First-day road-log, Trip 3, from Washington Ranch to Whites City, Walnut Canyon and Carlsbad Cavern

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WALNUT CANYON AND CARLSBAD CAVERN

FIRST-DAY ROAD LOG, TRIP 3, FROM WASHINGTON RANCH TO WHITES CITY, WALNUT CANYON AND CARLSBAD CAVERN

PETER A. SCHOLLE AND DANA S. ULMER-SCHOLLE

Assembly Point: Washington Ranch Road near tufa dam.

Departure Time: 8:00 AM

Distance: 22.3 miles round trip

Four stops (two optional stops)

SUMMARY

Today's trip will look at the Permian geology from the mouth of Walnut Canyon to Carlsbad Caverns that covers the transition from the main Capitan reef to the back reef sediments of the Tansill and Yates Formations.

0.0 Junction with U. S. Highway 62-180 and New Mexico Highway 7 in Whites City.

(7.3 miles from Washington Ranch). Set odometers to zero. Take NM 7 toward Carlsbad Caverns. Drive through Whites City.

0.5

0.5 Stop 1: Capitan reef to back-reef transition at the mouth of Walnut Canyon.

Park vehicles on paved pull-off on the right-hand side of the road at the Carlsbad Caverns National Park entrance sign. This stop is partially on National Park land and partly on private land. Collecting is not permitted within Carlsbad Caverns National Park, and you should get permission to access the eastern part of the outcrop since it is privately owned (if you ever come out here independently). **Since we are going to be on National Park Land for the entire day, please leave your rock hammers in the car/van! To collect samples, you must have a collecting permit from the Park Service.**

The Guadalupe Mountains contain some of the finest outcrops of reef and reef-related rocks in the world. The entire depositional spectrum from far-back-reef to deep basin deposits can be observed in outcrops of the Guadalupe Mountains (Fig. 1.3.1) and adjacent areas, with little or no structural deformation and slight vegetation or soil cover. The reef complex of this region is dissected by a series of deep canyons cut approximately at right angles to the regional facies strike. These canyons, especially McKittrick Canyon, provide exceptional cross-sectional views of the lateral and vertical relations of depositional environments through time.

Finally, the region is rather exceptional in that, at the end of Guadalupian time, the entire suite of facies was preserved (essentially pickled) by extremely rapid deposition of evaporites (gypsum/anhydrite, halite, sylvite, and more exotic salts). These Ochoan evaporites filled the Delaware Basin remnants and even covered adjacent shelfal areas (Fig. 1.3.2). Thus, original facies relations were preserved from extensive erosional modification, and late Tertiary uplift, coupled with dissolution of the soluble evaporites, has led to resurrection of original (Permian) topography, greatly facilitating facies reconstruction.

This locality, at the entrance to Walnut Canyon, provides world-class exposures of Guadalupian reef carbonates of the Capitan Formation. These carbonates were deposited in the warm equatorial seas of the Delaware Basin (Fig. 1.3.3). In this area, the fore-reef facies and part of the reef have been buried beneath the thick Ochoan (and some thin Tertiary-Quaternary) filling of the Delaware Basin (Fig. 1.3.4). The Castile Formation, although completely or partially removed in areas to the southwest, has been preserved in this area because of the northeastward tilting of

FIGURE 1.3.1. Photograph of the north side of McKittrick Canyon showing shelf-to basin sedimentary facies of the late Guadalupian Capitan reef. Strata at top of section at far left are Yates and Tansill back reef carbonates; massive triangular cliff just to the right is last phase of Capitan reef development; strata below and to the right of the reef are steeply-dipping fore-reef talus beds which flatten and pass gradually into the deeper-water sediments of the Delaware basin.

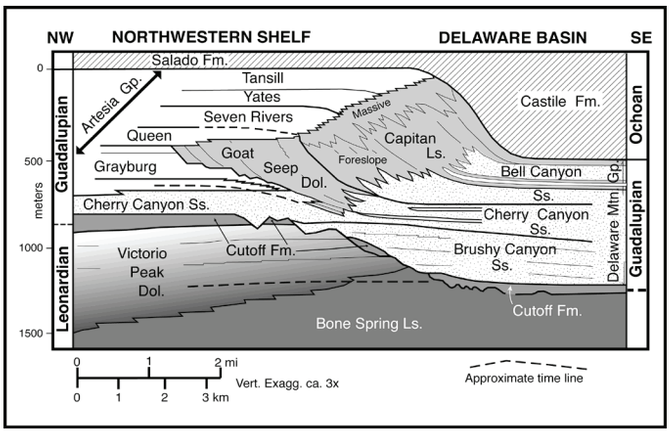

FIGURE 1.3.2. Stratigraphic nomenclature of the Permian strata exposed in the Guadalupe Mountains. Modified from numerous sources including King (1948), Hayes (1964), Tyrrell (1969), and Pray and Esteban (1977).

this region. Thus, only a small exposure of the reef-crest and its transition to the near-back reef are exposed.

We will examine the rock spur between Walnut Canyon and Bat Cave Canyon (across the street from the parking area) and look at how biology, diagenesis and structure have influenced these sediments and their hydrologic characteristics.

The Capitan reef is the main carbonate sediment producing facies and forms a more-or-less continuous, well-defined facies around the Delaware Basin (Fig. 1.3.5). This narrow facies band has broad bends and embayments similar to modern reefs, and it is also the zone of maximum faunal diversity with a complex framework of calcareous sponges, *Tubiphytes*, phylloid algae, *Archeolithoporella* and variety of other organisms (Figs. 1.3.6, 7).

Babcock (1974) noted a distinct faunal zonation within the reef (Fig. 1.3.8). He recognized an *Archaeolithoporella*-nodular

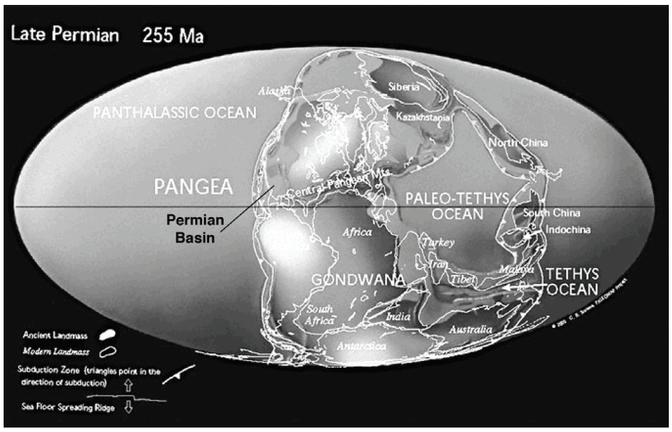

FIGURE 1.3.3. Continental reconstruction during Late Permian (Kazanian) time. In this model, the Permian Basin lay just north of the equator and close to the western margin of Pangea. Approximate present-day outlines are shown for reference (Scotese, C.R., http://www.scotese.com, PALEOMAP website, 2002).

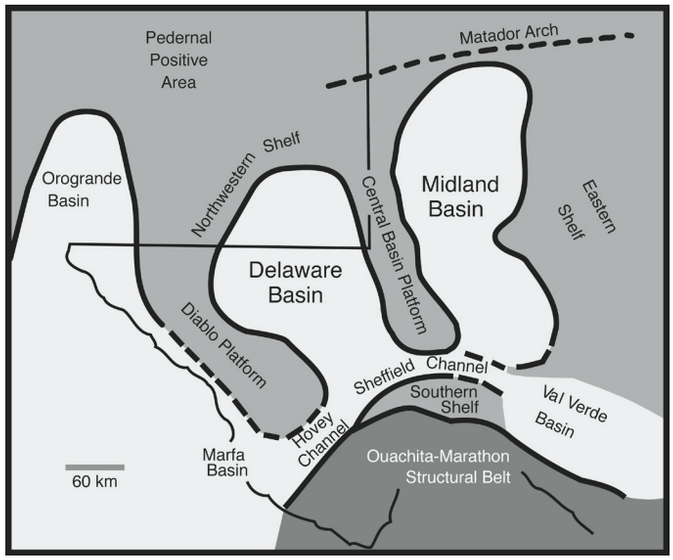

FIGURE 1.3.4. Location of the Delaware and Midland Basins during Leonardian and Guadalupian time. By Capitan deposition, the Midland Basin was filled, and the Delaware Basin and the surrounding platforms were the only active sites of carbonate deposition. Diagram modified from King (1948), McKee et al. (1967) and Williamson (1979).

boundstone, a phylloid algal boundstone, and a *Tubiphytes*-sponge boundstone/packstone as well as some transitional zones (Figs. 1.3.8, 9). In all these facies, there are four important elements: 1) an in situ framework of oriented organisms; 2) encrusting and

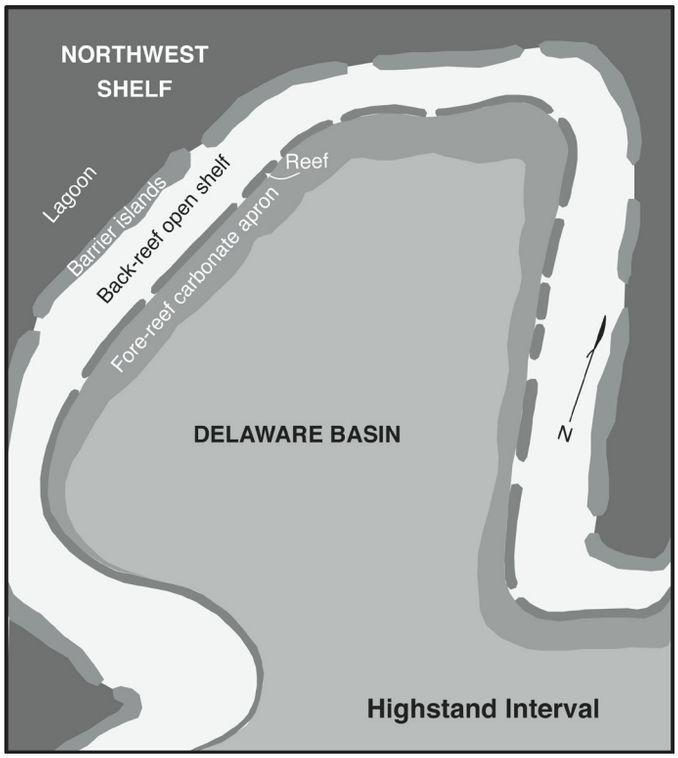

FIGURE 1.3.5. Distribution of facies during highstand Capitan deposition in the Delaware Basin.

FIGURE 1.3.6. A portion of the diorama of the Capitan reef produced by Terry L. Chase and displayed at the Permian Basin Petroleum Museum in Midland, Texas. The artist emphasized the framework sponges and the abundant encrusting fauna.

binding organisms which added stability to the framework; 3) internal sediment of skeletal fragments, pellets, or other grains which lodged in open pores in the framework; and 4) submarine cement crusts filling virtually all remnant porosity.

The dominant framework organisms in this complex are calcareous sponges. Many different types existed here (Figs. 1.3.11, 12), including members of the genera *Guadalupia, Amblysyphonella, Cystaulete, Gigantospongia* and *Cystothalamia*. Some of the sponges grew as upright, solitary organisms, others branched, and still others were platy, growing in cavities, with other species of sponges growing off of them (Wood 1999).

