban and Pray, 1977; Figure V-2b and V-2c) indicate that the laminated crust cement probably formed a coating or encrustation on the depositional surface. Esteban and Pray (1977) have documented that their botryoidal cement is thickest and most abundant high on the flanks of tepees and is mostly associated with pisolitic rocks.

The writer interprets the laminated crust cement to be equivalent to Assereto and Kendall's (1977) festooned cellular crust cement and Esteban and Pray's (1977) botryoidal cement. Relationships with internal sediment indicate that the laminated crusts formed early.

Assereto and Kendall (1977) noted that this type of cement had features similar to those described for Holocene supratidal splash or spray zone cements of the Persian Gulf. I was not able to confirm this observation. These cements in the Persian Gulf encrust the surface and cavities within beach rock (Purser and Loreau, 1973).

Assereto and Kendall (1977) reported that the microspar laminae commonly show ghosts of fibrous crystals. They postulated that the microspar resulted from neomorphism of an original, possibly aragonitic cement.

Sparry Equant Calcite - The sparry equant calcite cement is composed of calcite crystals ranging from 0.03 to 1 mm in diameter. It occurs as the last cement to fill the pores. The origin of this cement is probably the same as that of the sparry equant cement found in other types of pores. Its origin is discussed in the next section.

Other Cement Types (Those not found in sheet cracks)

Isopachous Bladed to Fibrous Cement - The isopachous bladed to fibrous cement consists of ghosts of fibrous to bladed crystals that range from 0.1 to 1 mm in length. (See Figures 25 and 26). The isopachous cement preferentially occurs in the coarser grained rocks, and the more basinward rocks. James et al. (1976) reported the occurrence of Mg-calcite isopachous bladed cement in shallow reefs of Belize, British Honduras. Submarine aragonitic isopachous fibrous cement has been documented by James et al. (1976) and Schroeder (1972). Evamy (1973) reported the occurrence of fibrous aragonite crystals isopachously lining the pores of intertidal rock.

Microcrystalline Cement - Microcrystalline cement is composed of micrite-sized crystals, ranging from 1 to 4 microns in diameter. The cement forms rims around pores and around grains and therefore, can probably be identified as a chemical precipitate and not a depositional clastic element, (See Figures 27 and 28). This cement is more common in the coarser grained rocks. Microcrystalline cement has been found in the Holocene in several different environments. Land (1971), Ginsburg et al. (1971), and James et al. (1976), reported Mg-calcite microcrystalline cements in Holocene reefs. Moore (1971), Roberts and Moore (1971), and Taylor and Illing (1971) reported aragonitic microcrystalline cement in Holocene intertidal beachrocks. Steinen (1974) reported microcrystalline cement in the fresh water vadose environment. Dunham (1972) reported microcrystalline meniscus cements within shelf strata in the Guadalupe Mountains. Meniscus cements have been documented in the fresh water vadose zone by Dunham (1971). The author interprets the microcrystalline cement to probably have originated in a variety of environments.

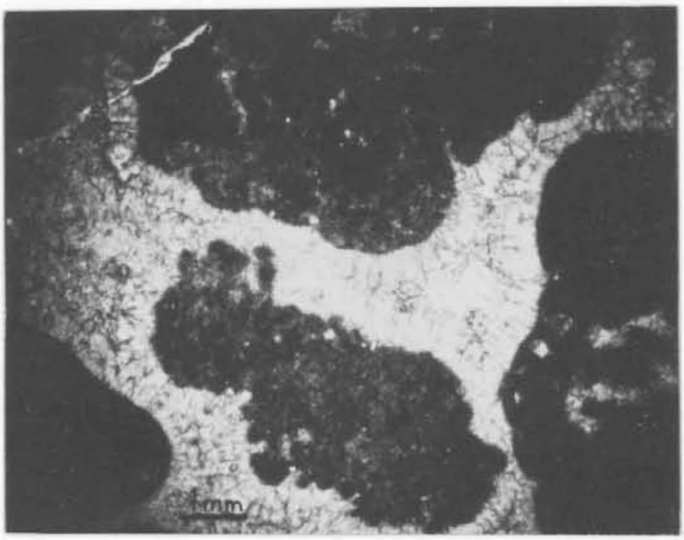

Figure 25. Thin section photomicrograph showing isopachous bladed cement lining a pore in a nonskeletal grainstone, Rock Type 2. (UW 1666/9).

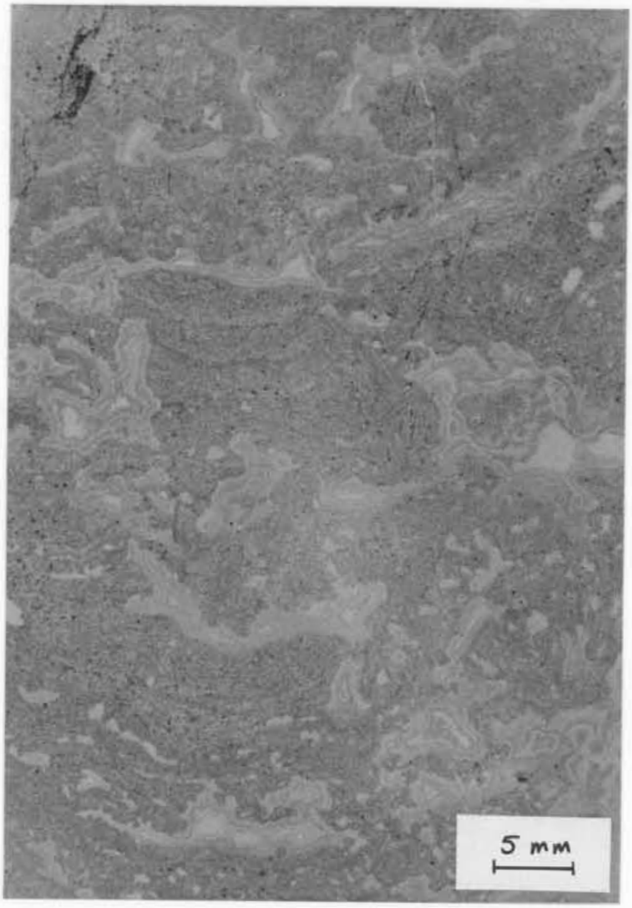

Figure 26. Thin section photomicrograph of a fenestral nonskeletal packstone, Rock Type 4. Fenestrae are lined with an isopachous cement, then filled with a late sparry equant cement. (UW 1666/10).

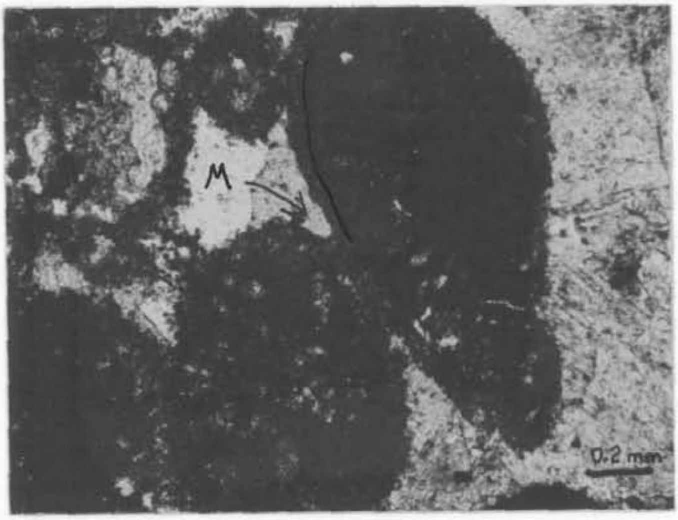

Figure 27. Thin section photomicrograph showing microcrystalline cement (M) and sparry equant cement. (UW 1666/11).

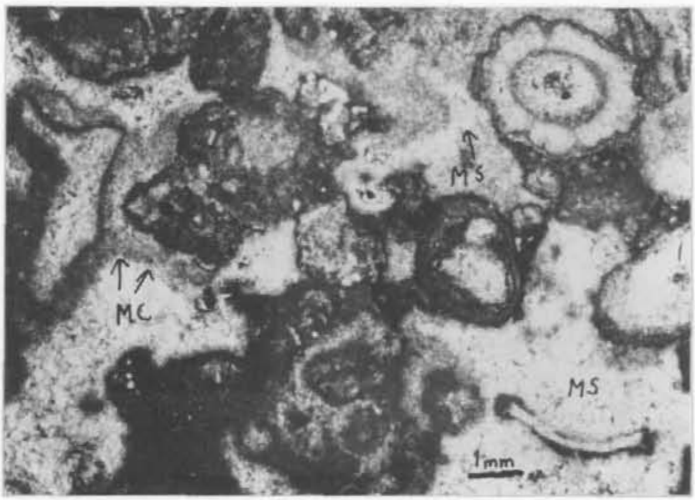

Figure 28. Thin section photomicrograph of a fenestral skeletal-nonskeletal grainstone, Rock Type 3. Pores are filled with microcrystalline cement (MC) and microspar cement (MS). The microcrystalline cement is the darker cement. (UW 1666/12).

<u>Microspar Cement</u> - Microspar consists of subhedral to anhedral equant crystals ranging from 5 to 20 microns in diameter (Folk, 1965). It occurs more commonly in the finer grained rocks. (See Figure 28). Steinen (1974) documented the occurrence of calcite microspar in the freshwater phreatic environment. Folk (1965) interpreted microspar to be the result of neomorphism of micrite. Taylor and Illing (1971) and Shinn (1969) have documented within the submarine environment neomorphism of original aragonite fibrous and micritic cement to Mg-calcite microspar. The microspar cement in the shelf strata of the Guadalupe Mountains probably evolved in a variety of different environments.

<u>Sparry Equant Cement</u> - Sparry equant cement is composed of subhedral to anhedral, equant clear calcite crystals ranging from 30 to 80 microns in diameter. It occurs as the last cement in the larger pores. (See Figure 26 and 27). For a time sparry equant cement was interpreted to have been precipitated in the freshwater phreatic environment. However, Schroeder (1972) reported the occurrence of sparry crystals of Mg-calcite in Holocene, Bermuda cup reefs. Schroeder (1973) also documented sparry calcite forming a meniscus outline in Pleistocene Bermuda reef rock. This indicates that sparry calcite can also form in the vadose environment. Folk (1974) has interpreted sparry equant calcite to form in waters with low Mg/Na ion ratio. This condition is present in the freshwater phreatic environment. Folk also noted that this condition occurs in connate subsurface waters.

Internal Sediment

Internal sediment is common in sheet cracks. It is also present in some fenestral pores. Three types of internal sediment are present: 1) red iron stained dolomite pellets, 2) dolomite pellet silt, 3) microcrystalline dolomite. (See Figure 29). Most of the internal sediment shows faint parallel lamination. In some localities it is normal graded.

Assereto and Kendall (1977) and Dunham (1972) reported finding internal sediment with skeletal grains (dasycladacean algae, ostracods, and *Tubiphytes*) derived penecontemporaneously from the surrounding sediments. As noted in the previous section on cements, internal sediment is commonly interlaminated with cement. The microcrystalline dolomite internal sediment commonly forms the micritic laminae that are present in the laminated crust cement. Internal sediment is not common in the most basinward or shelfward portions of the study area. It is most common in the fenestral and pisolitic rocks of the shelf crest.

Assereto and Kendall (1977) interpreted red iron stained dolomite pellets to be terra rosa. Terra rosa is a red soil which forms from subaerial solution of limestone. Terra rosa is a common fill in shallow subsurface (epigenetic) cracks of emergent carbonate lithified rocks in the Bahamas, Yucatan, and Florida (Roehl, 1967). I was not able to verify this interpretation.

Porosity

Several porosity types are present in the rocks of the study

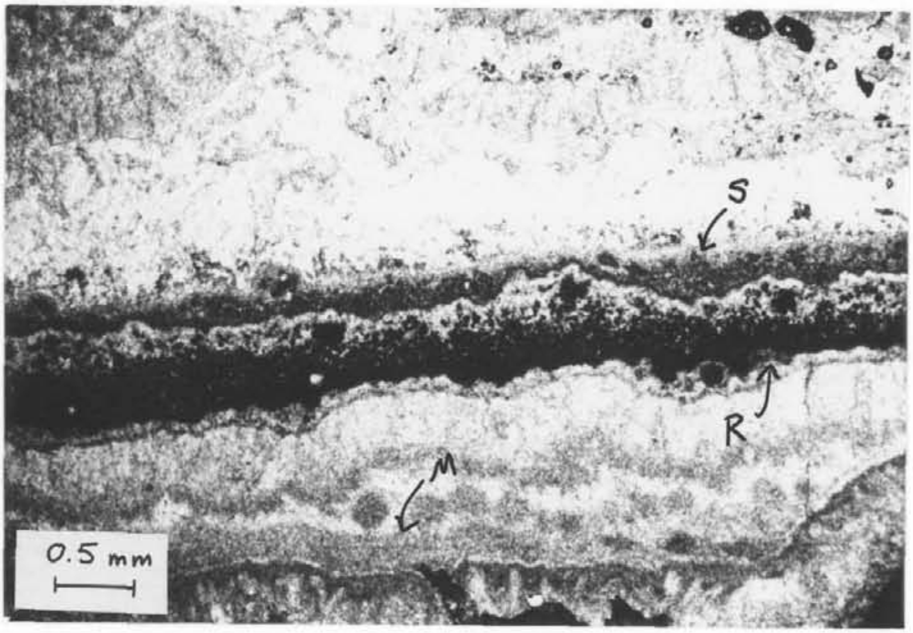

Figure 29. Thin section photomicrograph showing sheet crack filled with cement and three types of internal sediment; red iron stained dolomite pellets (R), dolomite pellet silt (S), and microcrystalline dolomite (M). (UW 1666/13).

interval: 1) fenestral, 2) interparticle, 3) moldic, 4) burrow, 5) intraparticle, 6) vug, 7) sheet crack, and 8) fracture. Only the fenestral and interparticle porosity is present in an amount of economic significance in a petroleum reservoir rock.

Porosity associated with fenestral fabric is the most abundant and best preserved (See Figure 15). Some of the fenestral pores have flat floors. This is the result of internal sediment partially filling the void. Some of the vug porosity was probably originally fenestral porosity that has been enlarged by solution.

Interparticle porosity is most abundant in the coarser grained pisolite strata (See Figure 32). The large size of the interparticle pores in these beds enabled them to remain open during cementation.

Moldic porosity is largely the result of solution of aragonite gastropods, dasycladaceans and evaporite minerals (gypsum and anhydrite). Evaporite molds are crystallotopic or nodular and occur in only the most shelfward strata. See Figure 20 for a photograph. Evaporites are discussed in more detail in the next section.

Burrow porosity was described in the section on rock fabrics. See Figure 19 for a photograph.

Intraparticle porosity occurs in minor amounts and only within the larger gastropods and some dasycladacean aglae.

Vug porosity is not common. Most of the vug porosity appears to be the result of enlargement by solution of other types of pores. See Figure 23 for a photograph.

Sheet crack porosity occurs only in the very large sheet cracks.

Most of the original porosity in the sheet crack is filled with cement and/or internal sediment.

Fracture porosity occurs in small amounts. These fractures probably originated during the uplift of the Guadalupe Mountains in the late Pliocene or early Pleistocene.

Evaporites

Evidence of evaporites, gypsum and anhydrite, is present in the form of crystallotopic and nodular molds (See Figure 20). Crystallotopic molds are the more abundant of the two types. Some of the crystallotopic molds are lens or disc-shaped, suggestive of gypsum. In the study area evaporite molds were found only in measured section F. The molds are randomly distributed in mudstone-wackestone beds approximately 30 cm thick. The beds underlying and overlying will commonly have a crinkly to flat algal lamination or burrows. Beds with evaporite molds grade out basinward into a nonskeletal packstone. Features indicative of dessication, fenestral fabric and mud cracks, are not present in the evaporitic beds.

Kinsman and Park (1976) described stromatolite structures and associated evaporites in Holocene sediments of the Trucial coast, Persian Gulf. They described a smooth mat stromatolite morphology which ranges from shallow subtidal to upper intertidal environments. In the upper 20 to 30 cm of the intertidal zone the precipitation of discoid gypsum destroys all algal structures.

Butler (1969) and Kendall and Skipwith (1969) described the evaporites and the associated fabrics of the supratidal environment of the Trucial Coast, Persian Gulf. Anhydrite is the dominant evaporite mineral, although gypsum is present. The anhydrite occurs as beds and masses with a nodular, nodular mosaic, or a mosaic fabric. Evidence of these fabrics is present in the strata of the study area but is not abundant.

The author interprets the evaporites to have been precipitated from hypersaline brines forming in a restricted lagoon. The evaporites were precipitated within shallow subtidal, intertidal, and low supratidal sediments.

Emergent Surfaces

A major diagenetic feature that is locally present is a weathered zone, calcrete or caliche, 1 to 5 cm thick, at the top of the dolomite unit in the Triplet Unit. This is of Guadalupian age. It is present in measured sections C, D, and E. In the field this zone has a reddish pink color and a micritic appearance. Esteban and Pray (1977) described thin sections from this zone that showed alteration of pisolith fragments, microchanneling and microsparitization. Esteban (1976) described the predominant caliche fabric to be a clotted, peloidal micrite with microspar channels and cracks. (See Figure 30). This description matches the fabric present in the zone at the top of the Triplet Dolomite. This supports the interpretation of this zone as a fossil calcrete.

At the Yates-Tansill contact, at an outcrop along the highway in Walnut Canyon, Esteban and Pray (1977) interpreted lithologic alteration in the underlying rock to be a fossil soil profile.

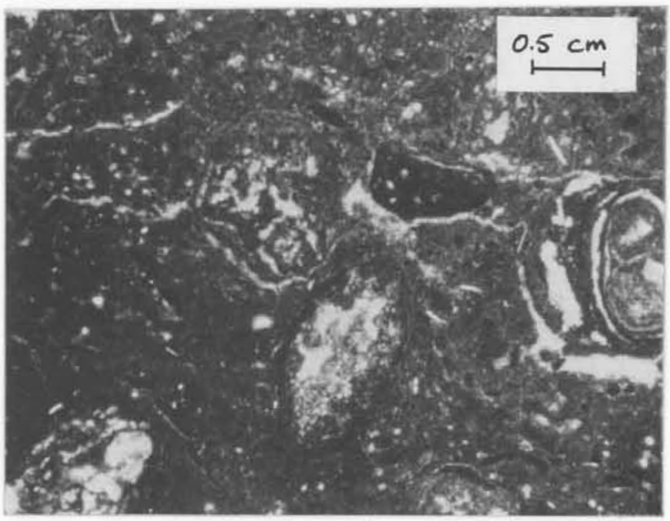

Figure 30. Thin section photomicrograph of a fenestral nonskeletal packstone, Rock Type 4, showing incipient calichefication. Note the microchanneling and clotted texture. (UW 1666/14).

The lithologic alteration occurs as a progressive upward increase in carbonate content in the upper foot of the Upper Triplet sandstone. Part of this carbonate occurs as lensoid nodules, 1 to 3 mm thick. Thin sections of this rock show microspar replacement of quartz grains, mottled micritic texture, microchanneling, and possible calcicutans. Esteban (1977, personal communication) noted that the interpretation of this zone as a fossil soil could only be done using thin sections; field criteria are not sufficient. The author did not find any similar zones in Rattlesnake Canyon. However, this may be due to the recessive weathering of the sandstones.

Thomas (1965), Dunham (1969, 1972), and Smith (1974b) have interpreted the pisolitic rocks to be caliche. Esteban and Pray (1977) interpreted the pisolitic rocks to be of submarine primary particle origin. The author agrees with Esteban and Pray and discusses the subject further in the section on genesis of pisolitic rocks.

Smith (1974b) also described some non-pisolitic caliche in the shelf strata. The non-pisolitic caliche was composed of a partly brecciated upper layer rich in laminar and fibrous carbonate that changes downward via a crudely nodular crumbly intermediate layer into unaltered rock. The author examined several outcrops of the non-pisolitic caliche and concluded that they are not caliche, but are breccias associated with tepees.

Dolomitization

Dolomite rocks are restricted to the shelf strata. The rocks of the most basinward shelf, the shelf edge and the basin are limestones. In the dolomitic rocks all grains are dolomite but the cement may be dolomite and/or calcite. The crystals in grains replaced by dolomite are anhedral and approximately 3 microns in size. This crystal size obliterates some of the microscopic features of grains but does not obliterate the general outline of grains.

The author interprets the dolomitization to be an eogenetic process. Examination of the cements within tepee strata provides the evidence for this conclusion. Two mineralogic types of cements are present within tepee strata: dolomite as interparticle cement and, calcite as sheet crack fill cement. The sheet crack filling cement was precipitated after the lithification necessary for sheet crack formation occurred. Therefore, the dolomitization occurred before the sheet crack cements were precipitated. Dunham (1972) also interpreted the dolomitization to have occurred at this time.

Sandstone Diagenesis

Sandstones are cemented with a dolomite microspar similar to the microspar present in the carbonate rocks (See Figure 31). Locally, there are pockets of calcitic cement. Porosity is present only in trace amounts as interparticle porosity. Examination of thin sections revealed that locally the cement may be replacing the quartz grains.

Hull, Jr. (1957) and Dunham (1972) both noted that there was a narrow belt, 1/4 to 1/2 km shelfward of the Capitan Massive Limestone in which feldspar content was anomalously low. This belt is coincident with the pisolite belt. Outside of this belt feldspar content averages 25%, within this belt the average is 1%. Dunham

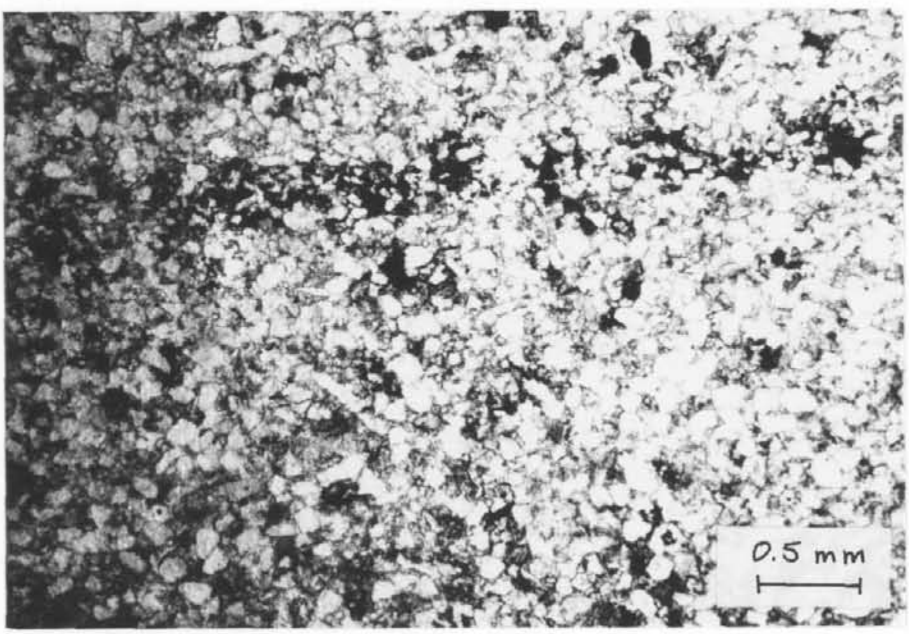

Figure 31. Thin section photomicrograph of a siliciclastic sandstone, Rock Type 7. Microspar cement and iron oxides occur between the grains. The iron oxides are the opaque areas in the photo. (UW 1666/15).

(1972) also examined his samples for kaolinite content. He found that the kaolinite content was higher for rocks within the belt as compared to rocks outside of the belt. Dunham concluded that the kaolinite was probably the result of alteration of the feldspar during subaerial exposure. I did not study this relationship.

ROCK TYPES

Introduction

Rock types were differentiated using several variables, principally: grain type, grain size, grain abundance, rock fabrics, and depositional texture. The rocks were separated into the various types to aid in the recognition of depositional environments and cyclic sedimentation in the shelf strata. Previously, the author described nine lithofacies (see Neese and Schwartz, 1977 in appendix). Some of the nine lithofacies are retained, some are discarded, and some were added. The author will note these changes. The term lithofacies has been discarded here and replaced by the term "rock type".

A total of eight rock types are recognized in the shelf strata Six carbonate rock types occur:

1) Skeletal grainstones

2) Nonskeletal grainstones

3) Fenestral nonskeletal-skeletal packstone-grainstone

4) Fenestral nonskeletal packstone-grainstone

5) Pisolitic packstone-grainstone

6) Peloid mudstone-wackestone

The other two rock types present are siliciclastic-rich:

7) Siliciclastic sandstone

8) Siliciclastic-rich peloid wackestone

These eight rock types are described below. A summary of the characteristics of the different rock types is provided in Table 1.

Rock Type	1	2	3	4
Depositional Features				
Depositional Texture	Grainstone	Grainstone	Grainstones & packstones	Grainstones, minor packstones
Grain Type	75% skeletal	75% nonskeletal	50% skeletal, 50% nonskeletal	75% nonskeletal
Grain Mineralogy	Calcite, minor dolomite	Dolomite, minor calcite	Dolomite	Dolomite
Sedimentary Structures	Cross & parallel lamination	Parallel & minor cross lamination	None	None
Rock Fabrics	None	None	Fenestrae, sheet cracks, tepees	Fenestrae, sheet cracks, tepees
Diagenetic Features				
Mineralogy	Calcite	Dolomite, minor calcite	Dolomite, minor calcite	Dolomite, minor calcite
Cement Type	Isopachous bladed/fibrous, equant spar	Microspar, equant spar	Microcrystalline, microspar, equant spar	Microcrystalline, microspar, equant spar
Porosity	Vug	Vug	Fenestral, burrow, intraparticle, vug, sheet crack	Fenestral, Burrow, intraparticle, vug, sheet crack
Internal Sediment	None	None	Present	Present

Table 1. Summary of Rock Type Characteristics

5	6	7	8
Grainstones, minor packstones	Mudstone & wackestones	Quartz arenite	Wackestones
75% pisolites	Nonskeletal, 75% peloids	Siliciclastics	Siliciclastics & peloids
Dolomite	Dolomite	Quartz, minor feldspar	Quartz & dolomite
Local inverse graded bedding	Local parallel lamination	Local cross & parallel lamination	None
Sheet cracks	Burrows, algal lamination, evaporite molds	None	Fenestrae(?), fluid-escape structures (?)
Dolomite, minor calcite	Dolomite	Dolomite, minor calcite	Dolomite, minor calcite
Microspar, equant spar	Microspar	Microspar	Microspar
Interparticle & Intraparticle	Burrow, evaporite molds	Interparticle	Interparticle, Fenestral (?)
Present	None	None	None

Rock Type 1, Skeletal Grainstone

Grain Type: 75% or more are skeletal: Dasycladacean algae, gastropods, brachiopods, pelecypods, bryozoans, sponges, echinoderm fragments, *Tubiphytes*. Nonskeletal grains are peloids and intraclasts. Most grains are well rounded.

Depositional Textures: Grainstone dominant, local occurrences of packstone interbeds.

Sedimentary structures: Abundant occurrences of low angle, 10 to 20 degrees dip, cross lamination in tabular and wedge shaped sets 10 to 30 cm thick (See Figure 32). Parallel lamination is also abundant. Beds are 3 to 5 m thick.

Rock Fabrics: None

Mineralogy: Predominantly limestone, shelfward patches of dolomite occur.

Diagenetic features:

Cements: Isopachous bladed and fibrous calcite, sparry equant calcite, traces of microspar clacite.

Porosity: Vug, less than 10%.

Internal Sediment: None

Other: Abundant neomorphism of grains into micritic grains.

Other Features: Coatings on many grains in the cross-sets. Coatings are accretionary and/or oncolitic. This rock type is equivalent to the Skeletal-rich limestone facies in Neese and Schwartz (1977).

Figure 32. Outcrop photograph of cross bedding in a nonskeletal grainstone, Rock Type 2. Cross beds have a dip of 10 to 15 degrees in a basinward direction. Outcrop occurs in Section B. Photograph taken by Lloyd Pray.

<u>Rock Type 2, Nonskeletal Grainstones</u>

Grain Type: 75% or more are nonskeletal: peloids predominantly, and intraclasts. Skeletal grains are dasycladacean algae, foraminifera, and gastropods. Most grains are well rounded.

Depositional Textures: Grainstones, local packstone laminae are present.

Sedimentary structures: Abundant occurrences of parallel lamination (See Figure 33), and local occurrences of cross lamination, with 10 to 20 degrees dip in tabular and wedge shaped sets 5 to 10 cm thick. Beds are 1 to 2 m thick.

Rock fabrics: None

Mineralogy: Predominantly dolomite, basinward patches of limestone occur.

Diagenetic features:

Cements: Microspar dolomite, sparry equant calcite, localized patches of isopachous bladed to fibrous dolomite.

Porosity: Vug, less than 10%.

Internal Sediment: None.

Other features: Accretionary coating locally present on some grains. This rock type was not previously described in Neese and Schwartz (1977).

<u>Rock Type 3, Fenestral Nonskeletal-Skeletal Packstone-Grainstone</u>

Grain Type: Skeletal and nonskeletal in about equal proportions. Local occurrences of skeletal-rich beds, especially dasycladacean algae rich beds. Skeletal grains are: dasycladacean algae, gastropods, foraminifera, ostracods, and

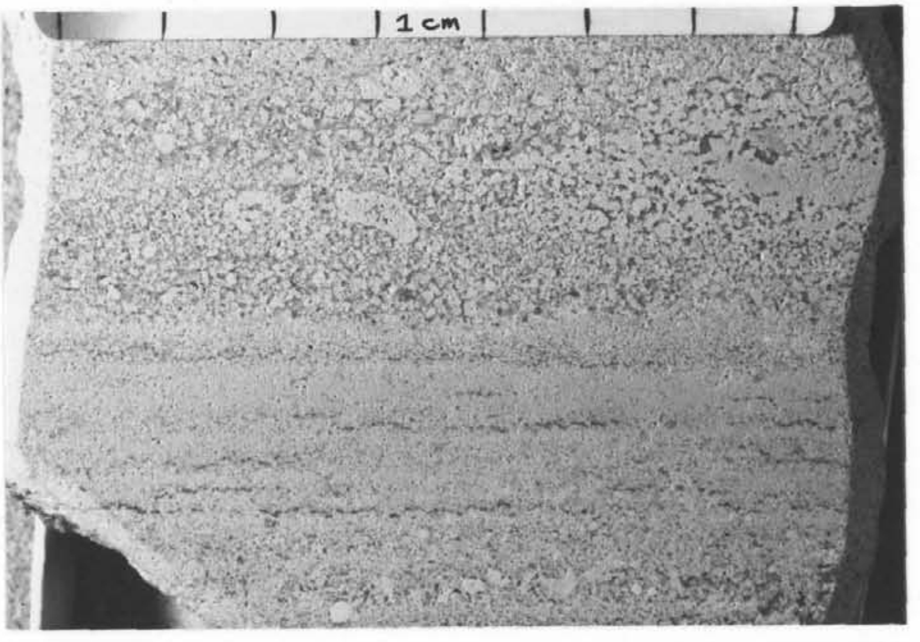

Figure 33. Parallel lamination in a nonskeletal grainstone, Rock Type 2. Sample is from Section B. (UW 1666/16).

thin shelled pelecypods. Nonskeletal grains are: peloids, intraclasts, and pellets.

Depositional Textures: Grainstones and packstones in equal proportions.

Sedimentary structures: None. Beds are 1/2 to 2 m thick.

Rock fabrics: Abundant fenestral fabric, although locally absent. Sheet cracks and tepees are also present. This rock type is present in both the cores of tepees and the intertepee depressions. Local occurrences of burrows.

Mineralogy: Dolomite.

Diagenetic features:

Cements: Microcrystalline dolomite spar, dolomite microspar, sparry equant calcite. Within sheet cracks: radial fibrous calcite and dolomite hemispheroids, mammillary fibrous calcite and dolomite crusts, sparry equant calcite.

Porosity: Fenestral, burrow, intraparticle (within gastropods), vug, approximately 10% total.

Internal Sediment: red iron stained dolomite pellets, dolomite pellet silt.

Other features: None. This rock type was previously described as the Mixed skeletal-nonskeletal dolomite facies in Neese and Schwartz (1977).

Rock Type 4, Fenestral Nonskeletal Packstone-Grainstone

Grain Type: 75% or more are nonskeletal: peloids, intraclast, pellets, and some scattered pisoliths. Skeletal grains are: dasycladacean algae, foraminifera, gastropods, and ostracods.

Depositional Textures: Packstone and grainstone in equal proportions.

Sedimentary structures: None. Beds 1/4 to 1 m thick.

Rock fabrics: Abundant fenestral fabric, sheet cracks, and tepees. This rock type occurs both in the cores of tepees and intertepee depressions. Local occurrences of burrows and algal lamination. Trace amounts of crystallotopic and nodular gypsum and anhydrite molds in the most shelfward portions.

Mineralogy: Dolomite.

Diagenetic features:

Cements: Microcrystalline dolomite spar, dolomite microspar, sparry equant calcite. Within sheet cracks: radial fibrous calcite and dolomite hemispheroids, mammillary fibrous calcite and dolomite crust, sparry equant calcite.

Porosity: Fenestral, burrow, intraparticle (within gastropods), moldic (evaporite), vug, and sheet crack. Total amount of porosity is 10%.

Internal Sediment: red iron stained dolomite pellets, dolomite pellet silt.

Other features: None. This rock type was previously described as the Non-skeletal dolomite in Neese and Schwartz (1977).

<u>Rock Type 5, Pisolitic Packstone-Grainstone</u>

Grain Type: 75% or more are pisoliths. Nonskeletal grains are: pisolites, pellets, peloids, and intraclasts. Skeletal grains are not abundant. Dasycladacean algae are present as nuclei of some pisoliths.

Depositional Textures: Grainstones and packstones. Grainstones are predominant.

Sedimentary structures: Inverse graded bedding is locally present. Beds are 1/4 to 1 m thick.

Rock fabrics: Sheet cracks, some as thick as 10 cm. This rock type is mainly localized in the intertepee depressions. There are local occurrences of polygonal fitting of pisoliths. For a more complete description see the section on pisolite genesis.

Mineralogy: Dolomite.

Diagenetic features:

Cements: Dolomite and calcite microspar, sparry equant calcite. Within sheet cracks: radial fibrous calcite and dolomite hemispheroids, mammillary fibrous calcite and dolomite crusts, laminated calcite and/or dolomite crusts, sparry equant calcite.

Porosity: Local occurrences of interparticle, 10 to 15%.

Internal Sediment: red iron stained dolomite pellets, dolomite pellet silt, microcrystalline dolomite.

Other features: See the section on pisolite genesis for a more complete discussion. This rock type was described previously as the Pisolith-rich dolomite facies in Neese and Schwartz (1977).

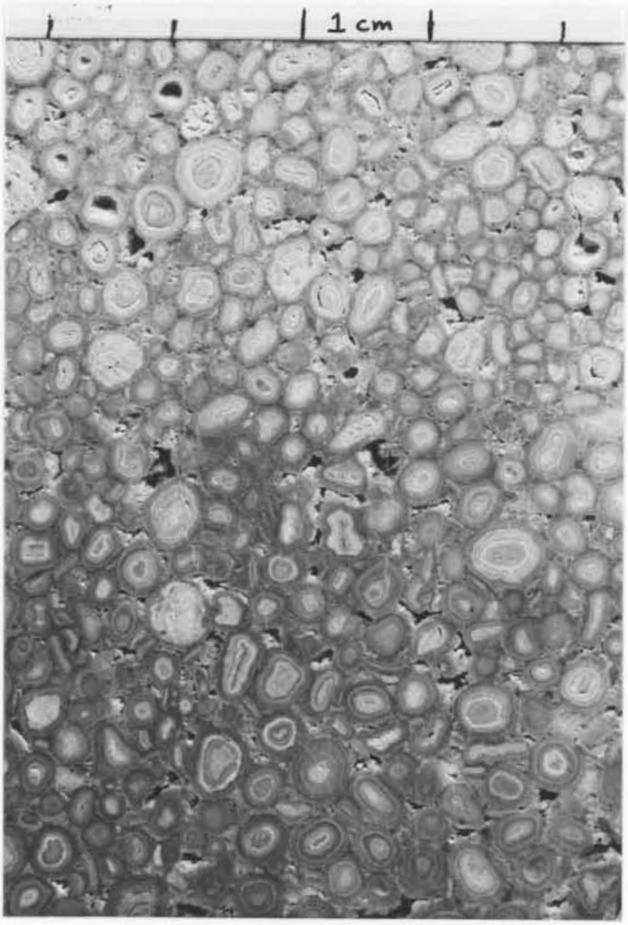

Figure 34. Polished slab of pisolitic grainstone, Rock Type 5. Note the interparticle and intraparticle porosity. (UW 1666/17).

Rock Type 6, Peloid Mudstone-Wackestone

Grain Type: Nonskeletal grains are predominant. 75% or more are peloids, other nonskeletal grains are pellets. Skeletal grains are: calcispheres, encrusting foraminifera, and ostracods.

Depositional Textures: Dominantly wackestone grading into mudstone locally.

Sedimentary structures: Parallel lamination occurs locally. Beds are 5 to 30 cm thick.

Rock fabrics: Local occurrences of burrows, algal lamination, and gypsum and anhydrite molds (both crystallotopic and small, 2 to 5 mm, nodular molds). In the basinward portions there are thin beds with laminar fenestral fabric.

Mineralogy: Dolomite.

Diagenetic features:

Cements: Dolomite microspar.

Porosity: Moldic; crystallotopic and nodular evaporite molds. Local occurrences of burrow porosity. Total amount of porosity is less than 5%.

Internal Sediment: None.

Other features: None. This rock type was previously described as the Mud-supported dolomite facies in Neese and Schwartz (1977).

Rock Type 7, Sandstone (Siliciclastic)

Grain Type: Siliciclastics, silt to very fine sand size.

Depositional Textures: Well sorted grains are subangular. These rocks are quartz arenites using the classification of Dott(1964).

Sedimentary strucutres: Local occurrences of parallel lamination. Also, local occurrences in upper portions of beds of small, 5 to 10 cm thick, sets of current ripple(?) cross lamination with 10 to 20 degrees dip.

Rock fabrics: None.

Mineralogy: Predominantly quartz, trace amounts of potassium feldspar, muscovite, biotite, tourmaline, zircon, and kaolinite.

Diagenetic features:

Cements: Predominantly dolomite microspar with local patches of calcite microspar

Porosity: Interparticle, less than 5%.

Internal Sediment: None.

Other features: Iron staining. Irregular and spherical iron oxide concretions, 1 to 4 cm in size occur (See Figure 35). Also, iron oxide crusts, 1/2 to 1 cm thick, forms on some bedding planes. This rock type was previously described as the Sandstone (siliciclastic) facies in Neese and Schwartz (1977).

<u>Rock Type 8, Siliciclastic-Rich Peloid Wackestone</u>

Grain Type: Silt to very fine sand size siliciclastics, predominantly quartz. Carbonate grains are: peloids, intraclasts,

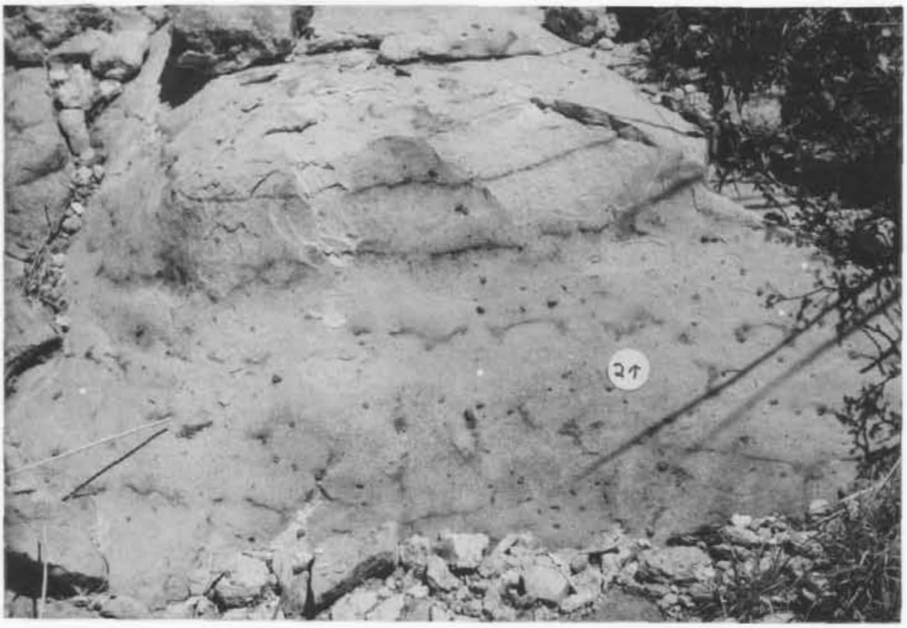

Figure 35. Outcrop of siliciclastic sandstone, Rock Type 7. Note the iron staining and iron concretions. Outcrop occurs in Section E. White circle is 2 cm in diameter.

dasycladacean algae, and a few scattered coated grains and foraminifera.