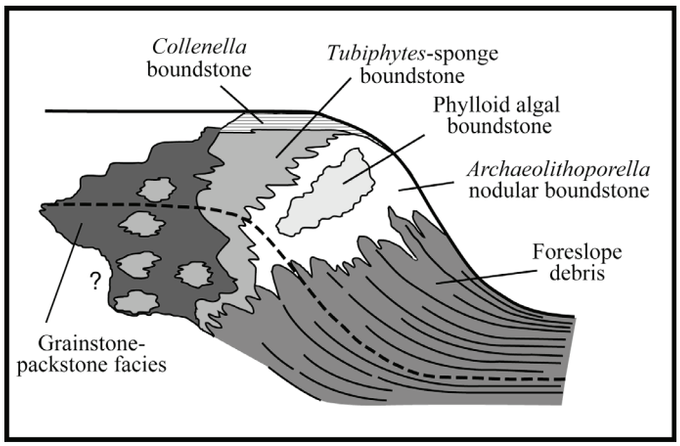

FIGURE 1.3.8. Idealized distribution of microfacies within the Capitan-age reef (adapted from Babcock 1977).

Other organisms such as *Tubiphytes* (of problematic affinities), stromatolitic blue-green algae, phylloid algae (Fig. 1.3.13), and bryozoans also form significant framework elements, at least locally. Solitary rugose corals can also be seen, although they do not form a volumetrically important part of the reef framework. Encrustation and stabilization of this skeletal framework was accomplished by *Archaeolithoporella* (a possible alga; Fig. 1.3.14), *Tubiphytes* (found as both framework and encrusting forms; Fig. 1.3.15), *Solenopora* (a probable red alga), *Collenella* (an algal form; Fig. 1.3.16) and other, less common, organisms. Such encrustation, seen also in modern reefs, probably contributed greatly to the strengthening of the reef framework.

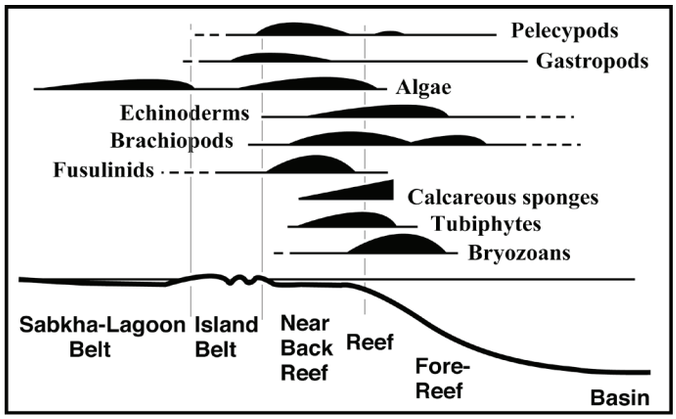

FIGURE 1.3.7. Distribution of the major faunal groups in the Capitan-age strata. Modified from Newell et al. (1953) and Schmidt (1977)

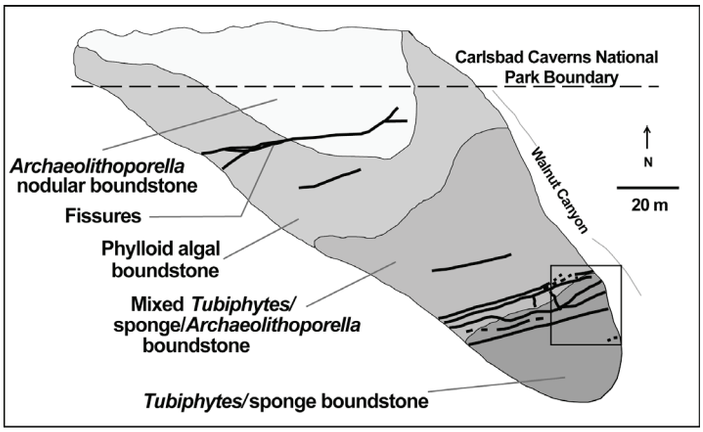

FIGURE 1.3.9. Map of microfacies distribution at the mouth of Walnut Canyon. Rectangle shows the detailed area of mapping in Fig. 10. Modified from Babcock (1974).

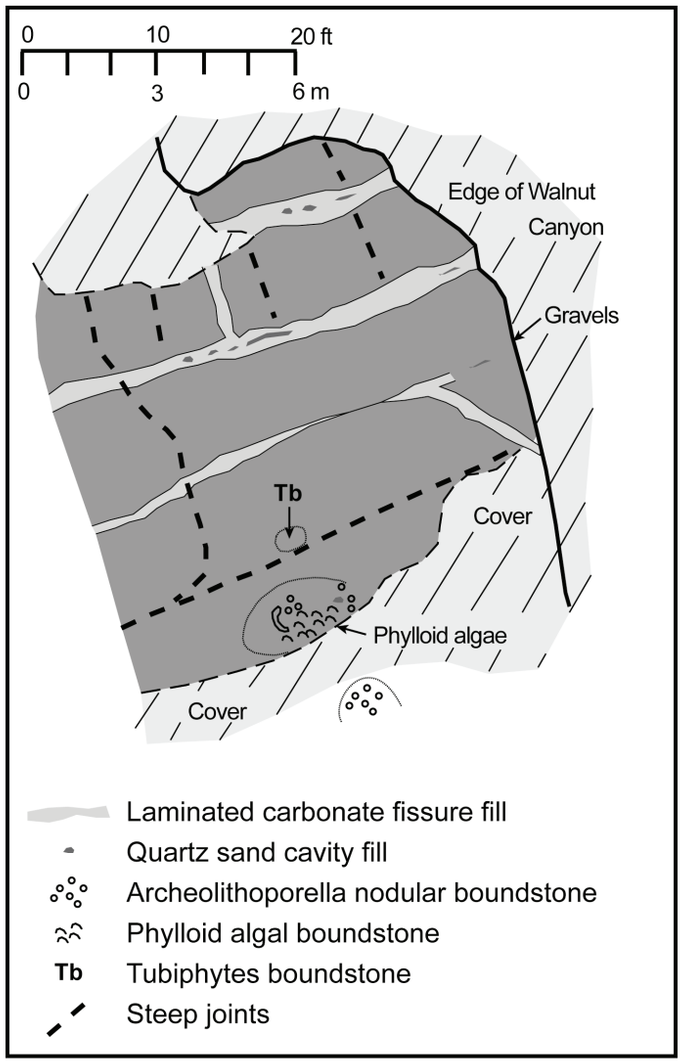

FIGURE 1.3.10. Detailed map of microfacies at the mouth of Walnut Canyon. Modified from Babcock (1974).

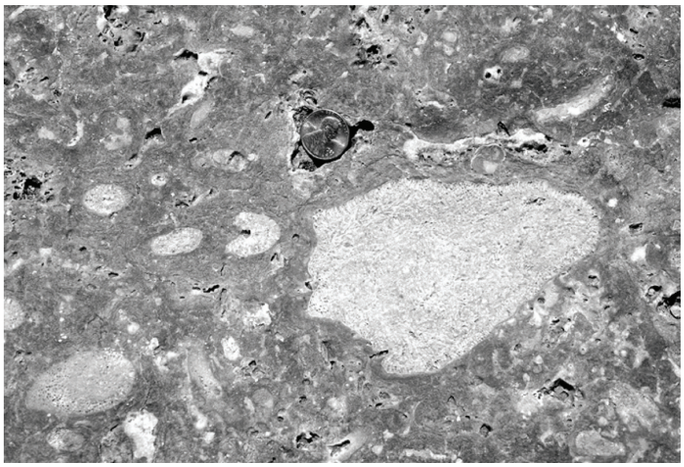

FIGURE 1.3.11. From the mouth of Walnut Canyon, large calcareous sponges (white) still in an oriented living position with long axes perpendicular to reef trend. Calcareous sponges were encrusted by *Archaeolithoporella* and marine cement (gray to black).

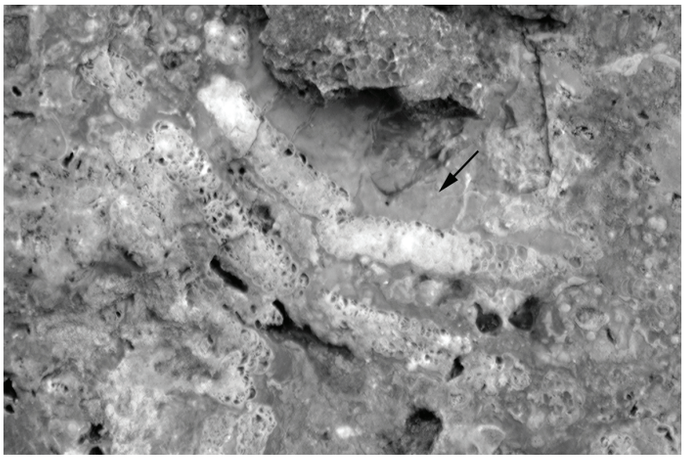

FIGURE 1.3.12. Chambered calcareous sponges (*Gigantospongia disconformia*) at the mouth of Walnut Canyon are another major reef framework organism. The light gray material (arrow) is mud trapped behind the sponge, and the darker gray material is a combination of *Archaeolithoporella* and marine cement.

Submarine cements form a very important component of the Permian reef. Coarse fans of radial-fibrous crystals fill much of the primary porosity in the reef and make up more than half the total volume of rock in many locations (Fig. 1.3.17). The cement fans, probably originally aragonite, are commonly interlayered with *Archaeolithoporella* or other encrusting organisms. The submarine cements are restricted to a relatively narrow zone which extends from one to two hundred meters (300-600 ft) down the fore-reef slope to perhaps 0.8 km (1/2 mi) shelf-ward of the reef crest. Very similar relations have been seen in modern reefs such as in Belize, Florida, the Bahamas, and Jamaica. In these areas, as in the Permian, submarine cementation, largely in the form of aragonitic and high-Mg calcite fans and crusts, are restricted to

FIGURE 1.3.13. Phylloid algal-*Archaeolithoporella*-marine cement facies in the Capitan reef from the spur at the mouth of Walnut Canyon. White, curved strips are oriented phylloid algae; white-gray patches are trapped micrite and darker encrustations consist mainly of *Archaeolithoporella* and marine cement.

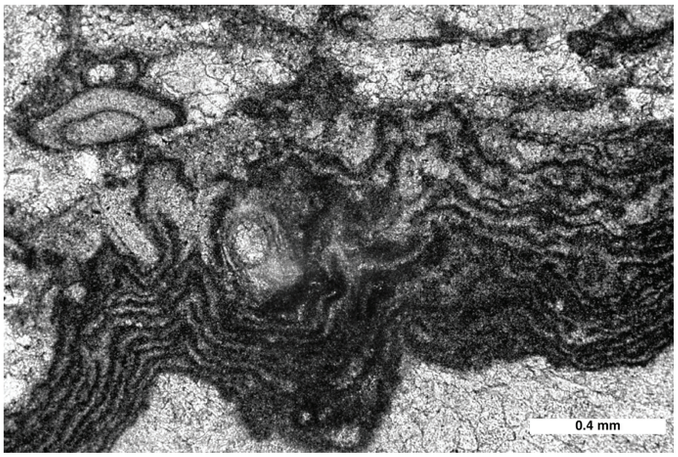

FIGURE 1.3.14. Thin section photomicrograph (plane-polarized light) of *Archaeolithoporella* from Rattlesnake Canyon. *Archaeolithoporella* encrusted organisms and grains within the reef facies of the Permian reef complex and played an important role in lithifying the reef.