Depositional Texture: Wackestone.

Sedimentary structures: None. However, there are two different bedding styles: 1) Parallel bedding, 2 to 10 cm thick, parallel laminae. 2) Wavy bedding, with lenticular, wavy, thin (2 to 5 cm thick) beds (See Figure 37). The waves have amplitudes of 5 to 8 cm.

Rock fabrics: In the parallel bedded units there are structures that resemble fluid-escape structures (See Figure 36). They are also similar to desiccation features described by Kinsman and Park (1976) on a carbonate supratidal flat. In the wavy bedded units porosity suggestive of fenestral fabric and nodular anhydrite molds occurs locally.

Mineralogy: Siliciclastics are predominantly quartz, with some kaolinite present. Carbonate is dolomite.

Diagenetic features:

Cements: Dolomite microspar, local patches of calcite microspar.

Porosity: Some porosity suggestive of fenestral, and some suggestive of nodular anhydrite molds. Also, minor amount of interparticle porosity. Total amount of porosity is less than 5%.

Internal Sediment: None.

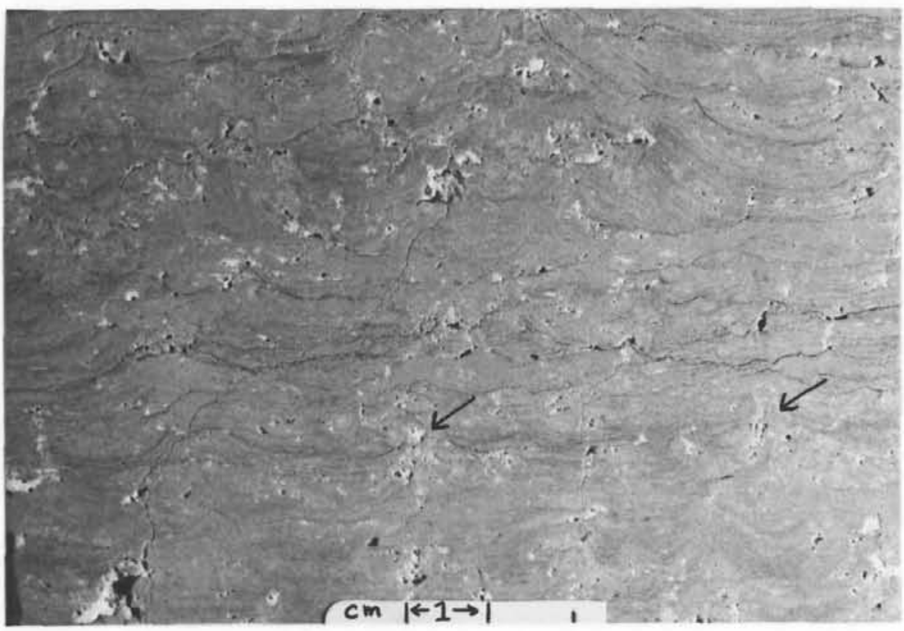

Figure 36. Polished slab of siliciclastic-rich peloid wackestone, Rock Type 8. Arrows indicate possible fluid escape structures. The author is unable to determine if the lamination is of algal mat or current origin. Sample is from Section C. (UW 1666/18).

Figure 37. Outcrop of wavy bedded unit of siliciclastic-rich peloid wackestone, Rock Type 8. Origin of wavy bedding is unknown. Outcrop occurs in Section C. Note the hammer for scale.

Other features: Some of the laminae in the parallel bedded units are iron stained. This rock type was previously described as the Mixed sandstone-dolomite facies in Neese and Schwartz (1977).

PISOLITE GENESIS

The origin of the pisoliths and pisolitic strata has been debated for many years. They have been interpreted as subtidal algal marine nodules, i.e. oncolites, by Ruedemann (1929), Johnson (1942), Adams and Frenzel (1950), Newell et al. (1953), and Kendall (1969). Another interpretation was proposed by Dunham (1965a, b, 1969, 1972) and independently by Thomas (1965). They argued that the pisoliths were not marine algal nodules, but in-place early vadose concretions. Dunham proposed that the pisolitic facies was probably a Permian fossile caliche. The most recent interpretation was proposed by Esteban and Pray (1975, 1976, 1977) and Esteban (1976). They interpreted the pisoliths to be primary particles deposited subaqueously in a peritidal environment.

Dunham (1969) provided excellent criteria and data for demonstrating the in-place non-algal origin of the pisolites. The following features were pointed out by Dunham to support his conclusions: 1) reverse or inverse graded bedding and a paucity of current traction sedimentary structures; 2) polygonal fitting of pisoliths, 3) downward elongations of pisolith laminae, 4) in-place fracturing, 5) perched inclusions incorporated in the upper part of the pisoliths, 6) absence of algal filaments or cells.

Esteban and Pray (1975, 1976, 1977) and Esteban (1976) pointed out that fragmented pisoliths are common as nuclei of other pisoliths. They also noted that the matrix between pisoliths, where present, may contain pisolith fragments. They concluded that these features suggest that some pisolith fragments were present as grains and were redeposited with pisolitic coatings. Kendall (1969) reported normal graded bedding and cross bedding in some of the pisolite beds. He suggested that this indicated that the pisoliths were not formed in-place but were subaqueously transported and then deposited.

Esteban and Pray (1975, 1976, 1977) and Esteban (1976) also pointed out that polygonal fitting, downward elongations, and perched inclusions in the pisolith laminae, occur in only some of the pisoliths and only in the outer laminae of these pisoliths. Esteban (1976) noted that, while inverse grading, polygonal fitting, downward elongation, and perched inclusions indicate an in-place origin, all but downward elongation do not require a vadose environment. Esteban and Pray concluded that in-place growth was evident only in the late stages of pisolith growth. They argued that the data suggested pisolith growth by an accretionary process.

The author examined the major pisolite bearing unit of the upper Yates, (informally the Hairpin dolomite), in Rattlesnake Canyon for features that would aid in determining the origin of the pisolite facies. This unit was chosen because pisolite strata are abundant and widespread (see cross section in appendix). In measured sections C, D, and E inverse graded bedding is present, but is not abundant. Polygonal fitting and downward elongation are present but they also are not abundant. The author observed that fragmented pisoliths are common as nuclei of other pisoliths. Normal graded bedding occurs in a few localities in sections C, D, and E. Cross-bedding was not observed in any locality. From observations of oriented hand samples polygonal fitting and downward elongation, where present, occur in the outer laminae of the pisoliths. However, it is also worth noting, as had been noted by Esteban and Pray (1976, 1977) that elongations in directions other than downward were also observed by the author in Rattlesnake Canyon.

Dunham (1969, 1972) had concluded that the pisoliths were formed in the vadose diagenetic environment in caliche soils. Esteban (1976) noted that the reported features of the pisolite beds in the Permian strata had little resemblance to the macroscopic features of caliche soil profiles. Esteban (1976) described the macroscopic features of caliche;

> "...a vertically zoned, subhorizontal carbonate deposit, normally developed with three main rock types: 1) massive chalky, 2) nodular-crumbly, and 3) laminated and/or pisolitic compact crust or caprock. The position and development of these rock types in the vertical sequence (profile) and laterally is highly variable. The only rather consistent relation is that the massive-chalky rock grades downward into the original rock or sediment through a transition zone, both with strong evidence for in-place alteration and replacement of the original rock or sediment..."

The author did not find any localities in Rattlesnake Canyon where these macroscopic features of caliche occur.

Esteban (1976) also noted that the reported petrographic features of the Capitan pisoliths differ greatly from those of vadose pisoliths in caliche profiles. Vadose caliche pisoliths, where developed, are micritic, indistinctly laminated, and rarely have more than five laminae. Capitan pisoliths have distinct laminae, commonly 10 or more (frequently 20 or more), and are composed of thin micritic layers alternating with mosaics of microspar that have ghosts of radial fibrous or bladed crystals.

A distinctive feature of caliche is the abundance of fine grained calcareous material with a clotted, peloid texture (Esteban, 1976). In the Capitan pisoliths micrite occurs in only some pisolite nuclei and laminae. The inter-pisoliths pores are usually occupied by cements or sand to silt size carbonate matrix material. Dunham (1969) noted the resemblance of the Capitan pisoliths with cave pearls formed in a vadose environment. Esteban (1976) reported, however, that Donahue (1965, 1969) and Ullastre and Masriera (1973) had demonstrated that cave pearls grow subaqueously.

From field studies, Esteban and Pray (1975, 1976, and 1977) reported finding two major, end member types of pisolitic strata, "in-place" and "clastic". The characteristics of in-place pisolitic strata are: 1) well sorted, with inverse grading, 2) little or no matrix, 3) absence of obvious clastic textures, 4) pisoliths are smooth, spheroidal with small nuclei. The characteristics of clastic pisolitic strata are: 1) good to poor sorting, locally showing clastic textures, 2) no inverse grading, 3) abundant calcarenite matrix, 4) broken and abraded pisoliths are present. (See Figures 38 and 39). Observations by the author in the Hairpin Dolomite in Rattlesnake Canyon in sections C, D, and E indicate that the two end member types of pisolitic strata do occur. Clastic pisolite beds are more abundant than in-place pisolitic beds.

Esteban and Pray (1975, 1976, 1977) argued that the in-place

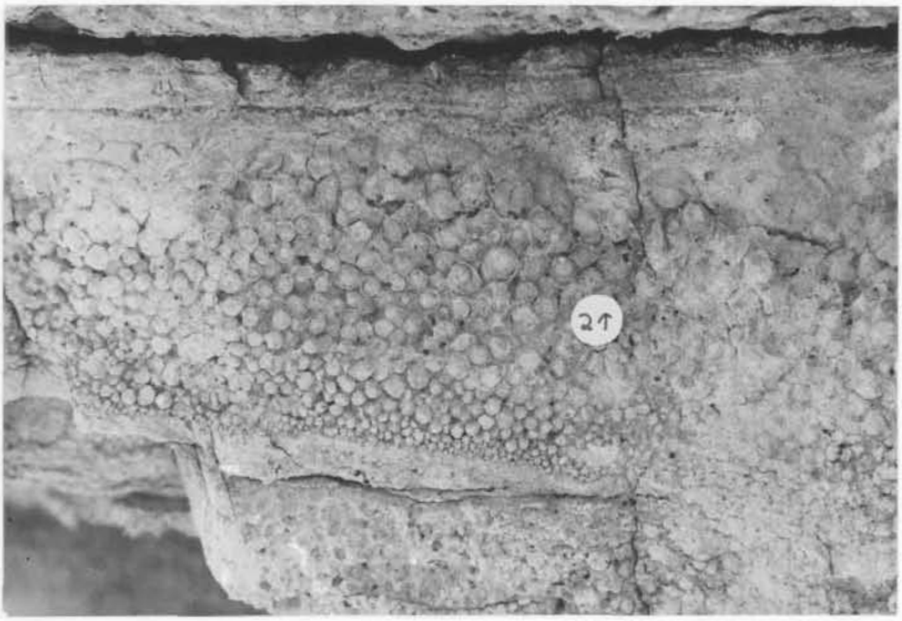

Figure 38. Outcrop of in-place pisolites showing reverse graded bedding. Outcrop occurs in Section E. White circle is 2 cm in diameter.

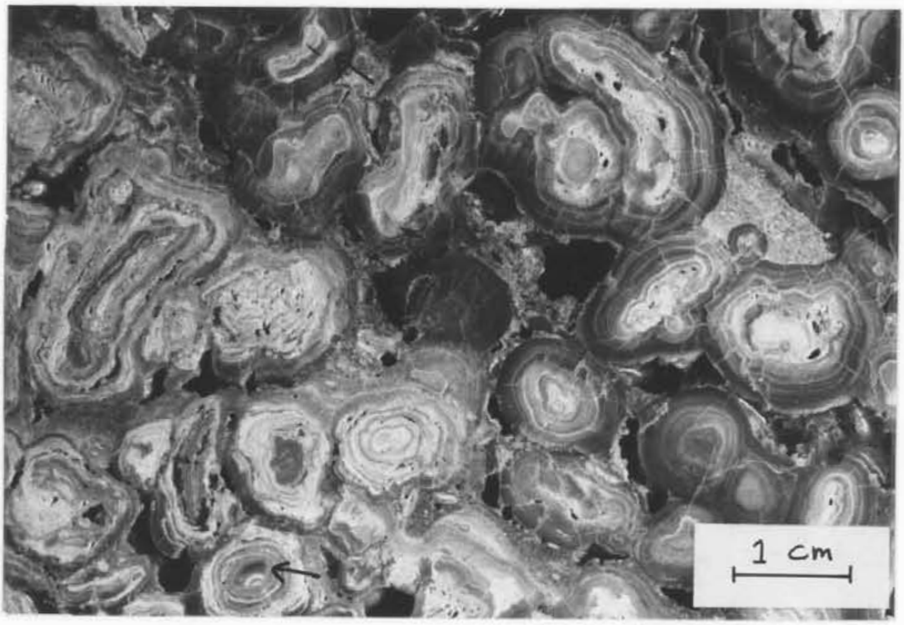

Figure 39. Polished slab of clastic pisolites. Note the irregular shape of these pisoliths and the broken pisoliths that are nuclei. Arrows indicate broken pisoliths that are nuclei. Sample is from Section D. (UW 1666/19).

pisoliths formed mostly by isopachous growth in a bed of loose particles at or near a subaqueous sediment interface. The upper particles in these beds were preferentially enlarged. They noted that the outer laminae of in-place pisolites may show the polygonal fitting, downward elongation, or perched inclusion described by Dunham (1969, 1972). They concluded that these vadose features occurred only locally in the last stage of evolution of the rock. Esteban and Pray argued that the clastic pisoliths probably formed both by accretion on freely moving subaqueous particles and as erosional products of in-place pisoliths. They noted that clastic pisolite beds are more abundant than in-place pisolite beds and many overlie in-place pisolite beds. This vertical relationship of clastic and in-place pisolite beds was also observed by the author in Rattlesnake Canyon.

Esteban and Pray were the first to note that the pisolitic beds were largely localized to the intertepee depressions. Esteban and Pray (1977) documented beds of pisolites which wedged out against the flanks of tepees as evidence of the syndepositional origin of pisolites, and as an argument for a subaqueous origin in possibly restricted water bodies in intertepee ponds. The author observed in numerous localities in Rattlesnake Canyon the localization of pisolite beds to intertepee depressions.

Donahue (1965, 1969) demonstrated that there were four requirements for producing accretionary particles, i.e. ooids, pisolites, and cave pearls. These four requirements are: 1) calcium carbonate deposition from a supersaturated solution, 2) availability of a nucleus, 3) agitation of grains, 4) a splash cup. I suggest that the intertepee depressions would have served as an excellent "megasplash cup" for the formation of the Capitan pisolites.

High on the flanks of some tepees, in-place and clastic pisolite beds are overlain by a rock type, that Esteban and Pray (1975, 1976, 1977) termed botryoidal pisolite. These rocks have both clastic and multilaminar boundstone fabrics, containing large (2 to 10+ cm) irregular pisoliths, many with multi-particle nuclei and coated intraclasts. They suggested that features within these rocks indicated that they may have formed at or near the water surface. Botryoidal pisolites occur in several localities in the Hairpin Dolomite in Sections C, D, E, and F.

Summary

The objective of this thesis was not a detailed study on the origin of the pisolites. Rather, the objective with respect to pisolite genesis, was to establish the facies mosaic of strata in which pisolite beds occur and to attempt to make observations bearing on the interpretations of pisolite genesis. In my study area, in measured sections C, D, E, and F the Hairpin Dolomite provided the best data on the pisolite beds. The author agrees with the interpretation proposed by Esteban and Pray (1977) that postulates a syndepositional primary origin for the pisolites. I interpret the in-place pisoliths to have formed subaqueously by isopachous growth in a bed of loose particles in an interpee depression. The intertepee depression served as a megasplash cup for the formation of the pisolites. Those pisolites near the top were preferentially enlarged because they could be rolled about for a longer duration. Late stages of some in-place pisolite formation show polygonal fitting, downard elongations, and perched inclusions. All of these late stage fabrics resulted only when agitation of the pisoliths ceased. The cessation of agitation could result from a rise in sea level that would place the intertepee depression below wave base. The cessation of agitation could also result if sea level dropped and the intertepee depression became emergent. The presence of downward elongation indicates that in the late stages at least some of the pisolite beds were emergent and in the vadose diagnenetic environment.

Esteban and Pray (1977) postulated that clastic pisoliths were deposited in shallower water than the in-place pisoliths. The broken and fragmented pisoliths indicates increased agitation. The author interprets the clastic pisoliths to have formed by accretion on loose subaqueous particles and as erosional products of in-place pisolites.

Esteban and Pray (1977) proposed that the botryoidal pisolite they described probably formed at or near the water surface. I noted earlier that their botryoidal cement, or my laminated crust cement, greatly resembles supratidal spray zone cements found in the Persian Gulf today. These cements in the Persian Gulf encrust the surface and cavities within beach rock (Purser and Loreau, 1973).

I suggest that the botryoidal cements were possibly the result of spray zone or splash zone precipitation during a time when the crests of the tepees were emergent. The large irregular pisolites and coated intraclasts may have formed in small depressions on the tepee flanks. Purser and Loreau (1973) also described in shallow depressions in the supratidal beach rock zone of the Persian Gulf large (2 to 10+ cm) rock clasts, that are coated with crusts similar to those encrusting the surface and cavities of the nearby beach rock.

DEPOSITIONAL ENVIRONMENTS

Introduction

It is the concensus of geologists that have studied this area that the Guadalupian shelf strata were deposited in shallow marine water. To achieve a better understanding of the Guadalupian shelf strata we should examine the various Holocene shallow water carbonate platforms for possible analogs. Holocene shallow water carbonate platforms are composed of a variety of environments. James (1979) noted that in Holocene sediments there is a common sequence of environments from the reef towards land. This sequence of environments is: 1) reef, 2) lime sand shoals, 3) tidal flats on the lee side of the lime sand shoals, 4) lagoon, 5) tidal flats on the shelfward side of the lagoon, 6) land (See Figure 40). James (1979) also noted that recognition of the tidal flat unit facilitates interpretation of the surrounding lithologies.

Pray (1977) divided the Permian shelf into five regions: 1) shelf edge, 2) outer shelf, 3) shelf crest, 4) inner shelf, 5) evaporite shelf (See Figure 5). Pray's outer shelf, shelf crest, and inner shelf are subdivisions of the marginal mound profile recognized by Dunham (1972). Comparison of Pray's division with James' (1979) division of Holocene shelves produces some interesting analogues. Pray (1977) proposed that the shelf crest was the region of the shelf topographically higher than the rest of the shelf. He proposed that the sediments of the Permian shelf crest were initially deposited in a shallow subtidal environment. As sedimentation pro-

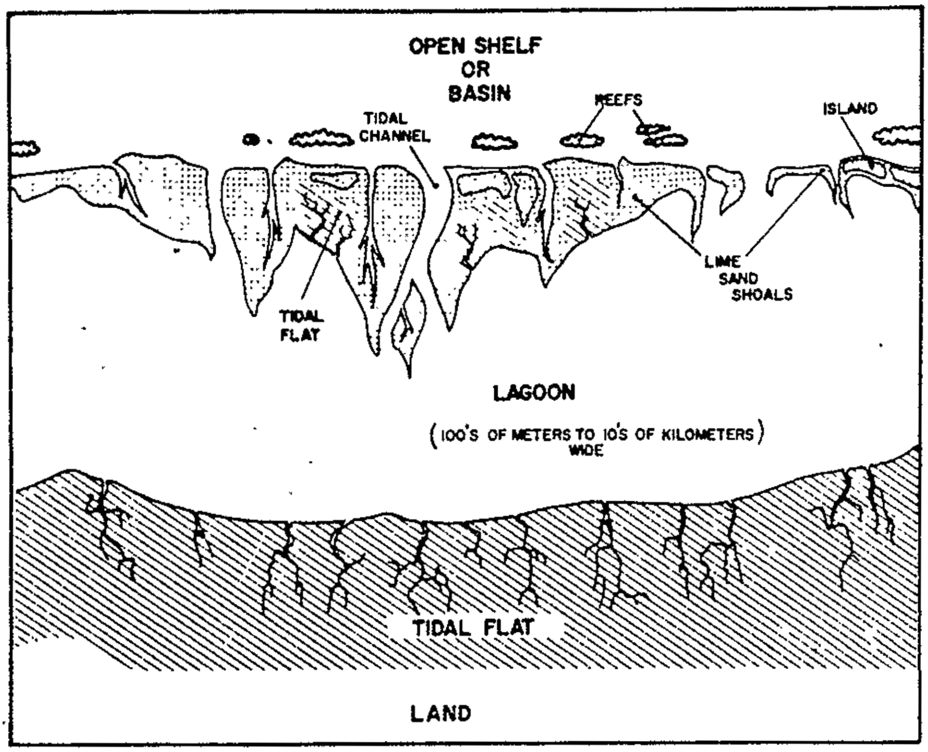

Figure 40. Plan view of a Holocene carbonate platform showing the sequence of environments. Taken from James (1979).

ceeded the shelf crest sediments built up to the peritidal level. Most of the sediments on the shelf crest are peritidal with small interbeds of subtidal sediments. I believe that examination of the profile across Holocene shelves indicates that the tidal flats on the lee side of the carbonate sand shoals most probably corresponds to the shelf crest of the Permian Reef Complex. It then follows that the outer shelf is equivalent to the lime sand shoals, the reef is equivalent to the shelf edge, and the inner shelf is equivalent to the lagoonal sediments.

James (1979) also noted that the transition from the reef across the lime sand shoals, tidal flats, and into the lagoon would be an abrupt succession. To determine if James' (1979) sequence of environments is an adequate model for the Permian shelf strata a comparison was made between features of Holocene environments and features of the Permian shelf.

Outer Shelf or Carbonate Sand Shoals

James (1979) and Wilson (1975) have described the characteristics of carbonate sand shoals. They are generally well sorted, oolitic, pelletoidal or skeletal lime sands. Oncolites may also occur. Bedding is planar with abundant cross lamination. Early cementation is characterestic, therefore intraclasts of cemented lime sand are common. These shoals are mostly deposited subtidally in shallow, 2 to 10 m depth, marine water.

Comparison of these features with features of the study area indicate that the skeletal grainstone (Rock Type 1) and the non-skeletal grainstone (Rock Type 2) present in sections A and B (see cross section in appendix) were probably deposited as subtidal non-emergent lime sand shoals. These rocks are well sorted, grains well rounded, with abundant cross lamination and parallel lamination. Oncolites are also present, but are not abundant. There is no evidence of emergence, i.e. fenestral fabric and mud cracks. Carbonate strata of the outer shelf have basinward dips of 5 to 15 degrees. Neil Hurley (1978,1979) demonstrated, using geopetal fabric measurements, that the primary depositional dip was basinward at approximately 8 degrees. Isopachous bladed or fibrous cement is common in Rock Types 1 and 2. As noted earlier, these cements are probably of submarine origin. Some of the most basinward skeletal grainstones contain fragments of organisms, e.g. Tubiphytes, sponges, bryozoans, and echinoids, that were probably able to live in this environment.

In Holocene carbonate shelves the carbonate sand shoals occur basinward of a tidal flat environment and shelfward of a reef environment. Also the carbonate sand shoals are relatively narrow, approximately 1 km. Examination of the cross section in the study area (see appendix) indicates that outer shelf strata, Rock Types 1 and 2, occur immediately basinward of strata with abundant fenestral fabric. In the field, the author traced outcrops of Rock Type 1 in Section A that rapidly grade basinward into shelf edge rocks of the Capitan Limestone. The lateral relationships of Rock Types 1 and 2 with other rock types also supports the author's conclusion that outcrops of Rock Types 1 and 2, in Sections A and B, were deposited as carbonate sand shoals.

Shelf Crest or Tidal Flat

James (1979) described tidal flats in detail, "...numerous environments may exist in very close proximity, not only perpendicular to the shoreline, but parallel to it as well, so that in the geologic record rapid local lithological variations are to be expected, both vertically and laterally, rather than a smooth succession of progressively shallower environments". Sediments of the tidal flats are generally recognized by the presence of fenestral fabric. In this discussion the author will consider the shelf crest to include rocks deposited predominantly in a peritidal environment.

I interpret the fenestral nonskeletal-skeletal packstone-grainstones (Rock Type 3), the fenestral nonskeletal packstone-grainstones (Rock Type 4), and the pisolitic grainstone-packstones (Rock Type 5) to have been deposited in a peritidal environment. The abundance of fenestral fabric and sheet cracks within and associated with these rocks indicates their peritidal nature. In an earlier section the author documented evidence that supports a peritidal origin for the Guadalupian tepees. Tepees are also abundant in strata containing Rock Types 3, 4, and 5 (See Hairpin Dolomite, Sections C, D, E, and F).

A comparison of the lateral relationships of Permian shelf crest strata with the lateral relationships in Holocene shelves suggests that the Permian shelf crest strata were deposited predominantly as a tidal flat in the lee of carbonate sand shoals. Basinward, Rock Types 3, 4, and 5 grade into outer shelf strata containing Rock Types 1 and 2. This transition was observed by the author in the Triplet Dolomite, and the Basal Dolomite between Sections B and A. Shelfward Rock Types 3, 4, and 5 grade into strata containing Rock Type 6. This transition was observed in the Triplet Dolomite between Sections E and F. The lateral relationship of Rock Types 3, 4, and 5 with outer shelf strata is analogous to the relationship of tidal flats on the lee side of shoals to the shoals themselves.

As noted earlier, in the geologic record the tidal flat environment should show local lithologic variation both vertically and laterally. In all three carbonate units in Sections C, D, and E local lithologic variation both vertically and laterally is evident. The local lithologic variation was greatly assisted by the formation of tepees. It is evident that the formation of tepees many produce a variety of mini-environments. Within one intertepee depression in-place pisoliths could be forming. At the same time in another depression clastic pisoliths could be forming. In another depression fenestral skeletal or mixed nonskeletal-skeletal strata could be forming. The author suggests that at any one particular point in time tepees could be developing at one part of the shelf crest, fenestral laminated rocks in another part, and pisolites in another part of the shelf crest. All of these localities could be along facies strike.

The author envisions the following sequence of events in the formation of these tidal flats. The first event would be deposition of a nonskeletal or mixed skeletal-nonskeletal sediment on the lee side of the carbonate sand shoals. James (1979) noted that the tidal flats form on the lee side of the carbonate sand shoals after beach ridges developed or currents had swept sand together to form island. As sedimentation proceeded the nonskeletal sediment and/or mixed skeletal-nonskeletal sediment would enter the intertidal environment and strata with fenestral fabric and local algal lamination would develop.

After the development of fenestral fabric, the formation of sheet cracks and tepees would be initiated, possibly by further desiccation. Precipitation of cements would then produce expansion of sheet cracks necessary to form mature tepees. As noted earlier the evidence supports a submarine origin for radial fibrous hemispheroid cements. The author suggests that if there was a small rise in sea level at this time the large pores in the cores of tepees would be ideal sites for the precipitation of cements resembling those present in the large cavities of Holocene reefs. As noted earlier the radial fibrous hemispheroid cements in sheet cracks greatly resemble radial fibrous hemispheroids of aragonite found within cavities in the Holocene deep reef-wall limestones in Belize, British Honduras (Ginsburg and James, 1976). Dunham (1972) noted the similarity of this type of cement to cements forming in caverns. He proposed a vadose origin for this type of cement. However, this cement has been found interlaminated with internal sediment containing marine biota Babock (1974), Yurewicz (1976), and Babcock, Pray, and Yurewicz (1977). This evidence indicates that the radial fibrous hemispheroids were probably submarine cements.

This proposed rise in sea level would also be an appropriate time for the initiation of in-place pisolith development. A drop in sea level could then produce the formation of clastic pisolith fabrics, and also local truncation of tepees. Botryoidal pisolites would possibly form at a time when the crests of tepees were at or above the water surface.

Esteban and Pray (1977) proposed the following sequence of events. A rise in sea level resulting from shelf subsidence would initiate development of in-place pisolites in intertepee depressions. As sedimentation proceeded water depth would become shallower and clastic pisolith development would be initiated. At the same time that these events were occurring botryoidal pisolites could be forming on the flanks of those tepees at or near the water surface. Esteban and Pray (1977) call for a single rise in sea level and sedimentation accreting towards the water surface (fill level) producing shoaler conditions.

The effectiveness of a change in sea level as a result of shelf subsidence in producing a change from one sediment type to another is dependent on the amount of sediment within an intertepee depression. For example, if one depression is filled to a level closer to the water surface than in another depression, a drop in sea level would probably initiate clastic pisolite formation in the higher one and might not initiate it in the one that is only slightly filled.

<u>Inner Shelf or Lagoon</u>

Wilson (1975) has described the characteristics of sediments of restricted marine shelf lagoons. They are characteristically mudstones or wackestones with peloids and/or pellets as the dominant grain type. These mudstones may be laminated, burrowed, or algal laminated. Locally, unfossiliferous unlaminated homogeneous micrite with crystallotopic or scattered nodular molds of evaporites, gypsum and anhydrite, also occur.

I interpret the peloid dolomite mudstone-wackestone (Rock Type 6) to have been deposited in a shallow subtidal lagoon. The author observed in Section F this rock type with local occurrences of evaporite molds, burrows, and algal laminations. Evidence of subaerial exposure is not present (mud cracks, sheet cracks, peritidal tepees, and fenestral fabric. The low abundance of biota, and the low diversity, plus the presence of evaporite molds indicates that the lagoon was probably restricted, producing hypersaline conditions.

James (1979) pointed out that on Holocene carbonate shelves lagoonal sediments most commonly occur shelfward of the tidal flats on the lee side of the carbonate sand shoals. Examination of the cross section of the study area shows that Rock Type 6 grades basinward into Rock Types 3, 4, and 5. The author observed this transition between Sections E and F. As described in the preceeding section the author interprets Rock Types 3, 4, and 5 to represent a peritidal flat that formed on the lee side of carbonate sand shoals. The lateral relationships observed in the Permian shelf strata is analogous to relationships observed on Holocene carbonate shelves. This further supports the conclusion that Rock Type 6 was deposited as a lagoonal sediment shelfward of tidal flats on the lee side of carbonate sand shoals.

<u>Summary of Carbonate Depositional Environments</u>

The profile across the Guadalupian shelf during carbonate deposition consisted of a shelf edge, an outer shelf, a shelf crest, an inner shelf, and an evaporite shelf. I interpret my study area to have been on the outer shelf, shelf crest, and the basinward portion of the inner shelf.

The rocks of the shelf edge form the Capitan Massive Limestone. The outer shelf rocks, largely skeletal grainstones (Rock Type 1) and nonskeletal grainstones (Rock Type 2) formed as carbonate nonemergent sand shoals deposited in a subtidal environment. The shelf crest rocks, fenestral nonskeletal-skeletal packstone-grainstones (Rock Type 3), fenestral nonskeletal packstone-grainstones (Rock Type 4), and pisolitic packstone-grainstone (Rock Type 5), were the result of peritidal depositon on a "tidal flat" on the lee side of the carbonate sand shoals. Rock Types 3 and 4 formed in the intertidal and supratidal portions of the "tidal flat". Rock Type 5 predominantly formed in the infratidal portions of the "tidal flat". The inner shelf rocks, peloid dolomite mudstone-wackestones (Rock Type 6) were deposited in a restricted probably hypersaline shallow subtidal lagoon. Comparison of the fabrics within these rocks and the lateral relationships of these rocks to fabrics and lateral relationships on Holocene carbonate shelves supports these interpretations.

Recently, Babcock (1974), Yurewicz (1976, 1977) Babcock *et al.* (1977), Cys *et al.* (1977), and Hurley (1978, 1979) noted a major difference between Holocene carbonate shelves and the Permian shelf in that the Capitan Limestone was deposited in water below wave base, 20 to 50 m in depth, whereas Holocene shelf edge sediments (reefs) are deposited in shallow water above wave base. These authors cited the following as evidence of this difference: 1) lack of major biotic zonation and appreciable skeletal fragmentation within Capitan Limestone, 2) lack of Capitan-derived sediment or erosional blocks of the Capitan within basinward shelf formations, 3) extrapolation of primary depositional basinward dip from the shelf crest to the Capitan Limestone.

Sandstone Depositional Environment

Determination of the environment of deposition of the siliciclastic sandstone units is not an easy task. Outcrops of the sandstones are not abundant, and sedimentary structures are not easily seen in outcrops or rock samples. Sedimentary structures are not easily recognized due to the excellent sorting and the lack of clays. Where recognizable, sedimentary structures are small current ripples, cross laminations, and parallel lamination. These features were observed in the upper portions of beds in measured Section A and in outcrop in Walnut Canyon.

Sandstones in the study area occur as two sheets 5 and 8 m thick. These two sandstone sheets are laterally persistent along facies strike over the entire shelf (Neese and Schwartz, 1977). The basal contact of these sandstone units with the underlying carbonate unit is sharp and erosional only in the area of the carbonate shelf crest. In the outer shelf these sandstone units thin and break up into smaller beds that are interbedded with the carbonate sand shoal beds. In the study area this transition occurs between Sections C and A (see cross section in appendix). Further basinward the sands become calcareous and grade into the shelfward portions of the shelf edge rocks.

The lateral continuity of the sandstone beds, and the uniformity of the siliciclastic grains suggests that the siliciclastic grains may have been transported to the carbonate shelf by eolian processes. Once they reached the shelf the question remains: were they deposited by eolian processes or were they deposited in a lagoon? Adams and Frenzel (1950), and Kendall (1969) supported an eolian depositional environment. Newell et al. (1953), Ball et al. (1971), Jacka et al. (1969), and Smith (1974b) supported a subaqueous depositional environment interpretation. Smith (1974b) noted that adhesion ripples, typical of eolian sand sheets, are absent, and fine-grained detrital white micas, rare in eolian environments, are widespread and locally abundant. Two subaqueous depositional environments have been proposed. Ball et al. (1971) proposed a lagoonal environment. Silver and Todd (1969) proposed a coastal plain environment.

Pray (1977) and Hurley (1978) have suggested that the siliciclastics were deposited at a time when the shelf subsided, producing a higher sea level. I suggest that the shelf subsidence may have increased the slope of the shelf. This increase in slope may have provided a setting in which transport of siliciclastics in a shallow subaqueous density current may have occurred. Blatt et al. (1972) noted that sandstone beds lacking any lamination,

"...are formed either by very rapid deposition from suspension or by deposition from very highly concentrated sediment dispersions". A grain flow is a type of sediment dispersion. Some of the characteristics of grain flow deposits are: 1) thick, ungraded, sharply bounded beds, generally massive internally; 2) presence of faint, dish-shaped laminae, oriented parallel with bedding (can be interpreted as fluid escape structures); 3) scarcity of sole marks; 4) lack of traction structures (Stauffer, 1967).

Deposition from a very highly concentrated sediment disperion (grain flow) could explain the absence of lamination and the presence of fluid escape structures in Guadalupian age siliciclastics. The traction structures present in the sandstone beds occur only in the upper portions of the sandstone units. These structures probably formed as sea level was falling, just before the recurrence of carbonate deposition on the shelf.

Hurley (1978) reported finding quartz siltstone beds which pinched out on the shelfward side of a presumed shelf crest. The thickness of these pinched out beds (3 m) and an estimated relief across the shelf crest of 3 m, suggests a water depth of 3 to 6 m for sandstone deposition. Therefore, I suggest that the siliciclastics were transported in highly concentrated sediment dispersions from which they were deposited in water depths of 3 to 6 m that resulted from subsidence of the shelf.

The thinning and grading of the sandstones into the basinward portion of the outershelf, indicates that sandstone deposition was probably contemporaneous with deposition of the outer shelf and shelf edge. I suggest that the carbonate sand shoals of the outer shelf may have served as barriers to deposition of sand beds on the shelf edge.

The siliciclastic rich peloid wackestone (Rock Type 8) represents short intervals of time when neither sandstone nor carbonate environments were dominant. Smith (1974b) noted the presence of sandstone dikes in some shelf edge rocks, suggesting that lithification of the sandstones was later than that of the carbonates. This difference in the time of cementation may help explain the wavy bedding found in the siliciclastic rich peloid wackestone (Rock Type 8). Wavy bedding could have evolved from compaction of a heterogeneous layer composed of lithified carbonate lenses surrounded by "soft" siliciclastic rock.

CYCLIC SEDIMENTATION

Introduction

Cyclic strata are present in most of the carbonate shelves that have been studied by geologists. Shelf strata of the Permian Reef Complex are no exception. Two forms of cyclic strata are present in the rocks of the study area. First, at a large scale, there is interbedding of sandstones and predominant carbonate units. Second, within the carbonate strata there are cycles of different carbonate rock types. Both of these types of cyclic strata are discussed below.

Sandstone-Carbonate Cycles

The sandstones on the shelf have received much attention but there is little agreement as to their depositional environment. Some of the characteristics of these sandstones can be explained by cyclic sedimentation.

Each cycle consists of a sandstone unit overlain by a carbonate unit. The basal contacts I observed of sandstone units with underlying carbonate units are sharp and erosional. In the Triplet Unit a weathered zone, described previously, formed at the top of the carbonate in the area of the shelf crest (see Sections C, D, and E). The contact of the sandstone with the overlying carbonate unit is in other places gradational over approximately 1/2 m.

Silver and Todd (1969) and Meissner (1972) proposed that the carbonate units were deposited at stages of high sea level, clastic units at stages of low sea level. Subaqueous and/or subaerial erosion of the carbonates occurred during short intervals of extremely low sea level. Pray (1977) suggested exactly the opposite, carbonate units were deposited at stages of low sea level, clastic units at stages of high sea level. He postulated that subsidence would allow saline lagoon waters to spill onto the shelf and inhibit carbonate production and allow shallow currents to transport sands onto the shelf area. My interpretation of sandstone deposition in highly concentrated sediment dispersions (see section on sandstone depositional environment) supports Pray's interpretation.

Silver and Todd(1969) and Meissner (1972) suggested that the sea level change was a result of eustasy or epeirogeny. Orogenic movements would not be able to produce the uniform cycles of large areal extent that are present. Pray (1977) suggested that episodic subsidence could have produced the sandstone-carbonate cycles.

Carbonate Cycles

Cycles within the carbonate strata were previously described by Smith (1974b) and Esteban and Pray (1977). Cyclic sedimentation is best developed in the strata of the shelf crest.

Cycles found by the author consist of the pisolitic packstone-grainstones (Rock Type 5) alternating with the fenestral nonskeletal packstone-grainstones (Rock Type 4) or pisolitic packstone-grainstones (Rock Type 5) alternating with the fenestral nonskeletal-skeletal packstone-grainstones (Rock Type 3). These cycles were observed in the Hairpin Dolomite in Sections C, D, E, and F; in the Triplet

Dolomite in Sections D and E; in the Basal Dolomite in Sections B, C, D, and E. These cycles are best seen in the intertepee depressions and cannot be correlated from one depression to another. These cycles are 1/2 to 1 m in thickness. The contact between the pisolitic unit and the overlying unit is commonly gradational over 3 to 5 cm, although locally it can be a sharp contact. The contact between the pisolitic unit and the underlying fenestral unit is sharp and probably erosional. If we accept Esteban and Pray's (1975, 1976, 1977) interpretation that the pisolites originate subaqueously, then the sequence from pisolitic packstones-grainstones into overlying fenestral packstones-grainstones is a shoaling upward sequence as they suggested. The nature of contacts I have observed between the units of cycles also supports this conclusion.

Esteban and Pray (1977) described cycles within the pisolitic rocks which they named the "walnutite cycle". The basal member of the walnutite cycle is a fenestral peloid packstone bed. This is overlain by a bed of in-place pisolites. The contact between the fenestral unit and the overlying in-place pisolitic unit is sharp and locally erosional. Overlying the in-place pisolite bed is a clastic pisolite bed. The contact between the in-place pisolites and the clastic pisolite beds is gradational over 2 to 5 cm. The clastic pisolite bed grades upward, over 4 to 6 cm, into the basal member of the next cycle. This type of cycle was observed in the Hairpin Dolomite in Sections C, D, E, and F; in the Triplet Dolomite in Sections D and E; in the Basal Dolomite in Sections B, C, D, and E.

The "walnutite cycle" is best developed in the intertepee depressions. On the flanks of tepees in-place pisolite units are poorly developed relative to clastic pisolite units. The "walnutite cycle" was interpreted by Esteban and Pray (1977) as a shoaling upward cycle. The fenestral peloid packstones occured first and these small grains may have provided the nuclei for the overlying in-place pisolites. Also, coincident with the formation of fenestral fabric, tepees would develop that form the "mega-splash" cups necessary for formation of in-place pisolites. The clastic pisolites would then form after the in-place pisolites.