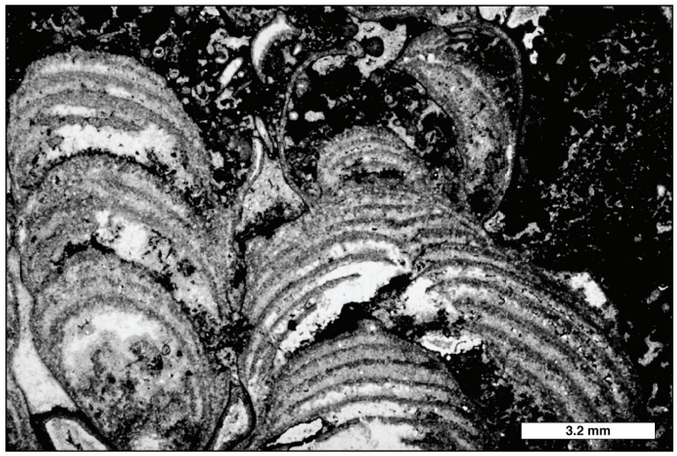

FIGURE 1.3.16. Photomicrograph of *Collenella guadalupensis* forming finger-like or domal skeletal structures that are part of the reef to near-backreef framework. The columns are composed of precipitated, not trapped calcium carbonate. Although viewed as a microbial deposit by some, it has been described as a probable stromatoporoid by others (J. A. Babcock, 2003, written commun.).

the reef face, upper fore-reef and near-back-reef zones (James and Ginsburg 1979, Macintyre 1977, Macintyre 1985). Petrographic evidence has indicated the former existence of cements with both primary aragonite and high-Mg calcite fabrics although both have subsequently been altered to low-Mg calcite (Lohmann and Meyers 1977; Loucks and Folk 1976).

Submarine cements also are seen filling synsedimentary fractures (Fig. 1.3.18). These fractures are shelf-margin-parallel features that probably formed due to compaction of underlying reef rubble below a rigid slab of cemented reef. The fractures were contemporaneous with sedimentation since they are lined with sessile organisms and have extensive marine cementation; local silt infiltration into fractures also argues for an early origin. Several such fractures are well exposed at this locality.

We will now walk up Walnut Canyon examining the transition from reef to near-back-reef sediments, eventually crossing from the south to the north side of Walnut Canyon. On this traverse, be sure to note changes in bedding character as well as sediment composition. You should see reef rubble and/or patch reefs. Also note the remarkably rapidity of the lateral lithologic changes. The most obvious change is from a boundstone fabric to one of grainstones and packstones containing ooids (Fig. 1.3.19) and skeletal grains. Cephalopods, foraminifers, pelecypods, bellerophontid gastropods (Fig. 1.3.20), and most importantly, dasycladacean green algae, particularly *Mizzia* (Fig. 1.3.21), and *Macroporella*,

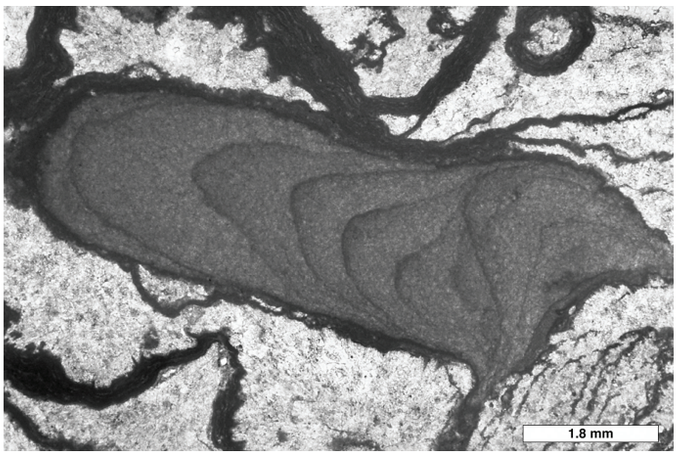

FIGURE 1.3.15. Photomicrograph of a typical Tubiphytes encrustation (in oblique longitudinal section) — here associated with *Archaeolithoporella* and marine cement in a sponge-cement reef. *Tubiphytes* appears to grow extensively on seafloor surfaces, but it may also have been a contributor to cryptic reef communities, growing in dimly lit recesses and cavities in reef and fore-reef settings.

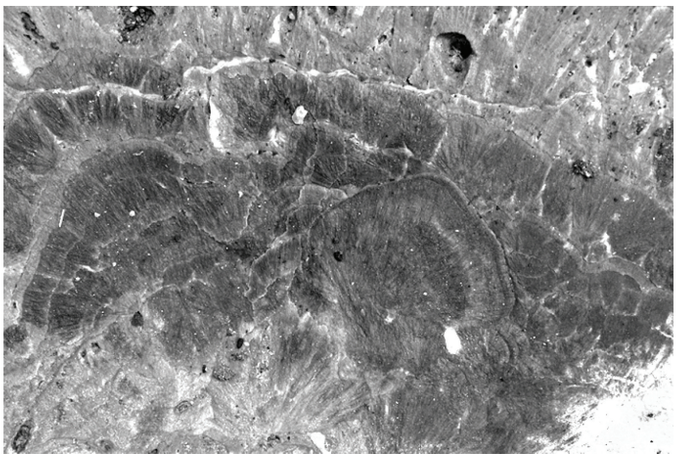

FIGURE 1.3.17. An acid-etched rock surface showing details of radial fibrous cement filling a large primary cavity in the upper Capitan reef framework from Walnut Canyon. Virtually the entire field of view is marine cement that may form as much as 80% of the total rock volume.

FIGURE 1.3.18. A synsedimentary fracture (neptunian dike) in the Capitan reef of Walnut Canyon showing parallel cement linings on opposite walls of fracture seen in an acid-etched outcrop. Latest stage filling in center of fracture is clastic terrigenous silt piped down from an overlying surface of transport. Fractures trend parallel to reef front and extend essentially vertically down through the reef and into upper fore-reef talus facies. These are the cause of the "spurs" seen in reef.

rapidly supplant sponges and bryozoans as the major skeletal components.

Bedding in these well-sorted grainstones is massive and indistinct, but still is far better defined than in the reef facies. Grain size in these near-back-reef deposits ranges from coarse silt to coarse sand; sorting is moderately good to excellent, and coated grains and ooids form a significant percentage of the total sediment.

Sediments further up-canyon (farther back-reef) show increasing amounts of dolomite, fenestral fabrics, coated (pisolitic) grains, algal (?) lamination, carbonate breccias, "tepee structures," and

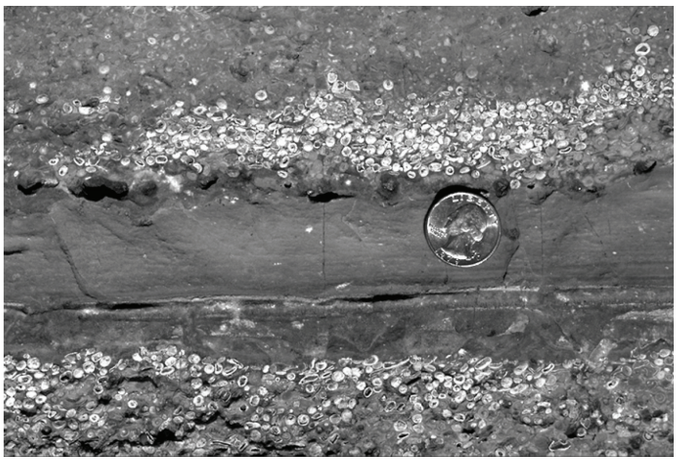

FIGURE 1.3.19. Oolitically coated fusulinid and algal grains in near-back reef strata of Walnut Canyon. Ooids are not common in the Capitan complex, but when they are found, it is most commonly in this setting associated with tidal channels and small islands or bars.

FIGURE 1.3.20. A photograph of large belerophont gastropods from Dark Canyon. Like present-day belerophont gastropods, they were common sediment grazers in the near back-reef environment.

thin, clastic terrigenous sandstone-siltstone units. The abrupt facies transition from reef to back-reef is similar to that seen in many modern settings. In the Florida Keys on the western side of Andros Island in the Bahamas, for example, the change from reefal boundstones to skeletal, back-reef grainstones takes place over distances of just a few tens to hundreds of meters. The near-back-reef areas in Florida and the Bahamas generally consist of complex, small-scale microfacies of green-algal (*Halimeda*) grainstones, grapestones (coated and coalesced grains), ooids, skeletal fragments, and other lithologies. In areas such as the Joulter's Cay region of the Bahamas, one can see these grainstone types closely intermingled as a series of submarine sand waves, islands, tidal channels, and beaches. Associated with these grainstones are mudstone-wackestone microfacies in sheltered areas of tidal flats and back-barrier coves.

The extremely varied lithologies within the Permian near-back-reef settings presumably reflect similarly complex microfacies

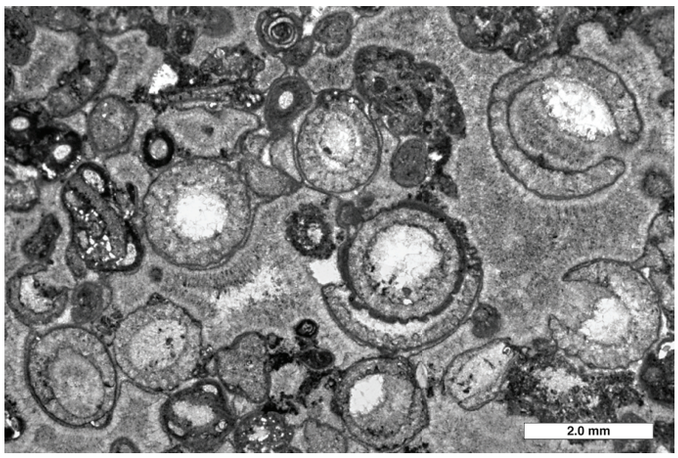

FIGURE 1.3.21. A photomicrograph of a green-algal grainstone, consisting mainly of *Mizzia* sp., with grains encased in typically cloudy (inclusion-rich), penecontemporaneous, isopachous marine cements.

patterns. This is also evident in the intimate mixture of diagenetic patterns in the Permian near-back-reef sediments. Submarine and vadose and phreatic nonmarine cements are all present in local zones in this area, probably as a result of local (island facies) input of nonmarine fluids (Fig. 1.3.22). Selective leaching of originally aragonitic skeletal fragments, probably in a subaerial setting, led to development of substantial, if localized, secondary porosity (Fig. 1.3.23). A major question in recent years has concerned the nature of the reef margin. Did the reef occupy a position as a topographic high point along the shelf margin or did it form in a slightly deeper water setting on the upper slope (with the ooid-*Mizzia* grainstones as the shelf edge "high"). Your observations on this traverse should focus, at least in part, on resolving this question. *0.2*

0.7 Near-back-reef *Mizzia*-dominated grainstones on right. These are equivalent to the outcrop visited on the left side of the canyon, and are also excellent outcrops to visit to view the near back reef sediments. *0.2*

0.9 Cross Walnut Canyon *0.2*

1.1 Pisolite-bearing dolomites and faulted upper Yates sandstones in roadcut. *0.1*

1.2 Upper Yates and lower Tansill sediments in canyon walls. This locality exposes mainly pisolitic dolomites and sandstones and is an excellent area for examining tepee structures. *1.0*

2.2 Chihuahua Desert Exhibit area on right. *0.3*

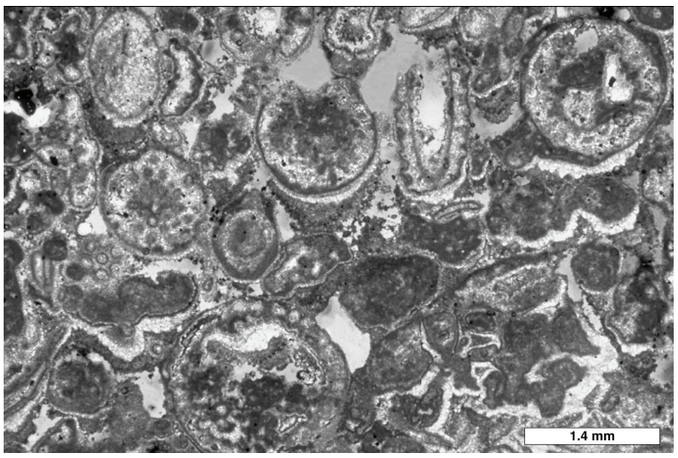

FIGURE 1.3.22. Thin-section photomicrograph (plane-polarized light) of a near-back-reef grainstone in the Tansill Formation approximately 1 km into Walnut Canyon. The predominant grains again are *Mizzia* green algae but the primary porosity has been partially filled with an early diagenetic cement produced in the vadose zone — note slight corrosion of tops of grains and pendants or microstalactites of calcite cement below the grains. Subaerial exposure may have occurred on small islands during sedimentation or in association with a major (early Ochoan) hiatus that followed Tansill deposition.

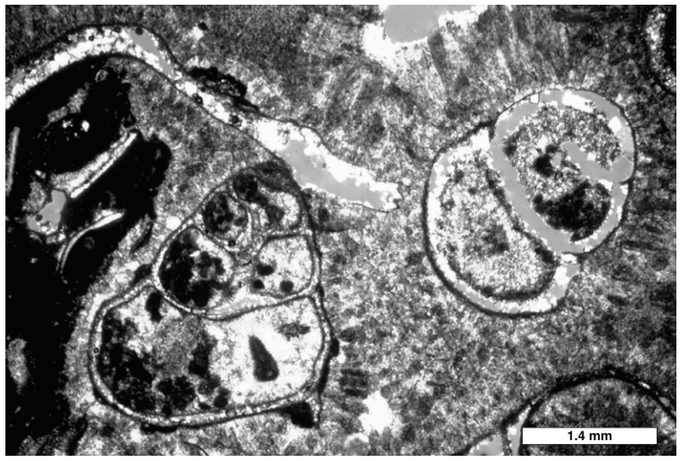

FIGURE 1.3.23. Thin-section photomicrograph (cross-polarized light with gypsum plate) of a near-back-reef grainstone in the Tansill Formation. The predominant grains are gastropods and bivalves that were encrusted penecontemporaneously with inorganic marine cement (cloudy, fibrous to bladed crusts). The sample subsequently was subjected to leaching of the originally aragonitic molluscan grains, presumably as a result of local meteoric exposure.

2.5 Road cuts on right expose Yates Fm. dolomite and sandstone. *1.0*

3.5 Optional Stop 1. Lower Tansill pisolites and Yates Sandstones

Parking area on left with exposures of pisolitic dolomites, tepee structures, and sandstones of Yates Fm. 0.5

4.0 Stop 2. Exhibit area (showing botanical diversity of the area) on the left.

Canyon wall on left has exposures of Yates Fm., including the large, sand-filled cavern described by Dunham (1972, Stop II-5). Please remember that we continue to be in the National Park and collecting of any kind is not permitted. Outstanding exposures of pisolitic dolomites of the upper Yates Fm. (see Pray and Esteban 1977, Dunham 1972, Esteban and Pray 1977) are exposed at this outcrop. The sandstone unit visible in this outcrop is called the Hairpin Sandstone, and it is within the Hairpin Dolomite of the Yates Fm. (Fig. 1.3.24). It is ~ 1 meter-thick, well-sorted, arkosic to subarkosic, fine sandstone to siltstone unit. This unit and the two sandstones at the top of the Yates Formation are remarkably consistent in grain size, thickness, composition and lateral extent on the shelf. Within the sandstones, faint traces of low-angle channel structures, ripples and subhorizontal bedding can be found, but little else in the way of depositional fabrics is visible. The sandstones overlie erosional surfaces on the underlying carbonates. Dunes migrating over an exposed carbonate platform and/or deposition in shallow, lagoonal waters may have produced the observed fabrics.

The concept of reciprocal sedimentation (Fig. 1.3.25) helps to explain a significant problem in Permian basin stratigraphy. The

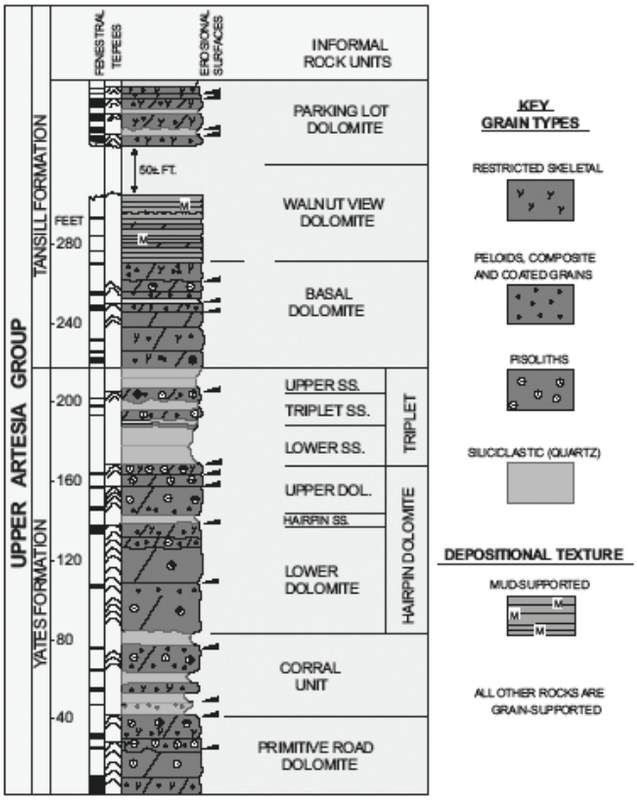

FIGURE 1.3.24. Stratigraphic section of the uppermost Yates Formation at the Hairpin Bend pisolite locality in Walnut Canyon, Carlsbad Caverns National Park, Eddy Co., New Mexico (modified from Pray and Esteban 1977).

Delaware Basin shelf sequences are mainly carbonate, but the basinal deposits are overwhelmingly siliciclastic. With reciprocal sedimentation, during periods of high sea-level stands, carbonate sedimentation predominates with reefs and grainstone shoals on the shelf and thin carbonates turbidites in the basin. Clastic sediments would have been trapped in vegetated dunes or in shoreline deposits further back on the shelf. During periods of low sea-level stands, the carbonate factory would shut down and eolian and fluvial sands and silts would move across the shelf and into the basin forming thick, basinal sandstones and siltstones. The Yates sandstones are the shelfal remnants of the "seas" of sand.

This locality illustrates numerous cycles of pisolitic, tepee-bearing sediments (termed "Walnutite cycles" by Pray and Esteban 1977). The main small-scale features to be seen at this outcrop are the abundant pisoliths that range from B-B-size to golf ball size. They have concentric laminations of thin carbonate coatings around nuclei of fractured pisoliths or, rarely, marine fossils (Fig. 1.3.26). The pisoliths, which have been completely replaced by aphanocrystalline dolomite, occur in cyclic beds, commonly with reverse grading (Fig. 1.3.27). In some (but not most) cases, pisoliths have intergrown or interlocking textures (Fig. 1.3.28).

There is considerable evidence to show that these pisoliths were originally aragonite in composition, but now replaced by

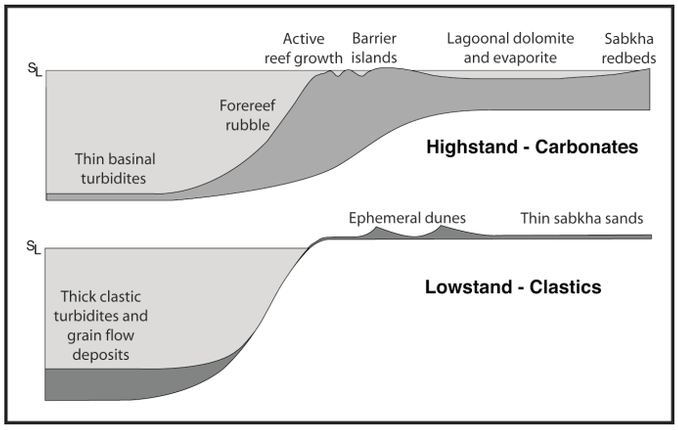

FIGURE 1.3.25. Relative high-stands of sea level resulted in extensive carbonate productivity on the flooded shelf, especially at its margin. Clastic terrigenous sediments are trapped in inland sabkha and dune belts, and basinal sedimentation is restricted to thin units of shelf-derived carbonate debris. During low-stands of sea level, carbonate sedimentation is minimal, and clastic terrigenous sand dunes migrate across shelves and spill into basinal areas where they are deposited as thick turbidite and grainflow deposits. Shelf sedimentation is restricted to sabkhas that trap only a fraction of the dune sands which traverse the region producing thin, sheet-like, shelf sandstones.

dolomite (Fig. 1.3.29). The pisoliths are associated with sheet cracks—broad bands of displacive, fibrous carbonate, presumably also originally aragonite (Loucks and Folk 1976) but now dolomite or calcite (Fig. 1.3.30). These displacive crusts are related to the origin of the tepee structures of this area for the tepees are expansion polygons formed by a volume increase of the associated sediments. This was most likely accomplished by in situ, near surface, displacive growth of aragonite and (or) evaporite minerals.

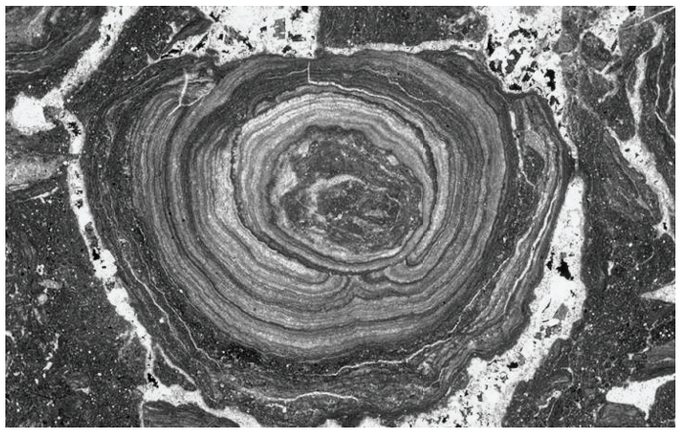

FIGURE 1.3.26. Thin-section photomicrograph (cross-polarized light) of a pisoid from the Yates Formation at 1708.1 ft depth in Gulf/Chevron PDB-04 well on Northwestern Shelf of Delaware basin. Note irregular, lumpy, partially concentric coatings; fracturing ("autobrecciation") of micritic-peloidal matrix; and evaporite plugging of remnant intergranular porosity.