Shelfward from the pisolite belt cycles within the peloid dolomite mudstone-wackestones (Rock Type 6) were observed in the Basal Dolomite in Section F. These cycles consists of beds of algal mat(?) laminated mudstones with local lenses of evaporite molds and burrows alternating with thin beds of laminoid fenestral wackestones. These cycles are not well developed and occur over an interval of 1/2 to 1 m. The two members of each cycle grade into each other over a vertical thickness of 5 to 10 cm. These cycles may represent a shallowing upward cycle from a shallow subtidal lagoon into a low intertidal environment. This type of cycle may have resulted from shifting of the shelf crest environment and the inner shelf environment.

In my study area another style of cyclic carbonate strata occurs. These cycles occur in an approximately 50 m wide belt, 150 to 200 m shelfward from the Capitan Limestone. This cycle was observed in the Basal Dolomite in Section A. This cycle has 2 members, the basal member of the cycle is either skeletal grainstone (Rock Type 1) or the nonskeletal grainstone (Rock Type 2). The basal member (1/2 to 1 m thick) usually contains traction structures, predominantly parallel lamination although cross lamination occurs locally. This member grades upward, over 1 to 5 cm, into the upper member, a fenestral nonskeletal-skeletal packstone-grainstone (Rock Type 3) unit, 1 to 2 m thick. The contact between the fenestral unit and the overlying basal member of the next cycle is generally sharp. It should be noted that these cycles are not common and may occur only once or twice in 50 m of vertical stratigraphic section. The author interprets this to be a shoaling upward cycle from a subaqueous environment into an intertidal environment. This cycle may have resulted from shifting of the outer shelf environment and the shelf crest environment.

Summary

Cyclic sedimentation on the shelf crest resulted from changes in sea level or lateral shifts of mini-environments along facies strike or episodic subsidence or a combination of these factors. The peritidal environment in which the pisolites and tepees formed can be composed of many subenvironments. I suggest that the lateral shifts of environment were possibly initiated by storms. Storms occur on most Holocene carbonate platforms and may be largely responsible for the transport of sediment on the tidal flat (Hardie and Garret, 1977). It has also been demonstrated that some of the morphology of Holocene tidal flats is determined by the occasional storm and not by the daily tides.

The carbonate cycles on the outer and inner shelf could also be the result of lateral shifts of environments along facies strike. However, these cycles and the cycles in the pisolite belt could arise from changes in sea level.

It is worth noting that in strata of my study area cycles consisting of outer shelf rocks overlain by shelf crest rocks in turn overlain by inner shelf rocks are absent. The absence of this type of cylic sedimentation indicates that changes in sea level were not large enough to cause large shifts in the environment. The author suggests that changes in sea level were only on the order of 2 m or less. Cycles are present at the transitions between environments; i.e. shelf crest to inner shelf, and shelf crest to outer shelf. Cycles are also present within the shelf crest. Smith (1974b) and Esteban and Pray (1977) suggested that these cycles are better explained by small 1/2 to 2 m changes in sea level. The author agrees with this suggestion.

The sandstone carbonate cycles are due to larger changes in sea level. Yurewicz (1976, 1977) suggests that the shelf edge sediments were deposited in depths greater than 15 or 20 m. Basinward the sandstones pinch out and grade into the shelf edge Capitan Massive Limestone. Also, erosion surfaces at the base of sandstone units are localized to the shelf crest, and are not present in inner shelf or outer shelf strata. This indicates that the rise in sea level resulting from shelf subsidence that initiated sandstone deposition did not expose sediments of the shelf edge. The author suggests that sandstone-carbonate cycles were the result of sea level fluctuation on the order of 5 to 10 m.

VARIATION OF THE SHELF CREST WITH TIME

Introduction

Cyclic sedimentation is one type of variability of the shelf environments with time. However, examination of the cross section of the study area (see appendix) indicates that the shelf was quite different during deposition of each of the three informal units, The Hairpin Dolomite, the Triplet Dolomite, and the Basal Dolomite. The three informal units differ from each other with respect to the abundance of different fabrics of the shelf crest and the width of the shelf crest. From observations in the field the author noted that the width of the outer shelf strata does not seem to vary with time. The shelf crest strata appear to expand in width only in a shelfward direction. This type of variation of the shelf strata was first reported by the author and Douglas Neese in Neese and Schwartz (1977).

Hairpin Dolomite

The Hairpin Dolomite is characterized by an abundance, both vertically and laterally, of tepees and pisolitic rocks. Tepees and pisolites, in association with fenestral packstones and grainstones occur from 1/4 to 2 1/4 km in back of the shelf edge (see Sections B, C, D, E, and F). The author was not able to trace this unit far enough shelfward to determine the location where the pisolites and tepees are no longer present because the rocks intersected with the summit peneplain surface (King, 1948) forming the top of the Guadalupe Mountains. It is evident that the peritidal shelf crest environment was well developed and wide during deposition of the Hairpin Dolomite. The lateral transition from the shelf crest strata to rocks of the shelf edge occurs between Section B and A in the Hairpin Dolomite. This transition is rapid, occurring over a distance less than 1/4 km. It is suggested that subsidence during deposition of the Hairpin Dolomite was slow. The slow subsidence enabled the peritidal shelf crest to become extremely wide. The slow subsidence would also allow for long periods of time during which peritidal diagenetic processes could produce the abundant tepees and sheet cracks.

Triplet Dolomite

The Triplet Dolomite is characterized by a paucity of tepees and pisolites (see cross section in appendix). However, fenestral fabric and thin sheet cracks are abundant. Examination of stratigraphic Sections C, D, and E (see appendix) indicates that tepees in the Triplet Dolomite appear higher in the section in a shelfward direction. Also, in Section B a 2 m thick bed of outer shelf non-skeletal grainstones (Rock Type 2) is present at the top of the Triplet Dolomite. This indicates a shelfward transition of the outer shelf and shelf crest environments during deposition of the Triplet Dolomite.

I suggest that subsidence during deposition of the Triplet Dolomite was rapid and that carbonate sedimentation was barely able to keep up with it. Examination of the cross section indicates that shelf crest strata, Rock Types 3, 4, and 5 in Sections C, D, and E occurred in a belt approximately 1 1/2 km wide. The paucity of tepees is possibly due to the suggested rapid subsidence. Rapid subsidence could shorten considerably the duration sediment spent in the peritidal diagenetic environment.

Basal Dolomite

In comparison with the Hairpin Dolomite and the Triplet Dolomite, the Basal Dolomite has characteristics that are somewhere in between. Tepees and pisolites are present and abundant (see Sections B, C, D, and E) but they are not as abundant and widespread as in the Hairpin Dolomite. In previous work, (Neese and Schwartz, 1977) it was noted that skeletal grains are more common in the Basal Dolomite, than in either the Triplet Dolomite or the Hairpin Dolomite. Examination of the cross section (see appendix), show that the distribution of shelf crest strata, Rock Types 3, 4, and 5, indicates that the shelf crest for the Basal Dolomite was wider than that of the Triplet Dolomite but narrower than that of the Hairpin Dolomite.

I suggest that the subsidence rate during deposition of the Basal Dolomite was slow enough to allow abundant formation of tepees and pisolitic strata. However, the subsidence rate was not so slow as to allow the formation of tepees to become extremely widespread. It is also worth noting that tepees in the Basal Dolomite are not as large nor do they have as much relief as those in the Hairpin Dolomite. The large number and type of skeletal fragments in this unit indicate that normal marine conditions were more prevalent rather than hypersaline conditions. I suggest that during deposition of this unit sea level was slightly higher and allowed marine water to circulate onto the shelf crest "tidal flat".

Summary

The Hairpin Dolomite, Triplet Dolomite, and Basal Dolomite differ from each other with respect to the abundance of different fabrics of the shelf crest and the width of the shelf crest. The author has suggested that these differences are possibly related to different rates of subsidence.

A very slow rate of subsidence would enable a large amount of sediment to accrete on the peritidal shelf crest. A very slow rate of subsidence would also produce long intervals of time during which peritidal diagenetic processes, tepee and sheet crack formation, could occur and become abundant. A fast rate would shorten significantly the intervals of time that sediment would be affected by peritidal diagenetic processes.

There is another possible explanation. Examination of the cross section (see appendix) shows that sandstone units were deposited between deposition of each of these three carbonate units. The affects of the sandstone deposition on the morphology of the shelf could have produced the differences between these three units. It has been demonstrated that differences in relief on the order of 1/2 to 1 m can produce significant changes in depositional environments.

CONCLUSIONS

1. The carbonate shelf strata were deposited in bands parallel to the shelf edge on a marginal mound profile (Dunham, 1972) which can be subdivided into five facies. In a shelfward direction these facies are: 1) shelf edge, 2) outer shelf, 3) shelf crest, 4) inner shelf, and 5) evaporite shelf (Pray, 1977). Rocks of my study area were deposited in the outer shelf, shelf crest, and inner shelf. The Capitan Limestone was deposited on the shelf edge.
2. Eight major rock types of the shelf strata of my study area are: 1) skeletal grainstone, 2) nonskeletal grainstone, 3) fenestral nonskeletal-skeletal packstone-grainstone, 4) fenestral nonskeletal packstone-grainstone, 5) pisolitic packstone-grainstone, 6) peloid mudstone-wackestone, 7) siliciclastic sandstone, and 8) siliciclastic-rich peloid wackestone.

   The outer shelf strata are composed of Rock Types 1 and 2. The shelf crest strata are composed of Rock Types 3, 4, and 5. The inner shelf strata are composed of Rock Type 6. The author interprets the outer shelf strata to have been deposited in a subtidal non-emergent carbonate sand shoal environment. The shelf crest strata were the result of peritidal deposition on a "tidal flat" on the lee side of the carbonate sand shoals. The inner shelf strata were deposited in a restricted, probably hypersaline, shallow lagoon. This sequence is analogous to that proposed by James (1979) for Holocene carbonate shelves.

3. Carbonate rocks of the inner shelf, the shelf crest, and the most shelfward outer shelf are dolomites. These rocks grade into limestones of the outer shelf and the shelf edge. Dolomitization is interpreted to have been eogenetic.
4. The presence of evaporites in the study area is indicated by crystallotopic and nodular gypsum and anhydrite molds. The author interprets the evaporites to have been precipitated within the sediments from hypersaline brines forming in a restricted lagoon, probably within shallow subtidal, intertidal, and low supratidal sediments. Bedded evaporites are not present.
5. The two sandstone units in the Triplet unit are continuous over most of the study area to within 200 m of the Capitan Limestone, where they thin and grade abruptly into the shelf edge carbonates. The author suggests that the siliciclastics were transported in highly concentrated sediment dispersions from which they were deposited in water depths of 3 to 6 m.
6. The core of most tepees is composed of fenestral laminated rocks (Rock Types 3 and 4) and sheet cracks filled with cement and internal sediment. Most tepees, and all large tepees occur only in shelf crest strata. Sediments in intertepee depressions generally wedge out against the flanks of tepees. This feature and erosional truncation of tepees indicates that tepee formation was syndepositional. Tepee formation may be the result of a two phase process, an early phase of desiccation contraction and a later phase of expansion from crystallization of carbonate cements, although the mechanics of each process are not completely

understood.

7. The author's observations supports Esteban and Pray's (1975, 1976, 1977) interpretation of pisolite genesis. Pisoliths are interpreted to be primary particles deposited subaqueously in a peritidal environment. Two end member types of pisolitic strata were observed, "in-place" and "clastic". Clastic pisolitic strata are more abundant than in-place strata and commonly overlie in-place pisolitic strata. Pisolitic beds are largely localized in the intertepee depressions. The intertepee depressions served as "mega-splash" cups for pisolite formation.

   Botryoidal pisolites formed on the flanks of emergent tepees. The botryoidal coating cements were probably precipitated in a supratidal spray zone.

8. Carbonate grain size and skeletal grain abundance decreases in a shelfward direction. Nonskeletal carbonate grains are the most abundant grains in the shelf strata.

9. Eight types of porosity are present in the carbonate rocks. They are: 1) fenestral, 2) interparticle, 3) moldic, 4) burrow, 5) intraparticle, 6) vug, 7) sheet crack, and 8) fracture. The fenestral porosity and interparticle porosity are present in significant amounts and are probably the only types of pores to be of economic significance in petroleum reservoir rocks.

10. Four cement types occur within former pores (other than sheet cracks). These four types are: 1) isopachous bladed to fibrous, 2) microcrystalline, 3) microspar, and 4) equant spar. Each of these cements probably formed in more than one environment. The

author interprets the ispachous, microcrystalline, and microspar cements to have been primary and eogenetic, probably precipitated out of marine to hypersaline water. The equant spar cement is later and interpreted as a fresh water cement.

11. Four cement types previously recognized by Dunham (1972), occur within the large sheet cracks in tepee cores. These four types are: 1) radial fibrous hemispheroids, 2) mammillary fibrous crusts, 3) laminated crusts, and 4) sparry equant calcite. The author interprets the radial fibrous hemispheroids and the mammillary fibrous crusts to have been precipitated early in a subaqueous marine to hypersaline environment. The laminated crusts are interpreted to have been precipitated in a supratidal splash zone environment. The sparry equant calcite was precipitated late in a fresh water phreatic environment.

12. Internal sediment is most common on the shelf crest and is associated with fenestral laminated rocks (Rock Types 3 and 4) and pisolitic rocks (Rock Type 5). Three types of internal sediment occurs within sheet cracks. These three types are: 1) red iron stained dolomite pellets, 2) dolomite pellet silt, and 3) microcrystalline dolomite. All three types can occur interlaminated within one sheet crack.

13. Two forms of cyclic sedimentation are present, sandstone-carbonate cycles and cycles within carbonate strata. The sandstone-carbonate cycles consist of a sandstone unit grading into an overlying carbonate unit. Sea level changes on the order of 5 to 10 m were

probably responsible for the sandstone-carbonate cycles. My interpretation supports Pray's (1977) theory that sandstone units were deposited at a high sea level stage and carbonate units at a low sea level stage.

14. Within the shelf crest strata cycles consist of Rock Type 5, pisolitic rocks, grading into either Rock Type 3 or 4, fenestral laminated rocks. Another type of cycle in shelf crest strata the "walnutite cycle" (Esteban and Pray, 1976, 1977) consists of fenestral peloid packstones overlain by in-place pisolites, overlain by clastic pisolites. In the transition zone from inner shelf to shelf crest the cycles consist of algal mat(?) laminated mudstones with local lenses of evaporite molds and burrows alternating with fenestral laminated wackestones. In the transition zone from shelf crest to outer shelf the cycles consist of skeletal or nonskeletal grainstones, Rock Type 1 or 2, alternating with fenestral laminated packstone-grainstones Rock Type 3. Cycles within the carbonate strata are the result of changes in sea level on the order of 1/2 to 2 m, a conclusion reached by Smith (1974b) and Esteban and Pray (1975, 1976, 1977).

15. The position and width of shelf environments were variable with time. The Hairpin Dolomite is characterized by lateral and vertical abundance of tepees and pisolitic rocks. In comparison with the Hairpin Dolomite, the Triplet Dolomite is characterized by a paucity of tepees and pisolites. The Basal Dolomite has characteristics that lie between those of the Triplet Dolomite and

the Hairpin Dolomite; tepees and pisolites though present and abundant, are not as abundant and widespread as in the Hairpin Dolomite.

The author interprets the widespread abundance of tepees and pisolitic rocks in the Hairpin Dolomite to have resulted from a slow rate of subsidence during deposition. During deposition of the Triplet Dolomite the rate of subsidence was high, therefore limiting the duration sediment spent in the peritidal environment. The rate of subsidence for the Basal Dolomite falls in between the rates for the Hairpin Dolomite and the Triplet Dolomite.

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APPENDIX

The following article is reprinted without any changes from: M. E. Hileman and S. J. Mazzullo (eds.), Upper Guadalupian facies, Permian Reef complex, Guadalupe Mountains, New Mexico and west Texas, 1977 Field conference guidebook: Permian Basin Section Soc. Econ. Paleontologists and Mineralogists, Pub. 77-16, p. 437-450.

FACIES MOSAIC OF THE UPPER YATES AND LOWER TANSILL FORMATIONS, WALNUT AND RATTLESNAKE CANYONS, GUADALUPE MOUNTAINS, NEW MEXICO

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INTRODUCTION

A stratigraphic interval approximately 150 feet thick centered on the contact between the Yates and Tansill Formations in the eastern Guadalupe Mountains was studied by the authors (M.S. theses) during the summer of 1976. This report presents preliminary results of the work.

Previous work by Tyrrell (1969), Dunham (1972), Smith (1974) and others has delineated broad regional facies belts. Our study attempted to determine the more detailed facies of individual units over a narrow belt (2 miles wide and 7 miles long) parallel and adjacent to the Capitan Front, and to provide a facies mosaic for better interpretations of the pisolite and associated facies. Our study focused on the strata of the grainstone and packstone belts shown by Smith (1974). Our data results from detailed observations and sampling of well-exposed stratigraphic sections and from walking out correlative units between these sections has yielded information on the geometry and local variability of the strata.

Douglas Neese worked in Walnut Canyon on detailed facies mapping parallel and 1-2 miles behind the Capitan Front. Arthur Schwartz worked in Rattlesnake Canyon on detailed mapping perpendicular to the Capitan Front. A geologic map showing locations of measured stratigraphic sections and a cross-section showing major stratigraphic units is presented in Figure 1.

The upper Yates-Tansill stratigraphic interval was chosen for study because exposures are excellent and accessible, the contact between the Yates and Tansill is easily recognized and several sandstone units occur that are laterally persistent and excellent for correlation. Informal names were applied to units in the upper Yates and lower Tansill Formations for ease of reference. The upper Yates consists of two informal divisions known as the Hairpin Dolomite and the Triplet Unit in this report. The lower Tansill consists of a unit, here informally referred to as the Basal Dolomite.

STRATIGRAPHIC SECTIONS

Stratigraphic sections were plotted in the field at a scale of 10 feet to the inch. Samples were taken at lithologic changes, and every 2 to 5 feet where possible. Approximately 1000 samples were taken during the course of the field

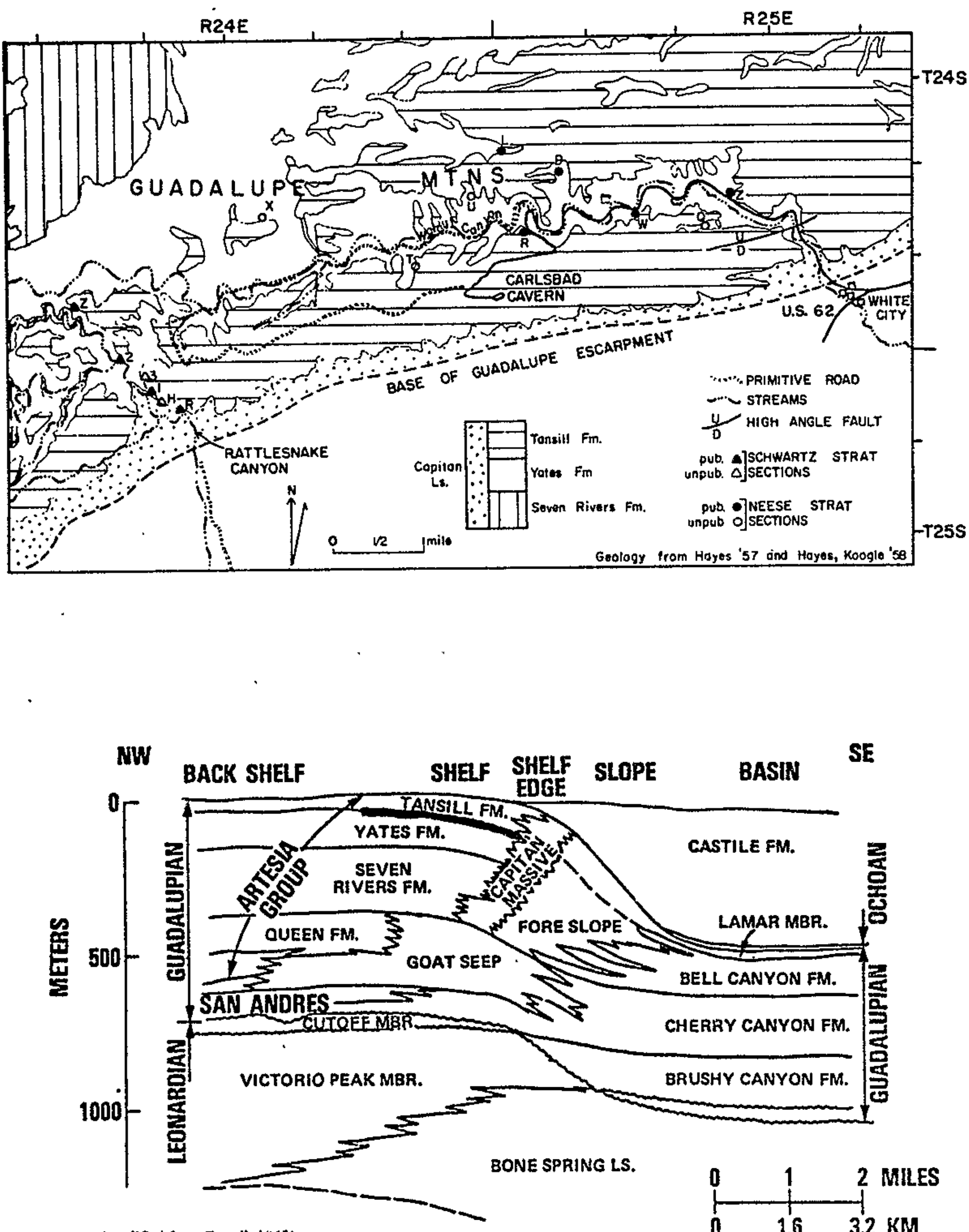


Figure 1. - (Top) Geologic map of Rattlesnake and Walnut Canyon area of southeastern Guadalupe Mountains showing location of stratigraphic sections. (Bottom) Cross-section showing major stratigraphic units. The dark line of the Yates-Tansill Formation contact shows the location of this study.

work. Many of these samples have been slabbed and polished and approximately 150 thin sections have been made. Microscopic examination of these samples was made to determine grain types and abundance, sorting and abundance of mud and other data. The resulting information, combined with field data, permittied the classification of the many rock types and sequences into major facies groups. These facies classifications are summarized below.

The key stratigraphic sections showing the classification of their facies are shown in cross-sections constructed parallel and perpendicular to the shelf edge. These are shown in Figures 2, 3 and 4.

FACIES CLASSIFICATION

The facies classification adopted herein is based on composition, depositional texture, grain type, and abundance of grains present. Two sandstone facies are characterized by abundance of siliclastics (quartz and some feldspar). Seven are largely carbonates, of which 6 have grain-supported textures and one is mud-supported. Most carbonate rocks are dolomite, but dolomitization has not altered details of depositional texture. The grain-supported carbonate facies are:

1. Skeletal-rich limestone
2. Skeletal-rich ("restricted") dolomite
3. Mixed skeletal — non-skeletal dolomite
4. Non-skeletal dolomite
5. Pisolith-poor dolomite
6. Pisolith-rich dolomite

The sandstone facies are :

7. Sandstone (siliclastic)
8. Mixed sandstone-dolomite

One carbonate facies is mud-supported :

9. Mud-supported dolomite

Grain-Supported, Skeletal-Rich Limestone

This facies is composed of grainstones and minor mud-lean packstones containing a biota characteristic of normal marine water. Skeletal grains occurring in the rocks are largely whole and fragmented bryozoans, echinoderms, forams, Tubiphytes sp., sponges, gastropods, and calcareous algae (dasycladacean and some encrusting forms). Micritic non-skeletal grains, mostly peloids and composite grains are common, especially in the finer grained and/or more poorly sorted rocks. Laminated coatings and micritic envelopes occur on some of the skeletal and non-skeletal grains. Unlike most rocks of the shelf, rocks of this facies are predomin-

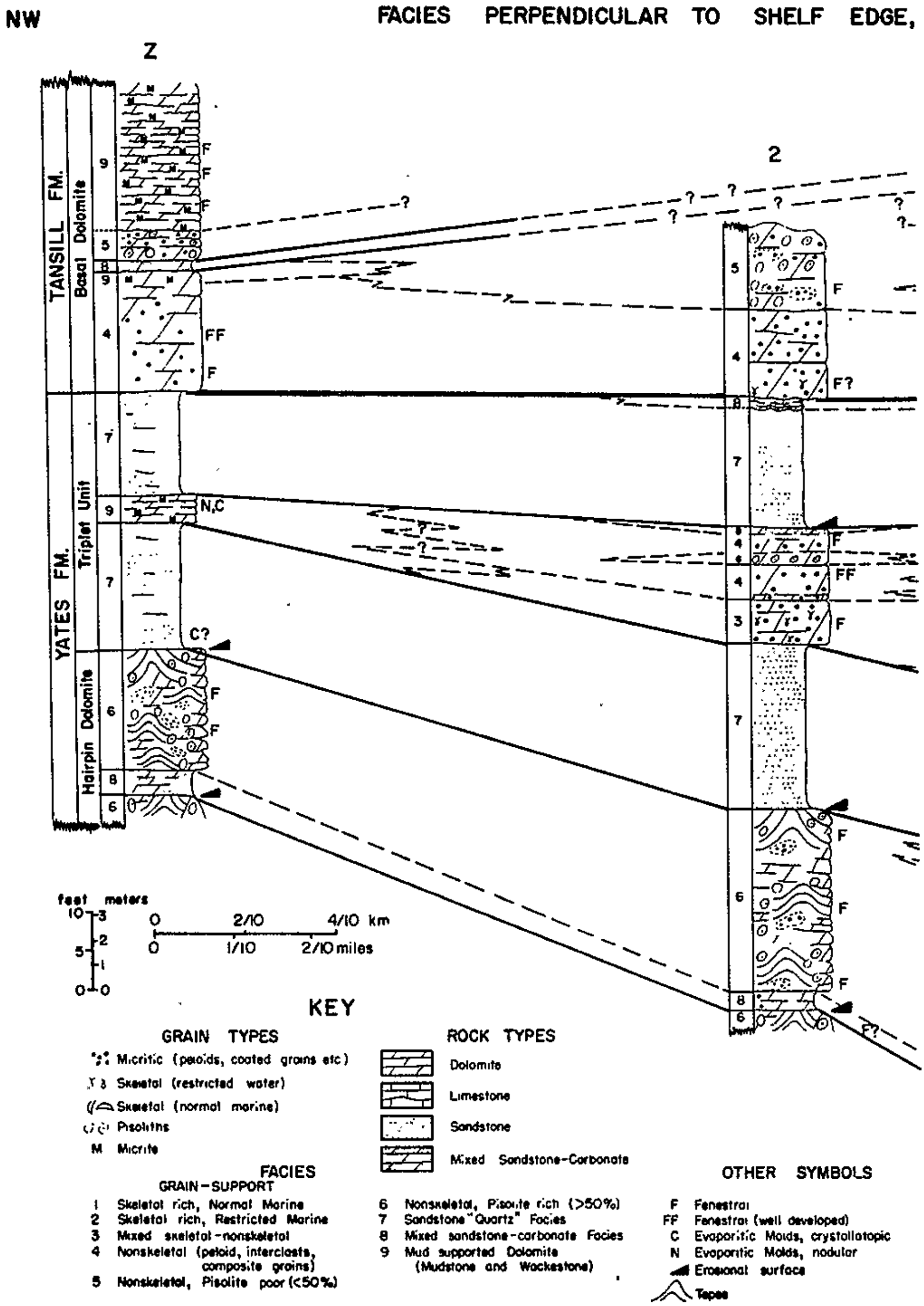
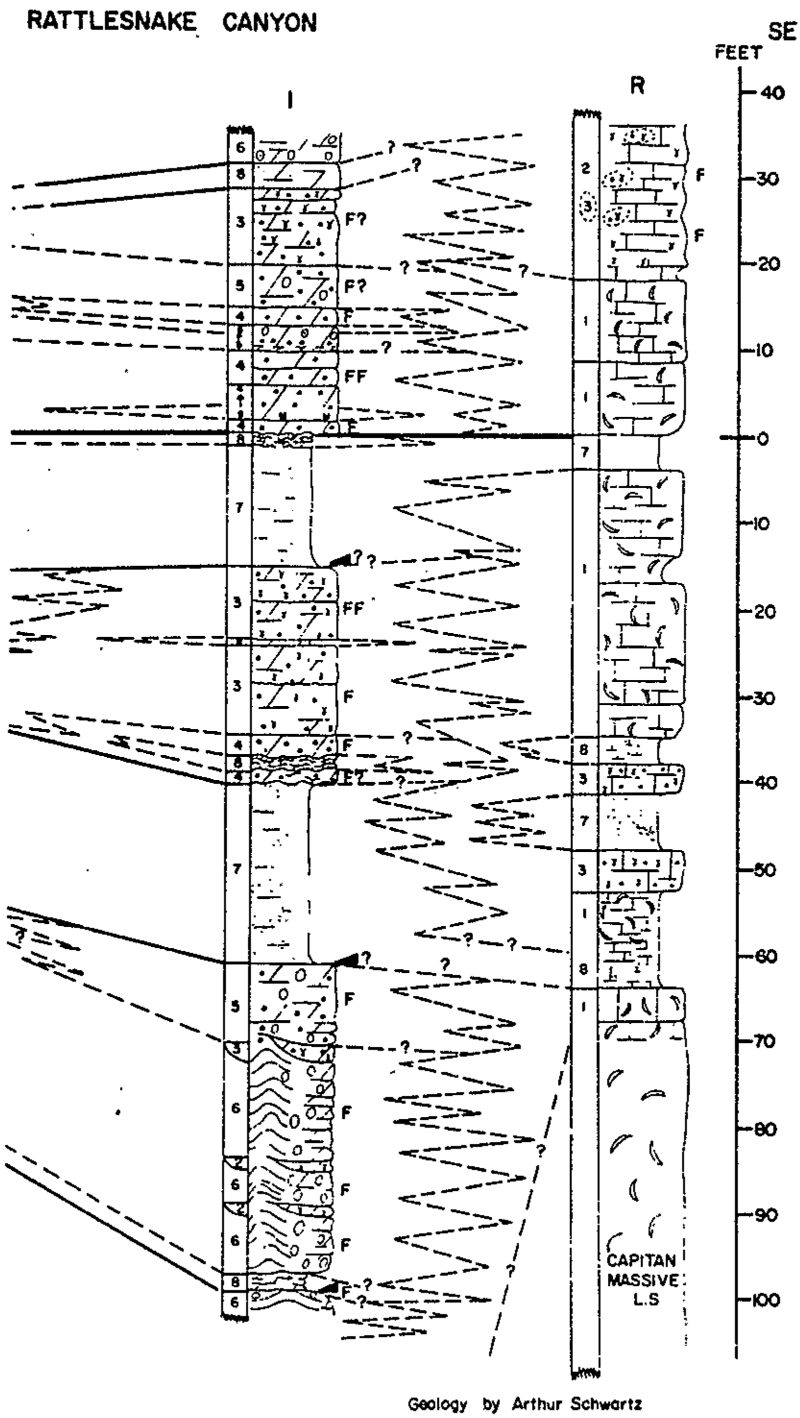


Figure 2. - Facies cross-section in Rattlesnake Canyon.


RATTLESNAKE CANYON
SE
FEET
I
R
CAPITAN
MASSIVE
L.S
Geology by Arthur Schwartz

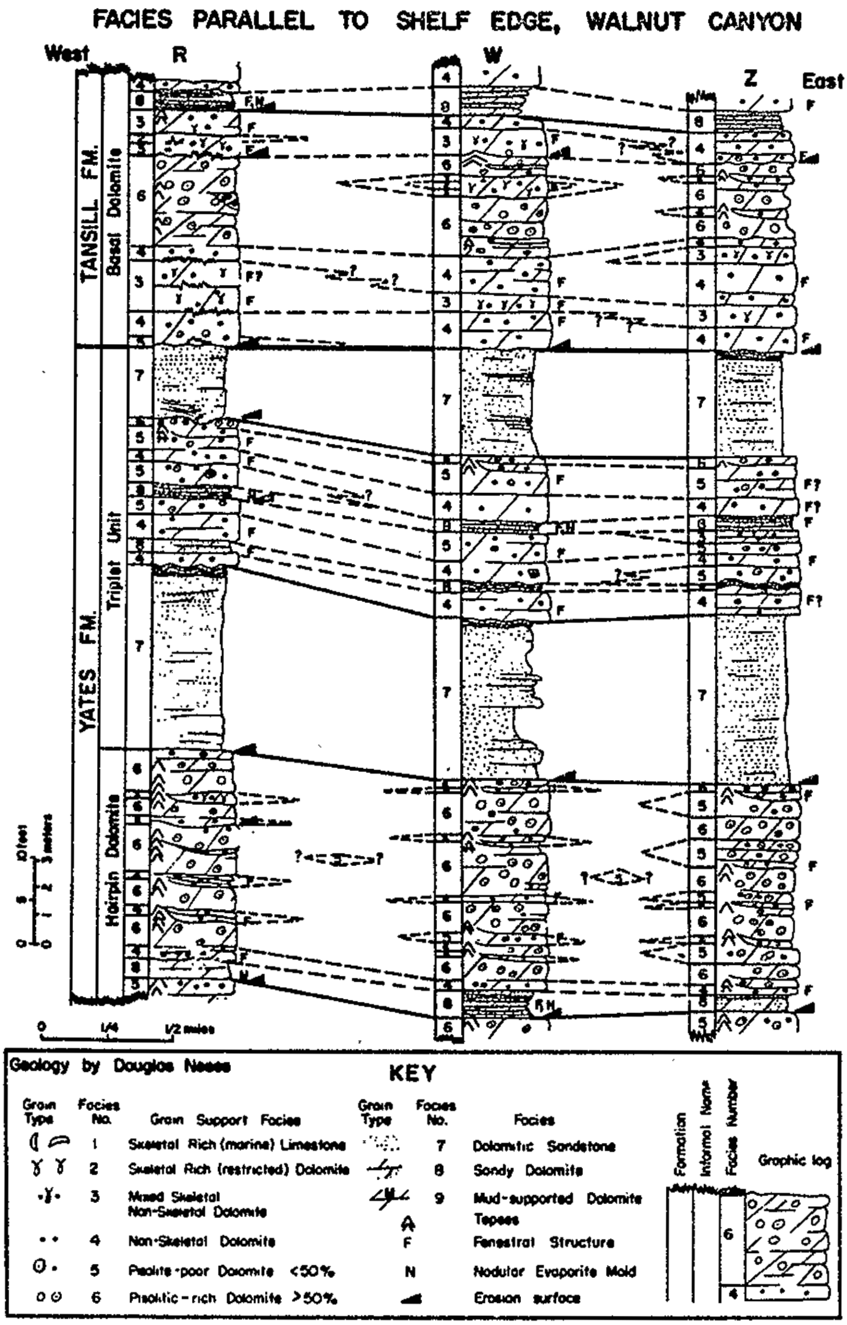


Figure 3. - West-east facies cross-section in Walnut Canyon.

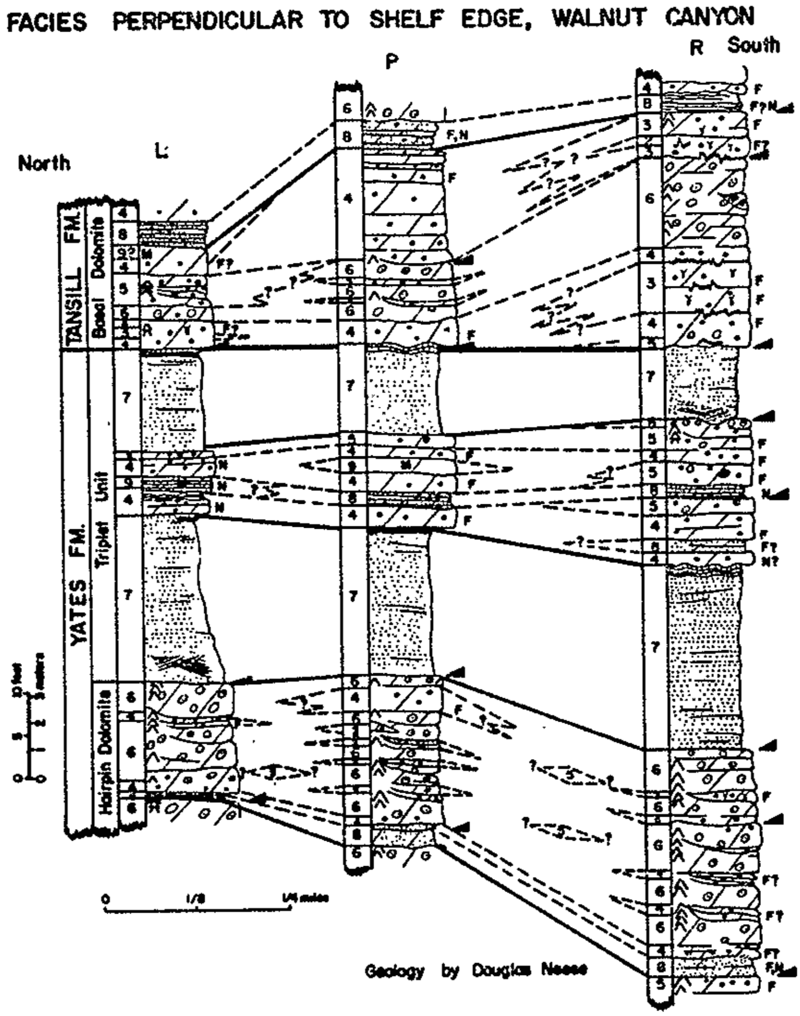


Figure 4. - North-south facies cross-section in Walnut Canyon.

antly limestone. These rocks occur as massive beds 3 - 5 m in thickness, and locally display low-angle cross-sets 10 - 30 cm thick. Cross-bed dips are primarily toward the basin (easterly), although shelfward dips (westerly) occur as well. The grains present in the cross-bedded units tend to be oolitically coated. The rocks of this facies occupy a narrow belt 100-500 m wide immediately shelfward of the Capitan Limestone. In this belt the shelf sandstones have thinned and are present as thin ($\frac{1}{4}$-$1\frac{1}{2}$ m) tongues that pinch and swell locally, and cannot be traced into the Capitan. Rocks of this facies were probably deposited in a turbulent subtidal environment.

Grain-Supported, Skeletal-Rich ("Restricted") Dolomite

This facies is composed of packstones and grainstones that contain a low diversity "restricted" water biota. Restricted water is used in this report to indicate that conditions of temperature and salinity were variable to the extent that they limited the development of a high diversity, normal marine biota. Grains are primarily skeletal, mostly dasycladacean algae, gastropods, forams, and ostracods. Dasycladacean algae are predominant, forming at least 80% of the skeletal grains. Non-skeletal grains are minor, including peloids, coated and composite grains, and intraclasts. Bedding is even and parallel; beds are $\frac{1}{4}$-$1\frac{1}{2}$ m thick. Some rocks of this facies also occur as small pockets or lenses within rocks of the mixed skeletal — non-skeletal facies. Rocks of this facies are present as partial fillings of inter-tepee depressions. Rocks of this facies are most commonly associated with the more basinward rocks of the mixed skeletal — non-skeletal facies and with the more shelfward of the skeletal-rich limestone facies. They probably were deposited under conditions very similar to those of the mixed skeletal — non-skeletal facies. They probably represent deposition in relatively shallow water at places of moderate turbulence when water conditions were favorable for high production rates of dasycladacean algae.

Grain-Supported, Mixed Skeletal — Non-Skeletal Dolomite

This facies consists of grainstones primarily, and minor packstones composed of skeletal and non-skeletal grains in approximately equal abundance (more than 1/3 of each). Non-skeletal grains include peloids, composite grains and coated grains (up to 5 mm). Skeletal grains are representative of a restricted water biota and consist mostly of dasycladacean algae, gastropods, forams, and ostracods. Both skeletal and non-skeletal grains may locally have thin laminated coatings. Locally, rocks of this facies are burrowed. Fenestral fabric is common and various shapes of fenestrae occur. Probable nodular anhydrite molds ($\frac{1}{2}$-1 cm) occur locally.

Rocks of this facies commonly occur in even parallel beds $\frac{1}{2}$ to $1\frac{1}{2}$ m thick. They also occur in lenses associated with tepees. The lenses thin over tepee crests and thicken between them. Rocks of this facies are most abundant in areas immediately shelfward of rocks belonging to the skeletal-rich limestone facies. Farther shelfward this facies loses the skeletal components and intertongues with rocks of the grain-supported, non-skeletal facies. These rocks were probably deposited in a subtidal to intertidal environment in quiet to occasionally turbulent water.