FIGURE 1.3.27. Typical lenticular deposit of pisolitic dolomite from the uppermost Yates Formation at Stop 2. Note reverse grading of coated grains. Pisoids have been attributed to algal growth, caliche formation, marine spray-zone precipitation, and seepage-spring development.

The origin of pisoliths and tepee structures in these sediments has been the subject of numerous studies and considerable controversy. Extensive discussions of these problems have been presented by Dunham (1972), Esteban and Pray (1977), Pray and Esteban (1977), and Handford et al. (1984) and so will be only briefly outlined here.

Basically there are three hypotheses: 1) the "all wet" model which proposes that the pisoliths were formed by organic (algal) or inorganic coating of grains in a shallow water shelf setting with each grain acting as a free, clastic particle; 2) the "caliche" hypothesis which suggests that pisoliths formed in situ as part of cyclic, reverse graded, caliche which formed by alteration of carbonate sediment brought into the area by storms or other episodic processes; and 3) salina "seepage" model which proposes that ocean water seepage through permeable barriers into sub-sealevel salinas can produce tepees and pisoliths by evaporation and precipitation. Advocates of any of these models can point to modern analogs (mainly from Persian Gulf, Red Sea, or Australian areas) with scattered, small-scale accumulations of aragonitic pisoliths in marginal marine, hypersaline settings. Yet nowhere have we discovered an analog that comes close to modeling the breadth and abundance of pisoliths that one sees in the Permian Basin deposits.

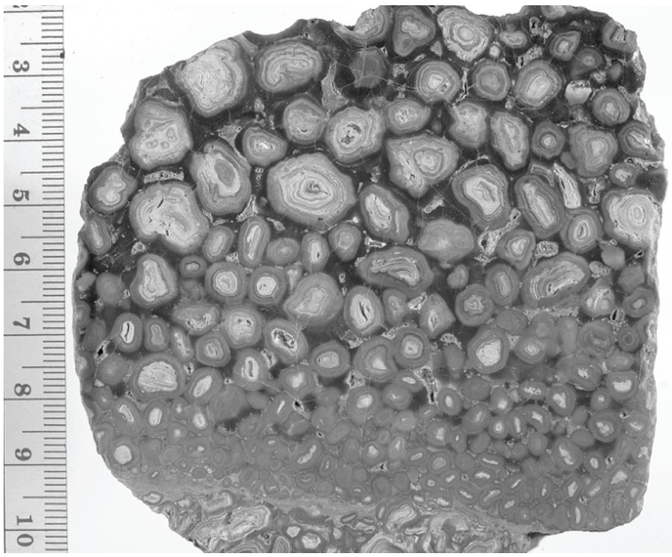

FIGURE 1.3.28. Polished slab of pisolitic dolomite from the uppermost Yates Formation at Stop 2. Note reverse grading of grains, and "fitted fabric" in which grains have interlocked boundaries produced by compromise growth of outer coatings.

FIGURE 1.3.29. Thin-section photomicrograph (plane-polarized light) of the transition from pisoid to botryoidal crust is from the Gulf/Chevron PDB-04 well on Northwestern Shelf of Delaware basin, 30 km ENE of Carlsbad. Note that the extremely elongate rays of cement that extend from upper surface of pisoid show squared crystal terminations. This has been used as evidence for an originally aragonitic composition for the cement (Loucks and Folk 1976).

The differences of interpretation of these deposits, although important from the point of view of fully understanding the rocks, are not of great significance to the petroleum exploration-

FIGURE 1.3.30. Transition from growth of free pisoids to a botryoidal cement crust in an outcrop of the uppermost Yates Formation. Last stages of pisoid precipitation occurred on the upper growth surfaces which eventually merged to form a crust of originally aragonitic cement.

ist. There can be little argument that this facies must have stood close to a paleotopographic high point in Guadalupian time. The persistence of this facies in space and time (it is present in Grayburg, Queen, Seven Rivers, Yates, and Tansill rocks), its consistent geometry (an elongate facies, parallel to the reef trend), and its equally consistent juxtaposition between open marine (grainstones with a high faunal diversity) and restricted (hypersaline mudstones and evaporites) environments indicate that the pisolite facies must either itself have been a major hydrographic barrier or it must have formed just landward of such a barrier. Nowhere in the world today are evaporitic mudstones and open marine, faunally diverse sediments in such close proximity without having an intervening barrier. It seems likely that to act as such a barrier, at least a narrow strip of land would have had to be subaerially exposed (except for tidal channels). This scenario would favor either the caliche or salina seep interpretations. Further support for the salina model may come from isopach studies of the thin sandstone/siltstone beds which are interspersed with the tepee-pisolith beds. Candelaria (1989) showed that these beds did not thin over the pisolite facies but did thin rapidly over the marine grainstones which lay just seaward of the pisolite facies. Although such a relationship may be due to reworking of shelf-margin sands by transgressive seas, it may also indicate depositional thinning over a topographic high in that area.

At this time, then, it seems most likely that the facies just seaward of the pisolites formed an elongate, irregular ridge of low-relief islands, tidal flats, and dunes (Fig. 1.3.31) which allowed marine water seepage into the back barrier lagoon. Such seepage zones saw massive precipitation of aragonite cements and formation of pisoliths as in the modern Lake McLeod and Yorke Peninsula examples (Handford et al. 1984, Lock and Burne 1986, Logan 1987).

Finally, it is possible that a combination of processes could have been involved in the formation of pisoliths. A number of different types of pisoliths can be seen in the Permian strata. These range from the small, irregularly coated grains (which almost

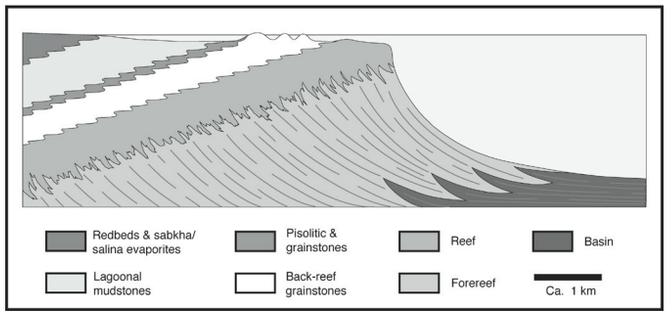

FIGURE 1.3.31. Shelf-to-basin spectrum of microfacies and interpreted depositional environments for the Capitan and Capitan-equivalent strata of the Guadalupe Mountains. Vertical axis is approximately 0.5 km.

certainly formed in a marine setting) to the larger, smoother, and more extensively encrusted grains present at this locality. Thus, a number of different origins can be envisioned for the various pisolith types.The tepee structures and sheet cracks found in association with pisolitic sediments can also be interpreted as either marine or nonmarine. Displacive aragonite crusts and tepees have been noted in submarine cemented areas within the Persian Gulf itself as well as in coastal caliches and sabkha surfaces of the surrounding, subaerially-exposed coastlines (Kendall 1969, Shinn 1969, Warren 1983). Return to vehicles and proceed to Carlsbad Caverns. *1.25*

5.25 Sharp bend in road; primitive road on right can be used as parking area to view exposures of Yates Fm. just ahead. *.05*

5.3 OPTIONAL STOP 2. Hairpin Turn pisolite locality.

Due to the heavy traffic coming and going from Carlsbad Caverns and the limited parking, we may not be able to stop at this locality. With smaller groups, this is an excellent stop to observe pisolites, tepee structures and the Yates sandstones. *.3*

5.6 Exhibit area on left; the thin sandstone-siltstone unit that marks the Tansill-Yates contact is exposed on the left. The road ascends into Tansill Formation dolomites. *1.9*

7.5 Stop 3. Tepee structures and pisolites.

The outcrops at the southwest end of the parking lot provide excellent exposures of tepee structures, sheet cracks, and pisolitic sediments of the Tansill Formation. Use the description from Stop 2 concerning the origin of tepee structures (Fig. 1.3.32) and pisolites to guide your observations. *0*

7.5 Stop 4. Carlsbad Caverns National Park.

Please remember that you are still in the National Park and collecting of any kind is not permitted. We will do the complete walking tour of Carlsbad Caverns (Fig. 1.3.33). The article by Mike Queen in this guidebook is a detailed walking tour of the cave.

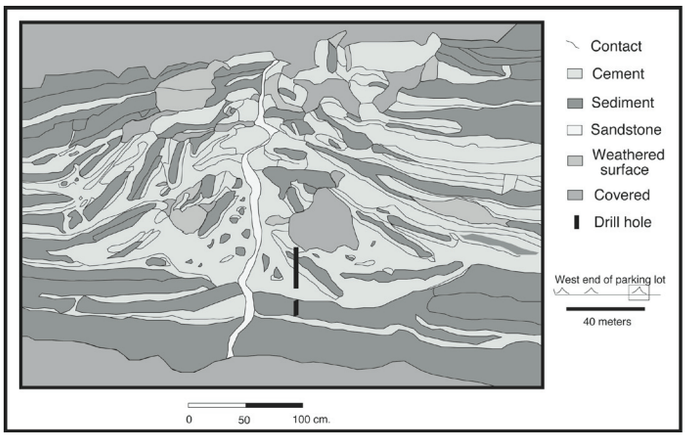

FIGURE 1.3.32. A diagrammatic representation of the tepee structure in Tansill Formation at the parking lot of Carlsbad Caverns.

The cave is the largest (but by no means the longest) cavern system in the world and has spectacular speleothems (Fig. 1.3.34). The cave is developed primarily in the fractured reef and fore reef Capitan Limestone, but the entrance (Fig. 1.3.35) and all of the upper level are in the back-reef dolomites of the Tansill and Yates Formations. The lowest parts of the cave extend down to a level approximately 260 m (850 ft) below the entrance. These deep parts of the cave are cut in the lower part of the reef as well as steeply-dipping fore-reef talus of the Capitan Formation. You should be able to see this fore-reef bedding on your circuit of the Big Room near the "Bottomless Pit" (or look back down the trail when you see the sign for "Bat Guano"). This deep level of dissolution is presumably related to the regional groundwater discharge surface in the Pecos valley to the northeast.

The history of development of the cave is extensively described by Jagnow (1979), Hill (1987, 1990, 1993, 1999) and DuChene and McLean (1989). The location and orientation of the Capitan reef and its early fracture system have controlled, to a large degree, the geometry of the local cave systems. Pliocene-Pleistocene and earlier uplift allowed percolation of phreatic groundwater through the joint system and eventual excavation of the caverns. H_2S, from the bacterial reduction of Permian sulfates in the presence of hydrocarbons, is common in groundwaters of the area and likely was important in forming sulfuric acid which dissolved many of the caves in this region.