Grain-Supported, Non-Skeletal Dolomite

This facies consists primarily of mud-lean packstones and grainstones in which over 2/3 of the grains are non-skeletal. The non-skeletal grains include peloids, composite grains, coated grains and intraclasts. The skeletal grains present are representative of a restricted water biota. They are largely dasycladacean algae, gastropods, forams, and ostracods. Some grains have a thin laminated coating. Most grains are sand to granule size except for larger whole or fragmented gastropods. Fenestral fabric occurs commonly. Burrows are present locally. In the most shelfward localities nodular anhydrite molds ($\frac{1}{2}$-1 cm) occur. Bedding in these rocks is even and parallel, layers are mostly $\frac{1}{4}$ - 1 m thick, but thinner beds occur within inter-tepee depressions. Towards the basin (to the east), beds of this facies commonly intertongue with the more shelfward beds of the mixed skeletal — non-skeletal facies. In the more shelfward outcrops they intertongue with dolomite of the mud-supported facies. These rocks were probably deposited in moderately quiet water of a subtidal to intertidal environment.

Grain-Supported, Pisolith-Poor Dolomite

This facies is composed primarily of packstones and some grainstones and contains between 10 and 50% pisoliths. Other grains include peloids, composite grains, intraclasts, and a few oncolites. Some skeletal fragments such as dasycladacean algae and forams occur but are minor. Most commonly the grains are poorly sorted. The pisoliths in this facies are small, mostly of granule size (2-4 mm). Their nuclei are mostly peloids and some skeletal fragments. They appear to be more micritic than the pisoliths of the pisolite-rich facies. Fenestral fabric, some distinct and others obscure, occurs locally.

The rocks of this facies may occur in distinct beds, up to 1 - 1$\frac{1}{2}$ m thick, especially in the Tansill Formation. However, more commonly the rocks occur in thinner units (less than $\frac{1}{2}$ m thick) that are transitional beds between rocks of the pisolite-rich facies and rocks of the grain-supported, non-skeletal facies. Rocks of the pisolite-poor facies have wide lateral continuity. In the most shelfward outcrops these rocks intertongue with rocks of the mud-supported dolomite facies. Rocks of this facies occur commonly in the uplifted central cores and flanks of tepees. They were probably deposited in subtidal to intertidal environments in turbulent to moderately turbulent water of hypersaline quality sufficient to repress skeletal grain production.

Grain-Supported, Pisolith-Rich Dolomite

This facies is composed of packstones and grainstones containing more than 50% pisoliths. Other rock types present in minor abundance as pockets or lenses are skeletal grainstones, non-skeletal grainstones, and mixed skeketal — nonskeletal grainstones. In the pisolith-rich rocks, sand to granule size peloids and composite grains, and fine sand to silt size siliclastics occur in minor abundance. Pisoliths in this facies range upward to 2 - 3 cm in size. Pisolith-rich units are almost always associated with tepee structures but the converse is not true. The pisolitic rocks are more common in the inter-tepee depressions. Beds of pisolith-rich rocks are laterally continuous only within individual depressions. The major occurrence of this facies is in the Hairpin Dolomite (informal name) of the Yates Formation, where pisolith-rich units as thick as $1\frac{1}{2}$ m occur. We consider the pisoliths as syndepositional particles, and believe this facies was deposited in a shallow peritidal environment (Esteban and Pray, 1975, 1976).

Sandstone Facies

This facies is composed of rocks containing predominantly siliclastic grains and largely devoid of carbonate grains. Rocks consist of well sorted, subangular, very fine sand to silt size quartz and some feldspar. The rock is cemented mainly by dolomite. Iron oxide as stains, concretions ($\frac{1}{2}$ - 10 cm), and crusts is abundant throughout. In the field the rocks of this facies weather to form a grassy covered recessive slope that is readily recognized. Most rocks are yellowish-orange on weathered surfaces and light gray on fresh surfaces. Bedding surfaces are usually poorly defined. Basal contacts of units with interbedded carbonates appear sharp and upper contacts are sharp or gradational. The weathering characteristics of the rocks makes the upper contact difficult to observe. Distinct primary sedimentary structures are uncommon. Obscure parallel to wavy lamination is common, and some small (5-30 cm) low-angle cross-sets occur. Rocks of this facies occur in two sheets (one 5 m, the other 6 m thick), both in the Yates Formation in our study area. They are laterally continuous over the entire study area until within 100 m of the shelf edge where they pinch out and/or intertongue with the carbonate rocks of the shelf edge. The rocks of this facies were deposited in a subaqueous environment at the shelf edge, and in a probable subaqueous environment over the area studied.

Mixed Sandstone - Dolomite Facies

This facies is composed of sandy dolomite and dolomitic sandstone, with both siliclastic and carbonate grains. Grains include silt to very fine sand size quartz (some feldspar), pisoliths, and peloids. Skeletal fragments occur and increase in abundance towards the shelf edge. These rocks are cemented with dolomite. The rocks of this facies weather slightly recessive.

In outcrop the rocks of this facies occur in thin (generally $\frac{1}{2}$ - 1 m thick) sheets. They are remarkably persistent laterally for their thickness and pinch out as they approach to within 100 m of the shelf edge. Two types of bedding styles occur : parallel and wavy. The former has even, parallel laminae (approximately 2 mm) and thin (2-10 cm) beds. The lamination may be disrupted by fenestral fabric or possible fluid-escape structures. This type of unit commonly overlies pisolith-rich strata associated with tepees. The basal contacts of these units are distinct planar erosion surfaces that truncate tepees in the underlying rock units. These units of dolomitic sandstone grade upward into dolomite. The wavy bedded deposits have lenticular, wavy, thin (2-5 cm) beds, the waves having amplitudes of 5 - 8 cm. These beds are underlain and overlain by discontinuous or semicontinuous pockets of soft, earthy weathering, sandy carbonate. Units with wavy bedding are not as laterally persistent as units with parallel bedding. Fenestral fabric and cavities suggestive of nodular anhydrite molds ($\frac{1}{2}$ - 1 cm) occur in both the wavy and parallel units, but are more common in the latter. The rocks of this facies were deposited in a subaqueous environment.

Mud-Supported Dolomite

This facies is composed primarily of dolomite mudstone and minor grain-lean dolomite wackestones. The grains are mostly peloids of medium sand size and skeletal fragments of a highly restricted (hypersaline) water biota (ostracods, calcispheres, gastropods and dasycladacean algae). Evidence of evaporites is abundant as nodular molds of anhydrite ($\frac{1}{2}$ - 1 cm) or crystallotopic molds of gypsum or anhydrite.

These rocks occur in outcrop in even, parallel beds ranging in thickness from 5 to 30 cm. At some localities, these rocks occur in units 10 to 20 cm thick, consisting of discontinuous, wavy parallel beds 2 - 5 cm thick, with amplitudes of 5 - 8 cm. Some of the beds exhibit a crinkly (approximately 1 mm) lamination that strongly resembles laminated algal mats. Burrows are locally present in the thicker beds. The rocks of this facies may be associated with rocks of the grain-supported, non-skeletal facies. They are more commonly found in the most shelfward portions of the study area, and suggest deposition from quiet, hypersaline water.

CONCLUSIONS

Preliminary conclusions from this study of the upper Yates and lower Tansill Formation of the marginal mound and outer shelf slope strata in Walnut and Rattlesnake Canyons are :

1. Stratigraphic units (0.5 - 6 m) of the sandstone (siliclastic) and mixed sandstone (siliclastic-carbonate) facies persist throughout the study area with but minor thickness variations. Beds of these facies pinch out abruptly at the shelf edge 100 - 300 m away from the Capitan massive facies. Few or no facies variations are evident in these units parallel or perpendicular to the shelf edge ; they are believed to be subaqueous deposits.

2. Facies of the carbonate units are laterally persistent parallel to the shelf edge. Three major carbonate units and several subunits (0.5 - 3 m) have been traced laterally for 10 miles. Perpendicular to the

shelf edge their facies change rapidly. Toward the shelf, carbonate rocks become more poorly sorted and increase in micrite content and evaporitic traces. Shelfward of the marginal mound, grain size diminishes.

3. Carbonate units between the two major sandstone units (upper Yates) decrease in thickness by one-half over the three mile interval shelfw to the Capitan.

4. Abundance and diversity of skeletal grains increases towards the Capitan as reported by others ; however, this change from normal marine to more restricted or hypersaline environments is abrupt in one upper Yates carbonate (the Hairpin Dolomite) and gradational in two others (the Triplet Dolomite and Basal Tansill Dolomite).

5. Facies patterns perpendicular to the shelf edge indicate that the relative width of facies belts was highly variable through time. Moreover, at various times some facies belts were absent. The carbonate facies studied did not simply prograde "in step" with the Capitan Limestone.

6. Most tepees are associated with fenestral laminated beds of the mixed (restricted) skeletal and non-skeletal facies. Although tepees are associated with pisolith-rich rocks, the pisoliths are best developed in inter-tepee depressions. Pisolith-rich carbonates and abundant, large tepees occur across at least two miles of the shelf, extending to within less than a half-mile of the shelf edge in the lower carbonate (Yates). They are poorly developed and occur in a narrow facies belt in the uppermost Yates and lowermost Tansill carbonate units.

7. Erosional surfaces are common at the upper boundaries of major carbonate units and within carbonate units. The most pronounced erosional surfaces are those where the sandstone facies or mixed skeletal — non-skeletal facies overlie carbonate units with truncated tepees. Some erosion surfaces show thin zones of alteration, representing possible soil development. Although erosion surfaces are found close to the shelf edge, the alteration zones are localized to areas 1 - 2 miles behind the Capitan Front.

8. The three major carbonate units (Hairpin Dolomite, Triplet Unit and Basal Dolomite) differ significantly in their shelfward facies progression, indicating that the governing environmental conditions determining carbonate facies, such as water quality, depth and turbulence, were only partly related to the distance behind the shelf edge. The major differences of these three carbonate units are summarized below :

Lower carbonate (Hairpin Dolomite), upper Yates Formation

The pisolith-rich facies dominates this unit, and is associated with large and abundant tepees. Beds of the mixed skeletal — non-skeletal facies alternate with beds of the pisolith-rich facies. The broad belt of pisolith-rich facies and tepees extends across most of the study area and persists to within $\frac{1}{4}$ - $\frac{1}{2}$ mile of the shelfward edge of the Capitan.

Middle Carbonate (Triplet Dolomite), upper Yates Formation

This unit is the best example of shelfward thinning. The Triplet Dolomite changes from 1 to 6 meters in thickness over a distance of three miles perpendicular to the Capitan Limestone. The unit has more admixed siliclastic silt and sand-size grains and micrite than the other two, and pisolith-rich facies and tepees are minor.

Upper Carbonate (Basal Dolomite), lower Tansill Formation

Mixed skeletal — non-skeletal facies persist in this unit through much of the study area, suggesting a deeper shelfward penetration of more open marine water than for the other two units. Overlying pisolith-rich units are in a basinward prograding, shoaling upward sequence that terminates upward at an erosion surface overlain by sandstone.

ACKNOWLEDGMENTS

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_________, 1969, Criteria useful in interpreting environments of unlike but time-equivalent carbonate units (Tansill-Capitan-Lamar), Capitan Reef Complex, West Texas and New Mexico, in G. M. Friedman (ed.), Depositional Environments in Carbonate Rocks : Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 14, p. 80-97.

LOCATION OF MEASURED SECTIONS

The location of each measured section is given as the location of the contact between the Yates and Tansill Formations within that section. Refer to U. S. Geological Survey Quadrangle Maps GQ 112 and GQ 167. Also, see Figure 6 in text for a map showing the relative positions of measured sections.

Section A is 1585 feet from the east line and 1900 feet from the south line of Section 4 in Township 25 South, Range 24 East. Elevation approximately 4250 feet.

Section B is 1850 feet from the west line and 2110 feet from the south line of Section 4 in Township 25 South, Range 24 East. Elevation approximately 4300 feet.

Section C is 1160 feet from the west line and 2640 feet from the south line of Section 4 in Township 25 South, Range 24 East. Elevation approximately 4350 feet.

Section D is 635 feet from the west line and 1585 feet from the north line of Section 4 in Township 25 South, Range 24 East. Elevation approximately 4450 feet.

Section E is 660 feet from the east line and 130 feet from the north line of Section 5 in Township 25 South, Range 24 East. Elevation approximately 4500 feet.

Section F is 1585 feet from the west line and 2640 feet from the south line of Section 32 in Township 24 South, Range 24 East. Elevation approximately 4600 feet.

KEY FOR MEASURED SECTIONS

GRAIN TYPES

Q = siliciclastics (95% quartz)

● = peloids

⦿ = pisolites

= dasycladacean algae

= sponges

= *Tubiphytes* sp.

∞ = intraclasts

= gastropods

★ = crinoids

$ = skeletal, unidentifiable

POROSITY TYPES

B = interparticle

I = intraparticle

E_C = moldic crystallotopic evaporitic

F = fenestral

E_N = moldic nodular evaporitic

EROSION SURFACE =

ACCESSORY CURVES

= fabric present in all rock types in that unit

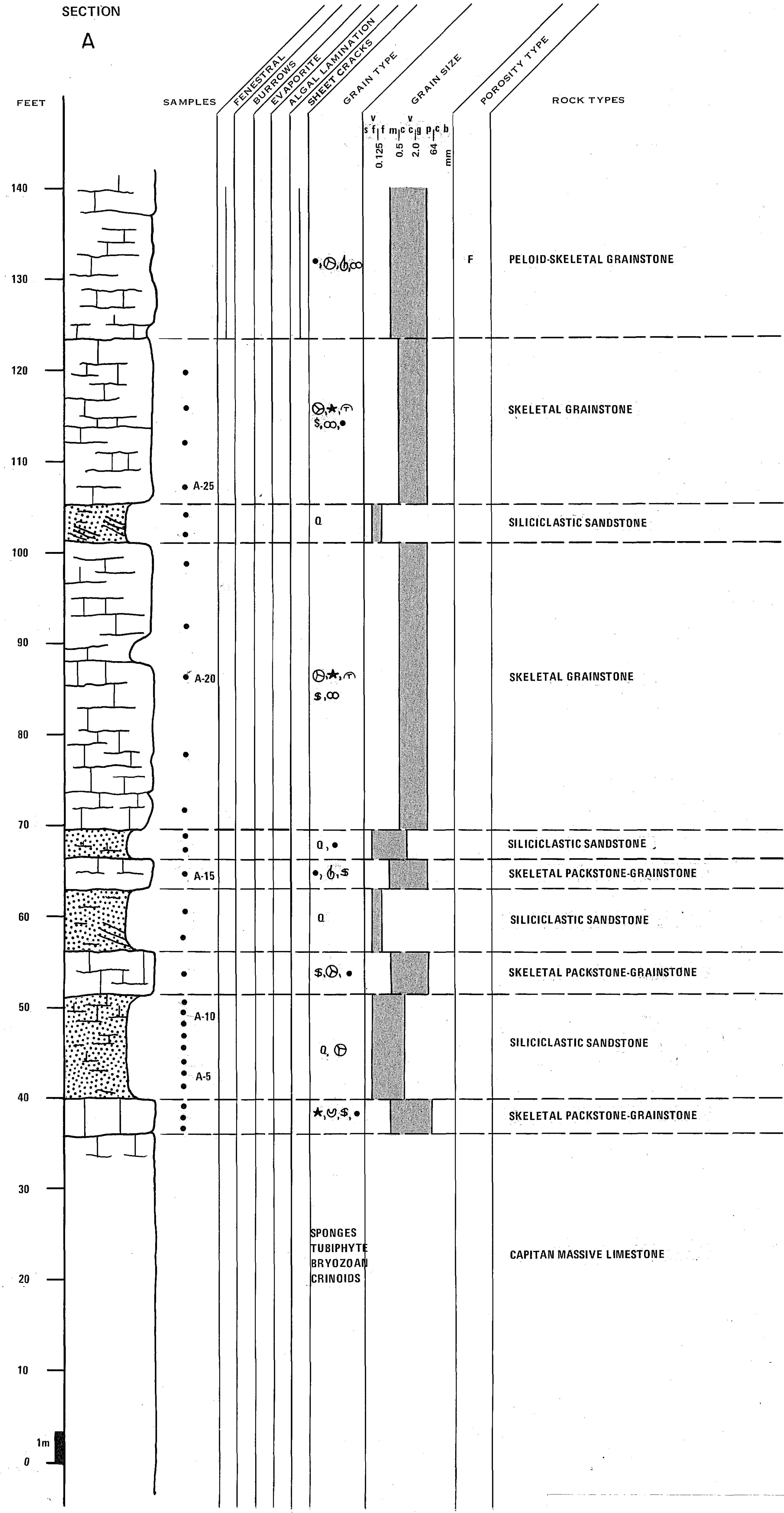
= symbol indicates grain type to which the accessory curve applies

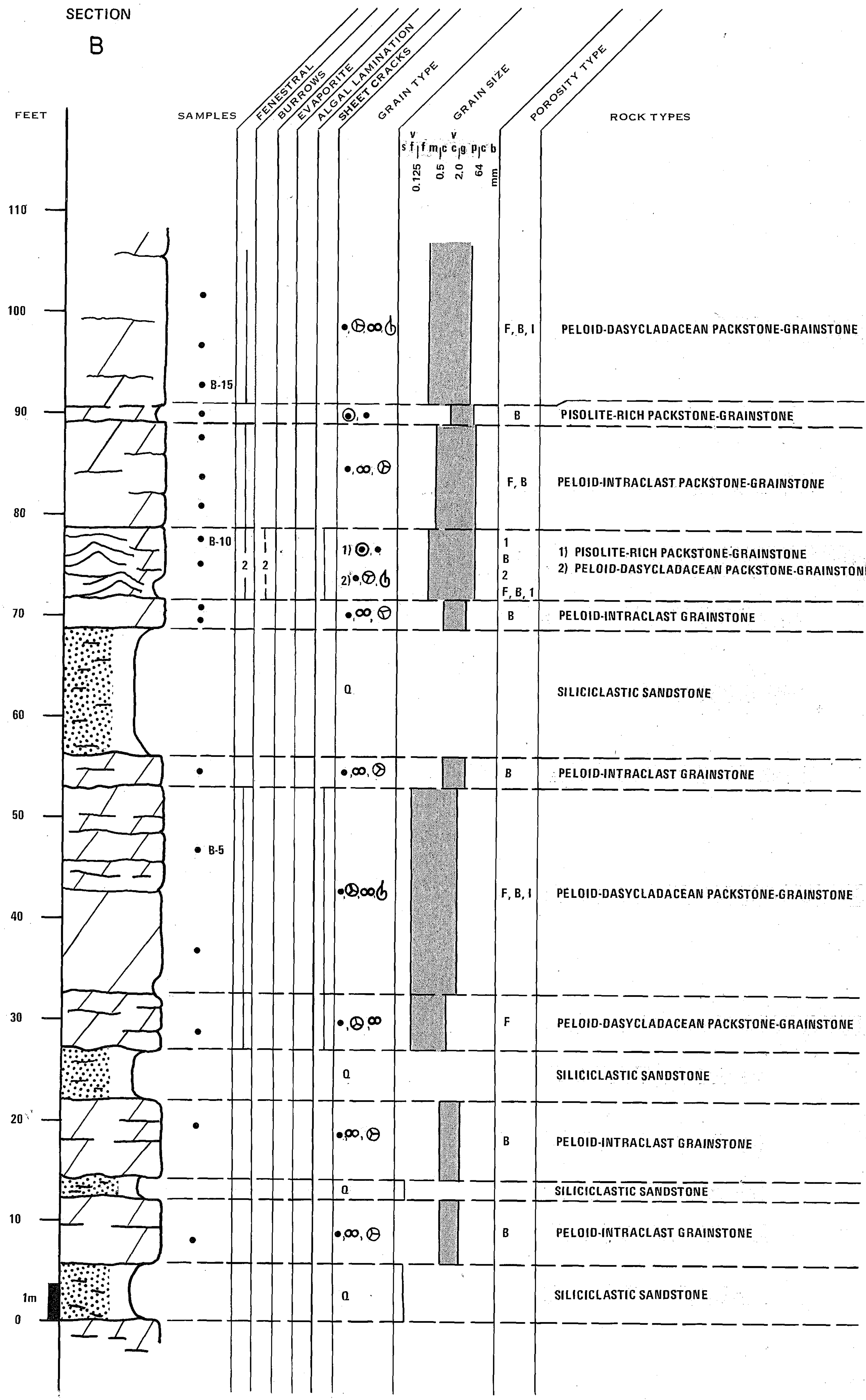
2 = number in accessory curves indicates the rock type of that unit in which the fabric occurs