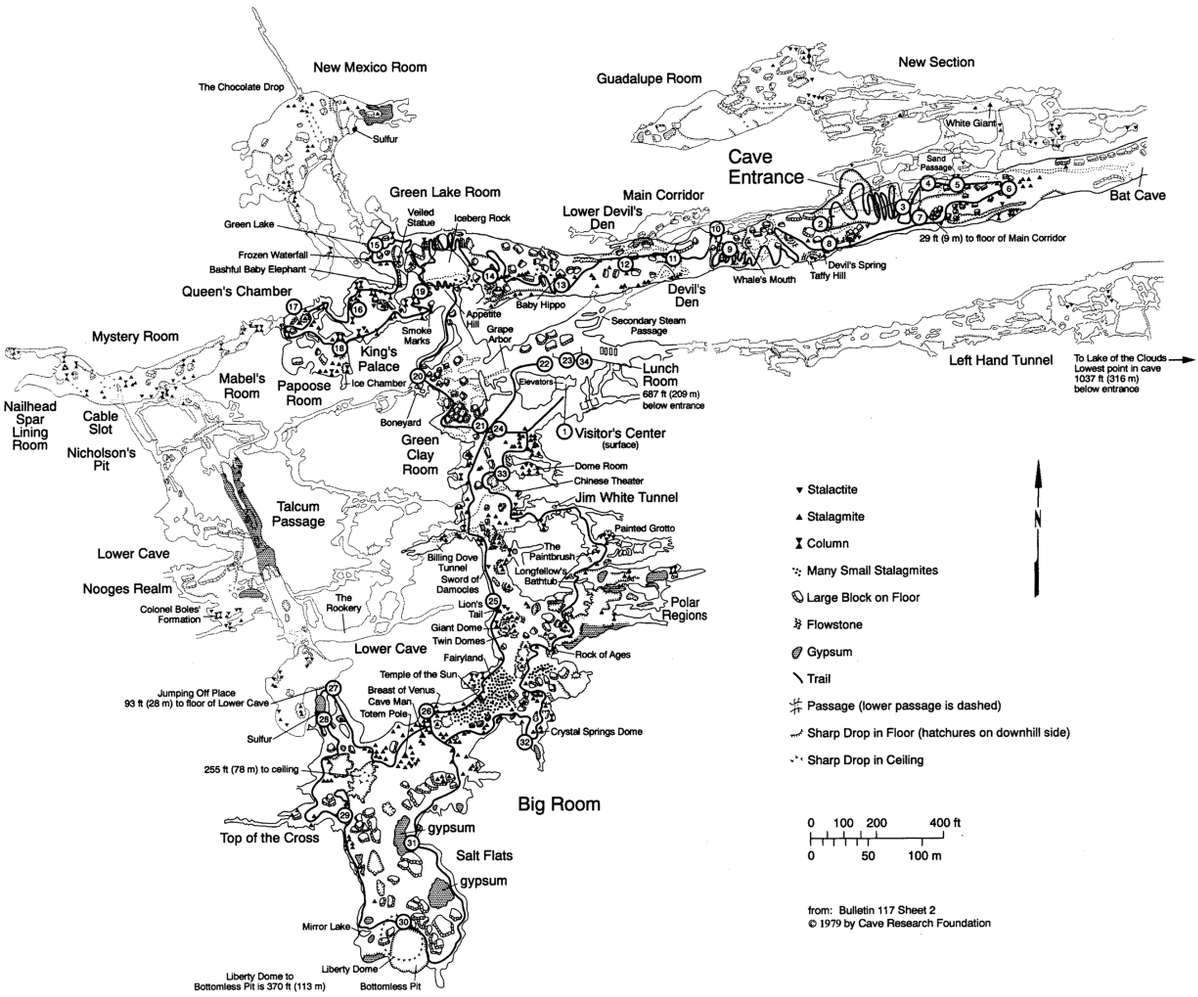

FIGURE 1.3.33. A walking map of Carlsbad Caverns from Hill (1993).

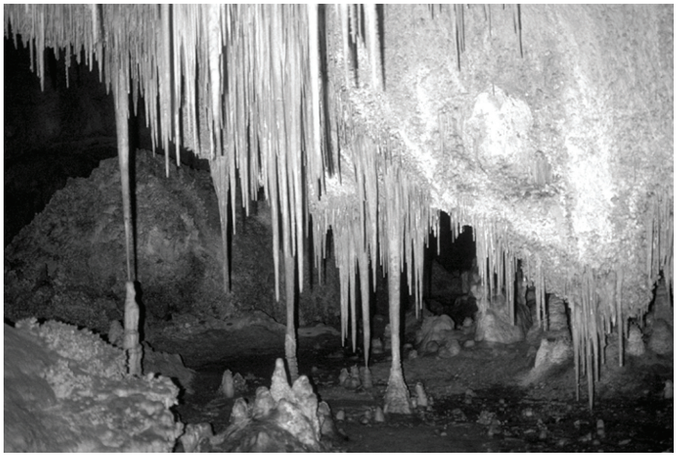

FIGURE 1.3.34. Stalactites, stalagmites, and "soda straws" that have precipitated within Carlsbad Caverns.

For an alternative view on sulfuric acid speleogenesis, see Brown minipaper, p. 36

The subsequent vadose history of the cave led to introduction (and later partial removal) of clay, silt, sand, and gypsum fills as well as calcitic (and even dolomitic) speleothems. Convective air circulation (Queen 1981, Hill 1987) may be responsible for much of the irregular distribution of precipitated calcite "popcorn" on the walls and formations of the cave (such as the Lion's Tail). With the exception of air circulation, the cave is largely inactive today except for some areas in the lowest cave levels.

FIGURE 1.3.35. A view back towards the natural entrance of Carlsbad Caverns.

See Melim et al. minipaper for discussion of cave pool deposits in Lower Cave, p. 38.

See Northup minipaper on cooperative cave research between biologists and geologists, p. 41

End of Day 1 - Trip 3
Return to vehicles and Washington Ranch.

UNRESOLVED PROBLEMS WITH SULFATE SPELEOGENESIS OF CARLSBAD CAVERN

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Over the last 30 years, sulfuric acid speleogenesis has become recognized as the major process forming many Guadalupe Mountains caves (Jagnow et al. 2000). Sulfur isotopes clearly demonstrate that much of the cave gypsum sulfate is derived from microbially reduced sulfide oxidized to sulfuric acid in a near-surface environment (Kirkland 1982). The most likely origin of this sulfate is in-situ oxidation of H_2S to sulfuric acid, dissolution of limestone to form the caves, and precipitation of calcium sulfate (Hill 1987). Cave morphology is consistent with sulfuric acid dissolution or mixing-zone dissolution (Palmer and Palmer 2000).

Despite the geochemical and morphological evidence for sulfuric acid speleogenesis, the exact process remains unclear, so variations of the sulfuric acid model have been proposed (Jagnow et al. 2000; Palmer and Palmer 2000). Also, chronology postulated with the early versions of the sulfuric acid model (Hill 1987) has not been supported by subsequent dating of cave filling events (Lundberg et al 2000; Polyak et al. 1998). The theory has had poor predictive value up to now, a sign that the theory is not yet mature. This minipaper briefly addresses some of the unresolved issues with application of the sulfuric acid speleogenesis model to Guadalupe Mountain caves.

H_2S source: The most significant problem is the source of H_2S. For sulfuric acid speleogenesis to be effective, oxidation of sulfide to sulfate must occur at the site of dissolution. In petroleum terminology, H_2S must migrate to the Guadalupe Mountains Capitan reservoir as either gas or dissolved in water. Petroleum-associated H_2S gas sources are not readily available. Of the 1011 valid Permian Basin gas analyses in the USBM gas analysis database, only 41 (4%) have H_2S concentrations greater than 0.1 %, and the maximum measured H_2S concentration is less than 5%. Gases from petroleum fields will have more hydrocarbon than H_2S, so the main result of gas leakage into the aquifer would be sulfate reduction, not sulfide oxidation.

Microbial H2S gas may also be sourced from the Ochoan sulfates (Hill 1987), but this H_2S gas is not in a position to charge the Capitan Formation. Most H_2S gas from the Castile Formation is likely to be vented to the atmosphere, because dissolution associated with sulfate reduction brecciates evaporite seals. Gas does not migrate downward due to its buoyancy in water, so H_2S

gas could not migrate into the Bell Canyon Formation from the Castile. The Bell Canyon Formation is an ineffective lateral gas migration pathway, anyway. It has very low permeability, and the few permeable pathways are oriented approximately north-south, a direction not favorable for charge to the Guadalupe Mountains.

H_2S is more likely to charge the Capitan aquifer as a dissolved component in moving water (Palmer and Palmer 2000). The H2S source still remains a problem. Dissolved H2S is associated with oil accumulations on the Northern Shelf and Central Basin Platform, but neither of these areas are located where water will migrate to the Guadalupe Mountains. Before development, water in the Capitan aquifer always flowed from the Guadalupe Mountains towards the basin along depositional strike (Hiss, 1980).

If dissolved H_2S is transported to the central Guadalupe Mountains, it must come from an upflow direction in the aquifer; that is, H_2S must be sourced from the west or southwest. The best potential H_2S source in this area is gypsum in backreef shelf carbonates. Backreef evaporite dissolution beds locally have higher heads than adjacent reef as well as high permeability. Gypsum was present in the past, and dissolved or particulate organic matter in water could have reduced this sulfate to sulfide.

Cavern paragenesis. Precipitation of massive gypsum is distinctly later than dissolution of the caves which it fills (Hill 1987). If isotopically light gypsum is not temporally associated with the main phase of speleogenesis, there is no compelling evidence that the main phase of cave formation was by sulfuric acid.

Gravels and silt deposition follows cave dissolution in Carlsbad Cavern, and gypsum was precipitated after silt deposition. The gypsum is widely distributed in the Big Room, so most of the Big Room floor was excavated prior to gypsum precipitation. Room excavation and gypsum precipitation appear to be two events separated by significant time. If sulfuric acid speleogenesis was responsible for cavern dissolution, the sulfuric acid and dissolved sulfate concentration were sufficiently low to prevent concurrent gypsum precipitation. It is unclear why the massive gypsum precipitation in Carlsbad Cavern is not associated with significant dissolution if it is a byproduct of sulfuric acid reaction with limestone. Palmer and Palmer (2000) suggest that much of the Carlsbad Cavern massive gypsum could be the product of sulfuric acid condensation-corrosion. This is possible, but gypsum deposition is followed by breakdown breccia, suggesting that vadose conditions did not develop locally until after massive gypsum precipitation (Hill 1987). Gypsum preservation requires that phreatic water filling the lower parts of the cave be saturated with gypsum.

The distribution of preserved massive gypsum in Carlsbad Cavern also indicates that its precipitation may occur by a process independent from formation of the cave. If massive gypsum is a byproduct of the main phase of cave formation, it should be distributed at all cave levels. Most massive gypsum in Carlsbad Cavern occurs in a relatively narrow depth range (3650 – 3750 ft; Hill 1987). Both age difference and depth distribution may reflect gypsum precipitation related to a mixing zone, as proposed by Queen (1994).

Isotopic data. Sulfur in Carlsbad Cavern massive gypsum is very light (near –20 per mil CDT; Kirkland 1982, Hill 1987). Permian sedimentary sulfate has sulfur with about +11 per mil CDT (Claypool et al. 1980). Assuming sulfur in the massive gypsum was derived by reduction of sedimentary sulfate, the fractionation is about –30 per mil, close to the maximum fractionation by single-cycle microbial sulfate reduction observed in nature (Canfield 2001). In other words, the Carlsbad Cavern massive gypsum is derived mainly from microbially reduced sulfur with at most 25% contribution from sedimentary sulfate heavy sulfur. This is not true for massive gypsum in other caves. The analyzed gypsum block from Cottonwood Cave (+5 CDT, Hill 1987) could be as much as 85% sedimentary sulfate, for example.

The light sulfur in Carlsbad Cavern massive gypsum significantly constrains porewater composition. The pore water precipitating isotopically light gypsum must have a high ratio of dissolved light H_2S to dissolved sedimentary sulfate, or the dissolved sulfate in the pore water must already be isotopically light. The only way to create water with low dissolved sulfate and high concentration of isotopically light H_2S is to introduce H_2S without water dissolving sedimentary sulfate. The easiest way to do this is by H_2S vapor dissolving into low sulfate water. This could happen by condensation-corrosion in the cave. It could also happen by rapid infiltration of rainwater through an H_2S-rich soil zone during aquifer recharge.

Brines derived from the deep Capitan aquifer east of the Pecos River are likely to be nearly gypsum saturated, because the formation contains abundant sulfate cement (Garber et al. 1989). This further eliminates an H_2S source from this area . In situ reduction of dissolved sulfate will not change the bulk sulfur isotopic composition. Rayleigh (closed system) fractionation would shift the H_2S towards the heavy sedimentary sulfate value. The light sulfur indicates that sulfate reduction occurred in an open system.

Discussion. If the sulfuric acid model is valid, it must have the following characteristics. (1) The H_2S was almost certainly introduced as a dissolved component in water, not gas. (2) Hydrological framework constrains the water source (and dissolved H_2S source) to locations west of the Guadalupe Mountains Capitan aquifer. (3) Dissolved H_2S concentration is low; otherwise, most cavern dissolution would be associated with concurrent gypsum precipitation. (4) Porewater sulfate had little contribution from sedimentary sulfate relative to H_2S. (5) Porewater probably derived its H_2S by interacting with a vapor phase in its source area. This is consistent with a vadose zone setting.

A case could even be made that most Carlsbad Cavern dissolution is not related to sulfuric acid. Sulfuric acid did modify the cavern, but it followed the main stage of cave dissolution and was volumetrically less significant. The only unique identifying feature for significant sulfuric acid speleogenesis is the light sulfur isotopes in massive gypsum, but this gypsum follows the main stage of dissolution in Carlsbad Cavern. Morphological features indicative of sulfuric acid speleogenesis could as easily indicate mixing zone origin. Acid-stable minerals and sulfur are clearly related to sulfuric acid speleogenesis, but most examples are in

settings where paragenetic relations are equivocal. This minimal sulfuric acid interpretation could probably be negated with more detailed study of timing of various cave fills and more sulfur isotopic analyses. The purpose for proposing this interpretation is to demonstrate how fragile the current model is.

The most likely scenario is hinted at by Queen (1994) and Palmer and Palmer (2000). Both papers emphasized the mixing zone morphology of the caverns and both note that H_2S oxidation is most likely a product of water mixing. The two water end members could be (1) relatively oxygenated low pCO_2 water saturated with calcite flowing down plunge in the Capitan aquifer and (2) H_2S-bearing water with high pCO_2 saturated with dolomite and calcite flowing down dip from the carbonate-gypsum transitional facies weathering in the vadose zone. As Palmer and Palmer (2000) note, it may be difficult to separate the effects of sulfuric and carbonate dissolution in a mixing zone setting where CO_2 generated by carbonate dissolution is not allowed to escape. Even in the absence of significant H_2S oxidation, mixing of such waters may enhance dissolution, so that the sulfuric acid component of dissolution may be subordinate to that resulting from mixing.

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THE UNKNOWN CRUST BENEATH YOUR FEET: CAVE POOL PRECIPITATES OF LOWER CAVE, CARLSBAD CAVERN, NEW MEXICO

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Introduction

Carlsbad Cavern is the best known of the many caves in Carlsbad Caverns National Park. Although most famous for the extremely large rooms and abundant decorations, Carlsbad Cavern also has a large number of cave pools, both modern and ancient (paleo-pools, now dry). We have started a systematic effort to describe cave pools in the Lower Cave section of Carlsbad Cavern, with particular focus on the minerals lining the pool bottoms and sides. We report here the results of the initial field studies; additional detailed petrographic and chemical studies are still in progress. Interpretation of these features awaits additional data.

Study Area: Lower Cave, Carlsbad Cavern

The Lower Cave area of Carlsbad Cavern is a large area of connected rooms sitting under the more famous Big Room. Access is through a pit with a ladder from the Big Room. The National Park Service runs guided tours of Lower Cave along a well marked trail making a loop past many of the pools, both wet and dry. Less traveled trails (but still marked) provide access to other pool areas. All pools are perched as the current water table is several hundred feet below the level of Lower Cave.

Methods

A total of 23 pools in 17 different areas were described and measured in detail. Proper cave conservation etiquette was followed to avoid any damage to pools or surrounding cave areas. A standard form was developed to ensure consistency between different workers. The form includes a check list to record dimensions of frequently observed features (e.g., pool area and depth, pool spar, etc.) and additional space for unusual or unique features. A scale plan drawing and cross-sections were done of each pool. Detailed observations were tied to the drawings. Associated speleothems such as overhanging stalactites or soda straws and adjacent flowstone were also noted. Most pools were also documented with digital photographs ranging from room shots

for context to detailed close-ups of individual features. Care was taken to include scale in most photographs, usually a metric tape measure.

Results

Detailed descriptions of 23 pools and reconnaissance observations on several more allow a general characterization of cave pools in Lower Cave. Some pool precipitates are present in nearly all pools, others are restricted to a few or even a single pool. The variability is quite high, often even for adjacent pools where one would expect similar conditions to apply. Seven of the pools currently have some water, but only three of these are full and possibily actively precipitating minerals. Anecdotal evidence suggests more pools were wet prior to drilling of the elevator shaft but no records were found for specific Lower Cave pools. A uniform black coating covers pools and walls alike in several areas. Since it is uniform, both surfaces had to have been dry during its formation indicating these pools, at least, have not been active for some time . Most of the pools described are less than 3 m in diameter but this is an artifact; smaller pools are easier to describe and many larger pools remain to be described.

Pool spar

Euhedral calcite crystals growing in pools are called pool spar (Hill and Forti 1997). All but one of the pools in Lower Cave are completely coated in a relatively uniform layer of pool spar (Figure 1.3.36). Pool spar usually defines the paleo-water line in dry pools as its sharp surface morphology contrasts strongly with the smooth surface of flowstone. The pool spar is clear to light brown but often appears darker due to very thin coatings of darker material, almost black in some areas. Besides the pool walls and floor, pool spar also coats sunken cave rafts and broken soda straws or stalactites, essentially anything sitting in the pool water for some time. Although bat bones are found on top of pool spar, we did not observe any entombed examples.

Pool spar is composed of very small (<1 mm) euhedral crystals of calcite that group together as round knobs (1-4 cm). These knobs coalesce into larger bulbous clumps (5-10 cm) which in turn sometimes group together to form 10-30 cm clusters (Figure 1.3.36b). All the pool spar in a single pool is usually similar in size; the range represents variability between pools. Internally, the individual crystals and knobs are not visible; rather they coalesce to form a radiating, fibrous fabric with sweeping extinction throughout each clump.

Shelfstone

Shelfstone grows from the edge of a pool along the water-air interface (Hill and Forti 1997). Twelve pools have shelfstone, most commonly along only part of the pool edge. Steeper pool edges are more likely to develop shelfstone but this rule is by no means universal. The upper surface of shelfstone is usually smooth and often coated in flowstone or dripstone. The undersurface is either knobby, like pool spar, or is covered in pool fingers or chenille spar (see below). In one case, there are multiple levels

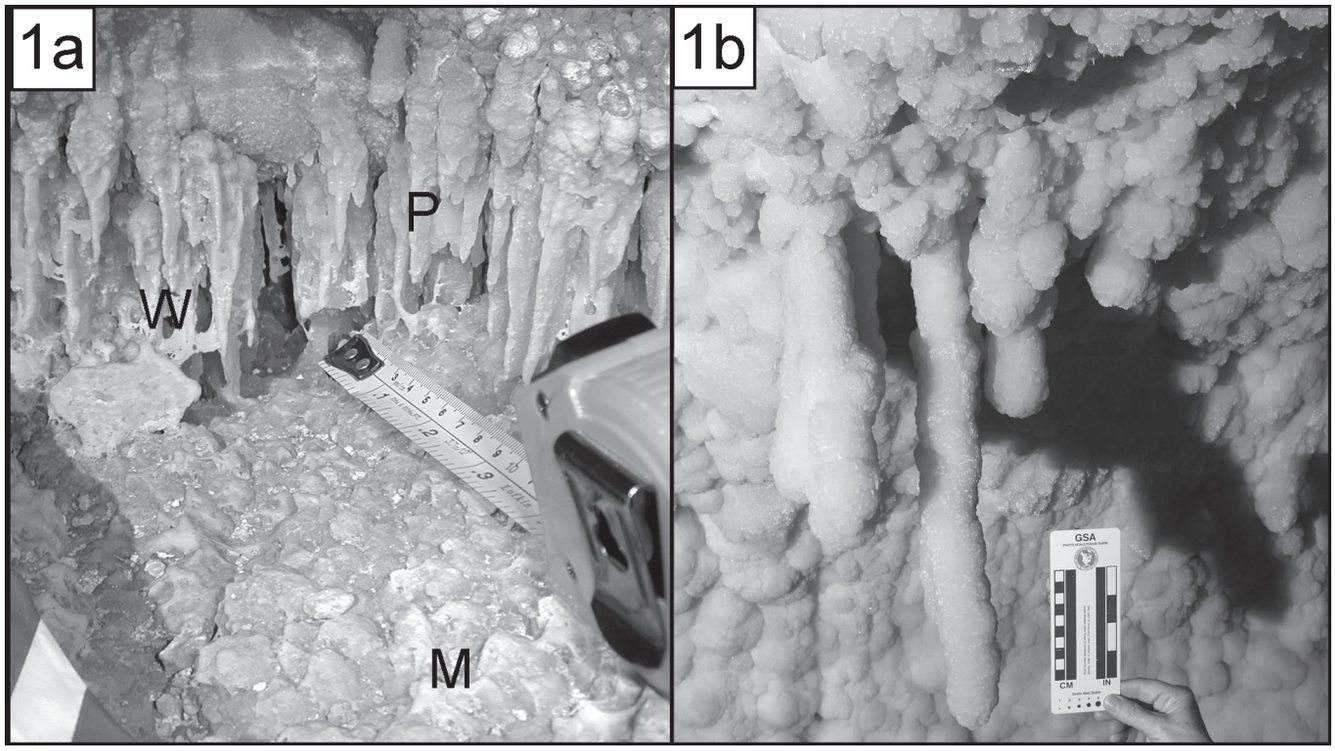

FIGURE 1.3.36. Lower Cave Pool features. (a) Webulite (W) and u-loops connecting pool fingers (perhaps chenille spar). The foreground pool spar is coated in peaks of pool meringue (M). (b) Giant pool fingers hanging from ceiling of pool; rest of wall is covered in knobs of pool spar. Photographs by K. Ingham.

of shelfstone indicating several stable pool water levels. Interestingly, an adjacent pool has only one level of shelfstone.

Crescent Shelfstone and Lily pad

Crescent shelfstone and lily pads are round to irregularly shaped features usually growing near the edge of a pool or on stalagmites in the center of pools (Hill and Forti, 1997). Crescent shelfstone, found in three pools, grows upward and outward from the sides of the pool resulting is a half-bowl shape with the raised rims extending 3-5 cm from the base up to the paleo-water line. Lily pads, found in two pools, are similar except they typically surround stalagmites or other protrusions that stick up in the center of pools. Unlike normal shelfstone, both crescent shelfstone and lily pads have a depression between the rim and the edge of the pool (shelfstone is continuous).