= indicates a local, or discontinuous occurrence

FACIES PERPENDICULAR TO SHELF EDGE, RATTLESNAKE CANYON

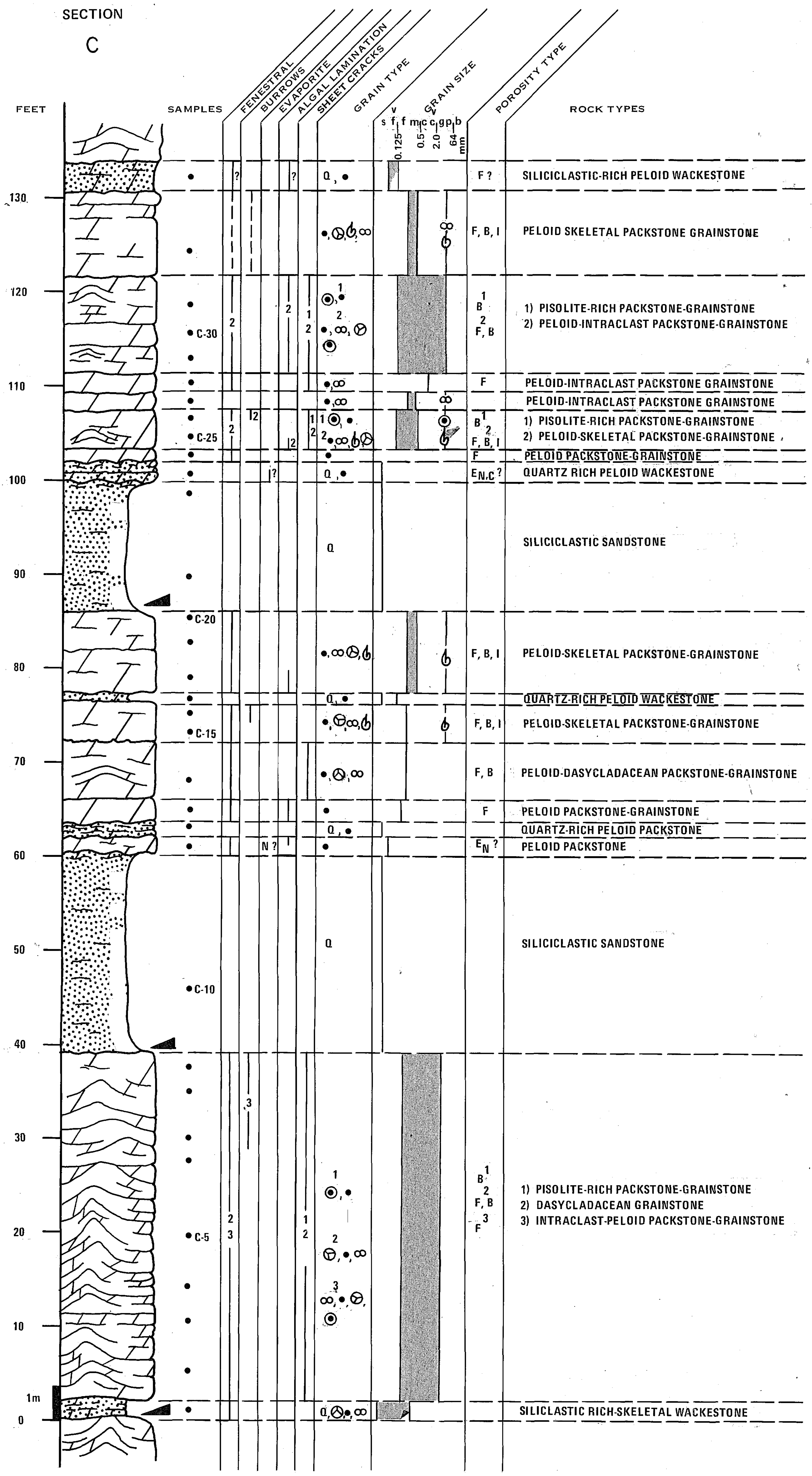
CROSS SECTION
RATTLESNAKE CANYON
FACIES PERPENDICULAR TO SHELF EDGE
790031 A


SECTION
A
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
v s f f m c v c g p c b
0.125
0.5
2.0
64
mm
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
1m
A-25
A-20
A-15
A-10
A-5
F
PELOID-SKELETAL GRAINSTONE
SKELETAL GRAINSTONE
SILICICLASTIC SANDSTONE
SKELETAL GRAINSTONE
SILICICLASTIC SANDSTONE
SKELETAL PACKSTONE-GRAINSTONE
SILICICLASTIC SANDSTONE
SKELETAL PACKSTONE-GRAINSTONE
SILICICLASTIC SANDSTONE
SKELETAL PACKSTONE-GRAINSTONE
SPONGES
TUBIPHYTE
BRYOZOAN
CRINOIDS
CAPITAN MASSIVE LIMESTONE



SECTION
B
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
s f f m c c g p c b
0.125 0.5 2.0 64 mm
110
100
90
80
70
60
50
40
30
20
10
1m
0
B-15
B-10
B-5
F, B, I
PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
B
PISOLITE-RICH PACKSTONE-GRAINSTONE
F, B
PELOID-INTRACLAST PACKSTONE-GRAINSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
1 B 2 F, B, 1
B
PELOID-INTRACLAST GRAINSTONE
Q
SILICICLASTIC SANDSTONE
B
PELOID-INTRACLAST GRAINSTONE
F, B, I
PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
F
PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
Q
SILICICLASTIC SANDSTONE
B
PELOID-INTRACLAST GRAINSTONE
Q
SILICICLASTIC SANDSTONE
B
PELOID-INTRACLAST GRAINSTONE
Q
SILICICLASTIC SANDSTONE

SECTION B



SECTION
C
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
s f f m c c gp b
0.125
0.5
2.0
64 mm
130
120
110
100
90
80
70
60
50
40
30
20
10
1m
0
C-30
C-25
C-20
C-15
C-10
C-5
SILICICLASTIC-RICH PELOID WACKESTONE
PELOID SKELETAL PACKSTONE GRAINSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
PELOID-INTRACLAST PACKSTONE GRAINSTONE
PELOID-INTRACLAST PACKSTONE GRAINSTONE
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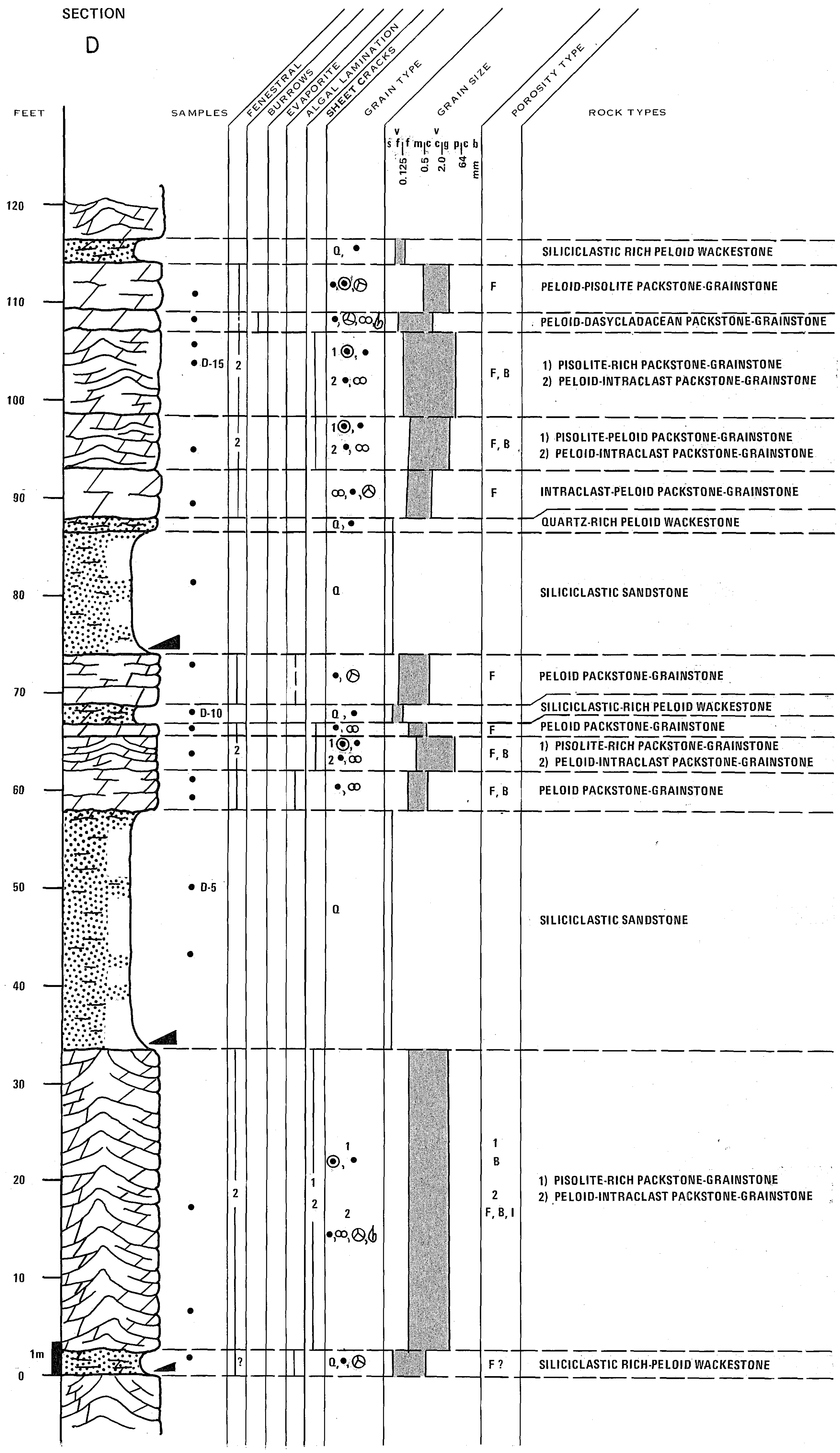
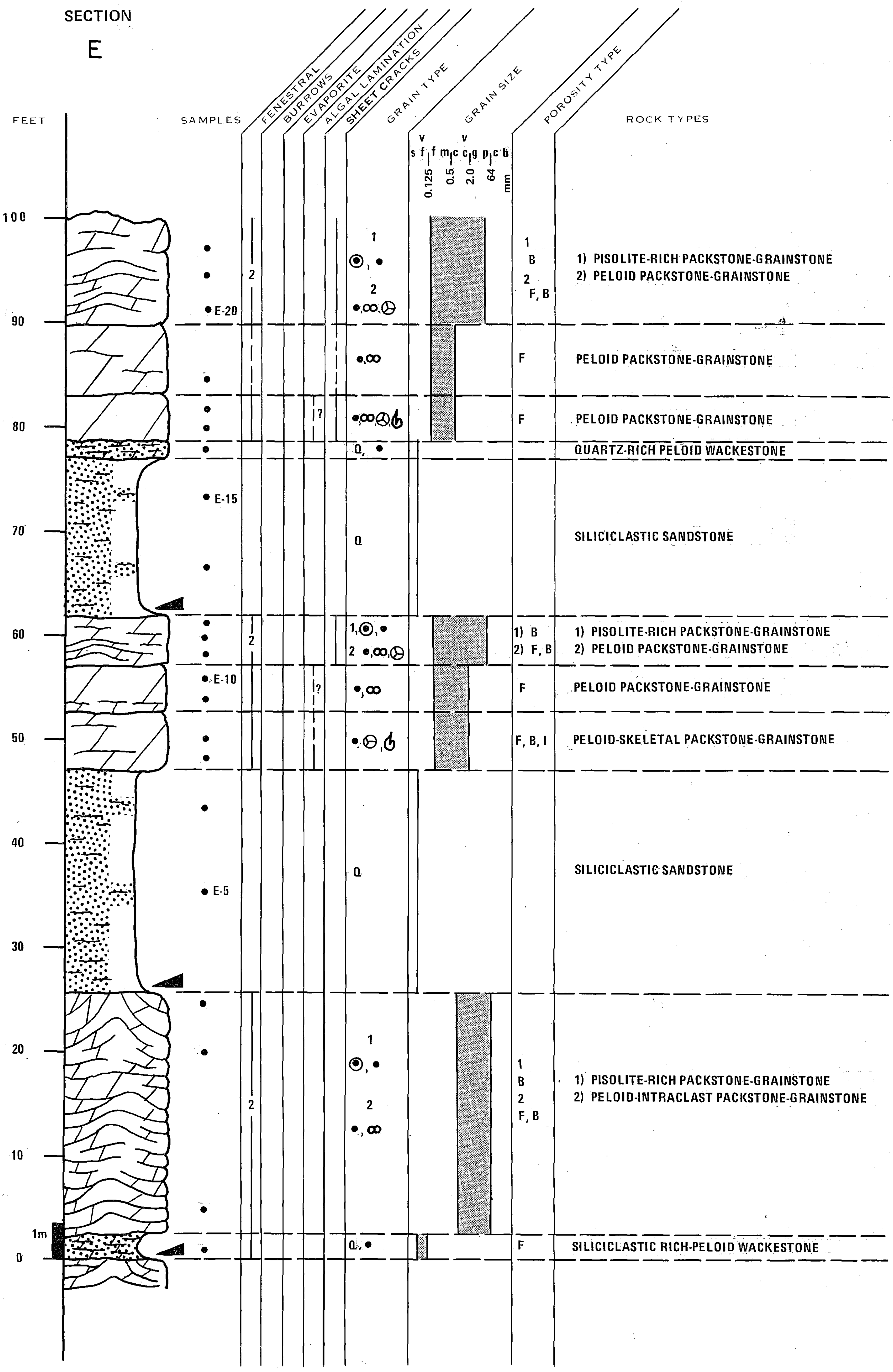
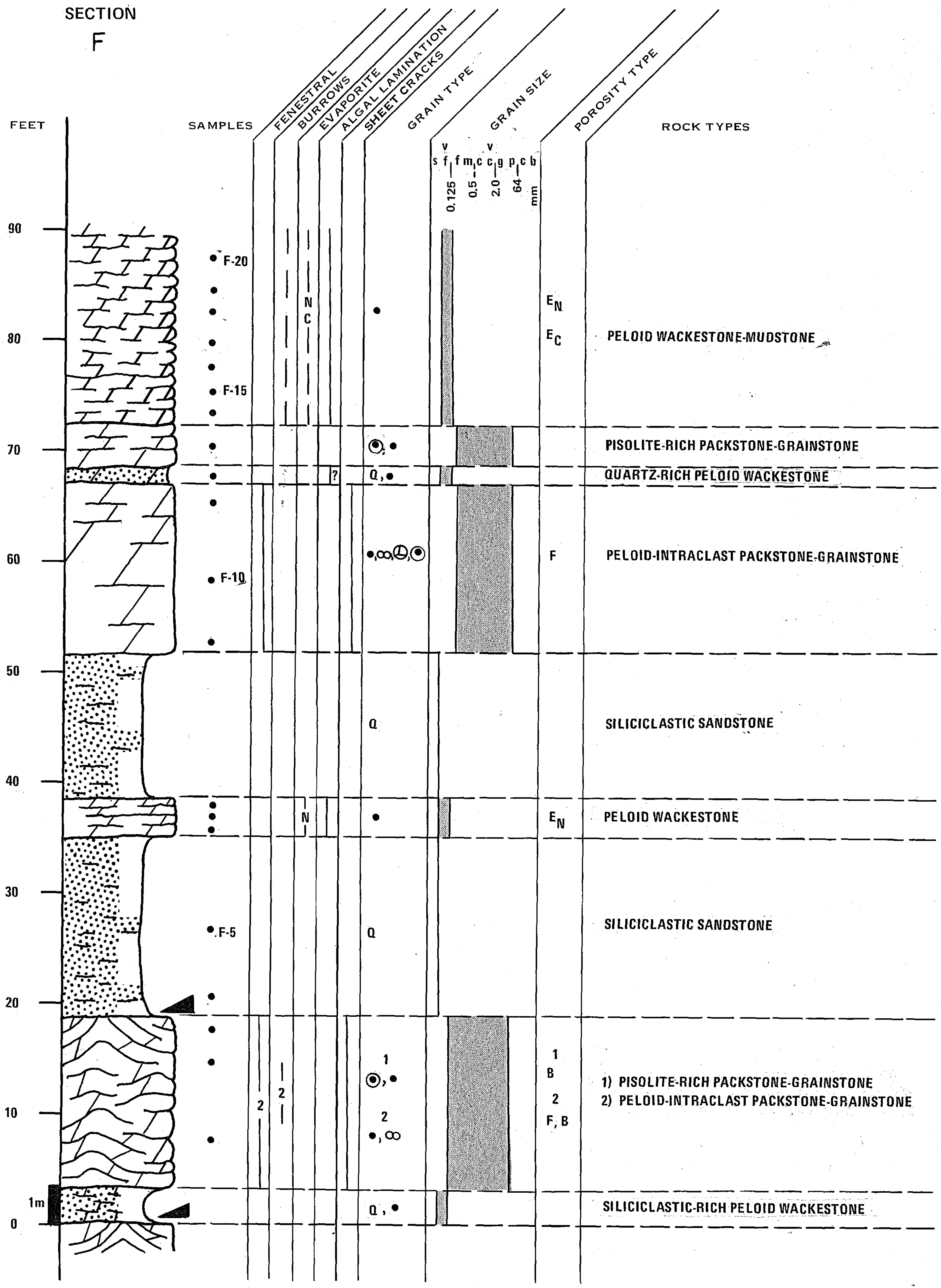
2) PELOID-SKELETAL PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
QUARTZ RICH PELOID WACKESTONE
SILICICLASTIC SANDSTONE
PELOID-SKELETAL PACKSTONE-GRAINSTONE
QUARTZ-RICH PELOID WACKESTONE
PELOID-SKELETAL PACKSTONE-GRAINSTONE
PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
QUARTZ-RICH PELOID PACKSTONE
PELOID PACKSTONE
SILICICLASTIC SANDSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) DASYCLADACEAN GRAINSTONE
3) INTRACLAST-PELOID PACKSTONE-GRAINSTONE
SILICLASTIC RICH-SKELETAL WACKESTONE
F ?
F, B, I
F
E N,C ?
F, B, I
F, B
E N ?


SECTION
D
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
0.125
0.5
2.0
64
mm
120
110
100
90
80
70
60
50
40
30
20
10
1m
0
D-15
D-10
D-5
SILICICLASTIC RICH PELOID WACKESTONE
PELOID-PISOLITE PACKSTONE-GRAINSTONE
PELOID-DASYCLADACEAN PACKSTONE-GRAINSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
1) PISOLITE-PELOID PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
INTRACLAST-PELOID PACKSTONE-GRAINSTONE
QUARTZ-RICH PELOID WACKESTONE
SILICICLASTIC SANDSTONE
PELOID PACKSTONE-GRAINSTONE
SILICICLASTIC-RICH PELOID WACKESTONE
PELOID PACKSTONE-GRAINSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
SILICICLASTIC SANDSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
SILICICLASTIC RICH-PELOID WACKESTONE


SECTION
E
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
s f f m c c g p c b
0.125 0.5 2.0 64 mm
100
90
80
70
60
50
40
30
20
10
0
1m
E-20
E-15
E-10
E-5
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
QUARTZ-RICH PELOID WACKESTONE
SILICICLASTIC SANDSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID PACKSTONE-GRAINSTONE
PELOID PACKSTONE-GRAINSTONE
PELOID-SKELETAL PACKSTONE-GRAINSTONE
SILICICLASTIC SANDSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
SILICICLASTIC RICH-PELOID WACKESTONE


SECTION F
FEET
SAMPLES
FENESTRAL
BURROWS
EVAPORITE
ALGAL LAMINATION
SHEET CRACKS
GRAIN TYPE
GRAIN SIZE
POROSITY TYPE
ROCK TYPES
0.125
0.5
2.0
64
mm
90
80
70
60
50
40
30
20
10
0
1m
F-20
F-15
F-10
F-5
PELOID WACKESTONE-MUDSTONE
PISOLITE-RICH PACKSTONE-GRAINSTONE
QUARTZ-RICH PELOID WACKESTONE
PELOID-INTRACLAST PACKSTONE-GRAINSTONE
SILICICLASTIC SANDSTONE
PELOID WACKESTONE
SILICICLASTIC SANDSTONE
1) PISOLITE-RICH PACKSTONE-GRAINSTONE
2) PELOID-INTRACLAST PACKSTONE-GRAINSTONE
SILICICLASTIC-RICH PELOID WACKESTONE