Pool finger and chenille spar

Pool fingers (Davis et al, 1990) and chenille spar (Hill and Forti, 1997) are both pendant features that hang down from submerged overhangs such as shelfstone, large pool spar knobs or pool-spar-coated bedrock walls. Pool fingers are round in cross-section while chenille spar is more bladed (Hill and Forti, 1997). However, the distinction is not always obvious in the field and some examples appear to contain both forms. Ten pools in Lower Cave contain either pool fingers or chenille spar. Although most are 1-2 cm in diameter and <5 cm long (Figure 1.3.36a), one pool contains large, slightly curved pool fingers 5-7 cm in diameter and 20-40 cm long (Figure1.3.36b). In Lower Cave, both forms are composed of calcite spar, indistinguishable from the local pool spar.

Conical crust, pool meringue

Conical crust and pool meringue are newly defined features growing on the pool bottoms, usually found in very shallow pools (<10 cm deep) (Rust et al., 2004; Queen and Melim, 2006). They are generally 1-2 cm diameter and grow up from the pool bottom into small cones or peaks (Figure 1.3.36a). Four pools contain these crusts; three are growing on top of round knobs of pool spar, the fourth appears to be growing on bedrock. Pool meringue appears to climb the wall of two pools and grades into webulite (see below).

Webulite, U-loops and Drips

Webulite, u-loops and drips are all calcitic extensions hanging from or draping between pool spar knobs or pool fingers. Webulite is sheet-like, punctuated by holes so it appears like a torn spider web (Davis et al. 1990). In Lower Cave, individual bridges are <1 mm thick and 1-5 mm wide. U-loops are single strands 1-3 mm wide draping between protruding features (Davis et al. 1990). Drips are like u-loops but they extend down 1-3 mm from a single knob (they may grade into chenille spar but no examples were found; Queen and Melim, 2006). Three pools contain one or more of these features. In two of these, pool meringue is also present and grades into webulite (Figure 1.3.36a).

Aragonite (?) bushes

One pool in Lower Cave contains delicate bushes growing in an active drip area. Based on the morphology, they are probably aragonite (no samples were taken).

Summary

Despite the relatively small number of pools described, there are a wide variety of features present. Pool spar coats nearly all pools and is the easiest way to identify paleo-pools as it marks the water line very effectively. Shelfstone, crescent shelfstone, and lily pads also mark the water line but only occur in about half the pools in Lower Cave, usually those with a steep edge. Pool fingers and chenille spar are pendant features found in ten pools. Unusual features including conical crust, pool meringue, webulite, U-loops and drips are found in four pools. Possible aragonite bushes occur in one pool.

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PARTNERING WITH BIOLOGISTS: BETTER ANSWERS THROUGH COLLABORATION

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Introduction

Multidisciplinary studies that involve geochemists, geologists, mineralogists, microbiologists, and molecular biologists can be crucial to our quest to investigate the role that microorganisms play in various geological materials and processes. Biologists have developed a suite of techniques such as DNA stains that reveal cell shapes in rock material, redox dyes that reveal live versus dead cells, and molecular phylogenetic techniques that assist in identifying living organisms present in geological materials using genetic sequences. Functional gene studies also can identify whether genes are present to carry out particular processes such as manganese oxidation. In our studies of microbial involvement in precipitation and dissolution phenomena in caves in the Guadalupe Mountains, we have come together from our separate fields of geology, geochemistry, microscopy, microbiology, and molecular biology to forge a partnership that is greater than the sum of the individual expertise. Here I will focus on the biological tools that have been used by our team to investigate how microorganisms participate in the formation of sulfur, iron, manganese, and carbonate speleothems.

Questions that Benefit from a Biological Point of View

Microorganisms are an integral part of our world, including the rocks and minerals that make up the Earth's crust. Beyond the sources of organic compounds that fuel microbial metabolism, there are many reduced mineral energy sources, such as hydrogen, sulfide, iron, manganese, etc. that fuel microbial metabolic pathways in surface and subsurface environments. Microorganisms have been found several thousand meters deep in the crust. Here, and in groundwater at varying depths, microorganisms may bring about the slow mineralization of organic compounds, releasing chemical waste products into their surrounding environment. Research has demonstrated that microorganisms are involved in the formation and dissolution of minerals in caves and other environments. Examples of processes that include a biological component include:

- The formation of moonmilk in caves.
- Formation of ferromanganese deposits on cave walls in Spider and Lechuguilla Caves
- Dissolution of limestone walls of caves to form punk rock.
- Formation of pool fingers and other pool precipitates in carbonate cave pools.
- Formation of caves through sulfuric acid-driven speleogenesis.
- Formation of gypsum and sulfur deposits in caves.

What biology can offer in these studies is (1) the identification of the microorganisms associated with these processes and products, (2) disclosure of locations in which microorganisms are present and active, (3) establishment of the microbial processes active in association with mineral environments, and (4) documentation of the ability of microorganisms in culture to form identical minerals to those found in a particular environment.

Visualizing Microbes

When investigating the role of microorganisms in mineral environments one needs to visualize where they occur and whether they are metabolically active, or just hanging around. Several stains exist that help visualize microbes on rock or soil particles. For example, in our studies we have used acridine orange to visualize cells in ferromanganese deposits. Acridine orange intercalates into DNA and makes the cells "light up" when visualized with an epifluorescent microscope (Figure 1.3.37A). This and other stains can be effective in differentiating cells from soil or rock particles when combined with cell morphology. Investigators can also determine whether microbial cells are metabolically active in an environment by using redox dyes such as INT or CTC. We have used INT with acridine orange counter staining to establish that microorganisms are metabolically active in the punk rock that underlies ferromanganese deposits, one step in showing that microorganisms are participating in the creation of these deposits on cave walls.

What Genetics and Microbiology Can Provide Geomicrobiology: Catching Microbes in the Act of Forming Minerals

In geological settings in which microorganisms are actively producing or assisting in the production of minerals, molecular

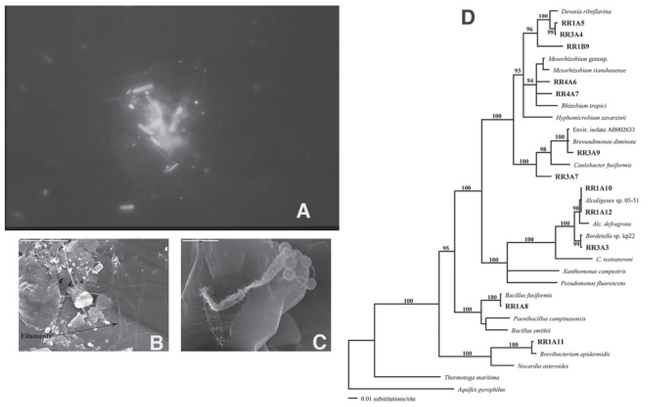

FIGURE 1.3.37. (A) Bacteria from biovermiculations in Cueva de Villa Luz, stained with acridine orange. (B) Filamentous microorganisms from a rock samples from Spider Cave. Note septa. (C) Beads-on-a-string bacterial morphologies with holdfasts that allow the bacteria to anchor to the rock in Spider Cave. (D) Phylogenetic tree of Rainbow Room sequences from Lechuguilla Cave.

microbial ecologists can provide essential information about the organisms involved. In the last two decades, researchers have discovered that we can culture less than one percent of the microorganisms present in the environment using standard culturing techniques. Thus, one cannot study biogenic mineral production with culturing techniques alone. One tool in molecular microbial ecologists' repertoire is what is called molecular phylogeny. Researchers take samples of the mineral deposits, extract the DNA (genetic material), and through a series of steps involving cloning and sequencing, produce a record of the genetic sequences for the organisms present. These genetic sequences are analyzed based on their evolutionary relationships to produce a "family" tree, called a phylogenetic tree (Figure 1.3.37D). Such trees can provide information that is useful in identifying the role that microorganisms are playing in the creation of mineral deposits and can guide microbial culturing efforts to more closely study mineral production by microorganisms. Two examples illustrate how this is done.

To investigate whether ferromanganese crusts on cave walls were biogenic, we extracted DNA from these deposits in Spider and Lechuguilla Caves in Carlsbad Caverns National Park. The molecular phylogenetic trees that were generated showed the presence of a novel organism that grouped with two other known manganese oxidizers, a slightly "smoking gun." Closely related genetic sequences can have similar or very different metabolic capabilities, so further evidence was needed. Therefore, our team cultured microorganisms from ferromanganese deposits from the Rainbow Room in Lechuguilla Cave that produced manganese minerals in culture, while killed controls did not produce minerals. We extracted DNA from these cultures and the resulting phylogenetic tree revealed closer relationships to known manganese oxidizers, providing a stronger case that these deposits were the result of manganese oxidizing bacteria (Figure 1.3.37D). The biologists provided the information on the microorganisms, while the geologists provided information on the geochemistry (i.e. nutrient sources for the microorganisms) of the habitat and the nature of the resultant minerals in culture, some of which matched those seen in the cave habitat.

Other useful molecular techniques include community fingerprinting, such as denaturing gradient gel electrophoresis (DGGE), which produces a pattern of community diversity on a gel. The more bands on the gel, the more organisms in that community and the brighter a particular band, the more individuals of that species that are present in the community under study. We have used this technique to address the question of whether one or more organisms occur in the different colors of cave ferromanganese deposits. We have also compared the number of organisms in different communities using this technique. DGGE allows us to investigate the question of whether one or more species are responsible for the production of all colors of deposits, showing a progression through time as different minerals are produced.

Telling Abiotic from Biotic in Microscopy

When geologists view scanning electron micrographs of their study objects they evaluate the morphologies present in terms of their knowledge of mineralogy and mineral shapes. Add a biologist, especially a microbiologist to the equation, and another dimensionis added to the identification process. Minerals mimic biological shapes, but there are some hallmarks of microbial life that can definitively say: "This is (or was) living." Cell structures such as holdfasts (cellular extensions for attaching to surfaces, Figure 1.3.37C) and septa (divisions cells in filaments, Figure 1.3.37B) and visual evidence for dividing cells are among the criteria that are used to differentiate biotic from abiotic.

Elements of Successful Partnerships

The key elements of bio-geo partnerships are communication, mutual respect, and an interest in learning something about the "other side." Within our team we often go to the field or work on the scanning electron microscope together. We've discovered that we view deposits of interest in situ in very different ways. Geologists see many clues to the origin of the deposits in the surrounding rock. Biologists see structures in the rock that suggest organisms were once present. Working together in the field is essential to meld these observations to achieve new understanding. The hard part is learning to communicate with each other, since we each come to the table with a different set of vocabulary to describe what we see. In creating keywords for a database of scanning electron micrographs we have decided to use words that describe morphology without making conclusions about whether the feature being described is biological or geological. In creating these jointly we have learned much about each other's science. Being willing to ask questions and to learn new things is essential to the successful partnership. Sharing equally in the credit and authorship of papers is also important. Valuing each other's contributions, points of view, knowledge, and perceptions are of paramount importance.

Summary

Biologists and geologists have a lot to gain from working together to discover the multitude of ways in which microorganisms interact with minerals. Questions about the past and present of our planet often involve microorganisms and better understanding can be achieved through working together. Microbes are truly the engineers of our planet and we have a great deal to learn about them.

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