Second-day road log, from Washington Ranch to Rader debris flow, Orla Road Castil outcrop, State Line Castile gypsum roadcut, Chosa Draw evaporite karst features, and Parks Ranch Cave

Lewis Land and David Love

2006, pp. 49-84. https://doi.org/10.56577/FFC-57.49

in:

Caves and Karst of Southeastern New Mexico, Land, Lewis; Lueth, Virgil W.; Raatz, William; Boston, Penny; Love, David L. [eds.], New Mexico Geological Society 57th Annual Fall Field Conference Guidebook, 344 p. https://doi.org/10.56577/FFC-57

This page is intentionally left blank to maintain order of facing pages.

EVAPORITE KARST OF THE CASTILE FORMATION AND REGIONAL GEOLOGY OF THE DELAWARE BASIN

SECOND-DAY ROAD LOG, FROM WASHINGTON RANCH TO RADER DEBRIS FLOW, ORLA ROAD CASTILE OUTCROP, STATE LINE CASTILE GYPSUM ROADCUT, CHOSA DRAW EVAPORITE KARST FEATURES, AND PARKS RANCH CAVE

LEWIS LAND & DAVID LOVE

Assembly Point: Washington Ranch Road near tufa dam

Departure Time: 7:30 AM

Distance: 88.8 miles

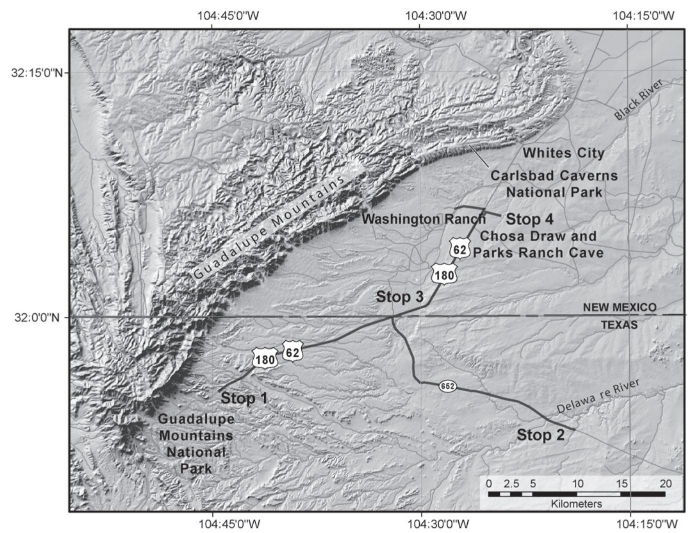
Four stops and three optional stops

SUMMARY

The second day of the conference will focus on evaporite karst in the upper Permian (Ochoan/Lopingian) Castile Gypsum; and aspects of regional geology of the Delaware Basin along the southwestern margin of the Guadalupe Mountains.

See Lucas minipapers for discussion of recent advances in Permian Stratigraphy, pages 60 and 62.

Our route will follow National Parks Highway southwest into Texas, running parallel to the Guadalupe Escarpment for several miles. As we proceed southwest, exposures of the Capitan Reef will become increasingly prominent, capped by layered strata of the backreef Yates and Tansill Formations. Stop 1 will visit the classic Rader submarine debris flow, consisting of large blocks of limestone that were eroded from the Capitan Reef escarpment, transported downslope, and are now imbedded in deep water sands of the basinal Bell Canyon Formation. After a brief stop for refreshments we will then proceed east on Orla Road to Stop 2 and examine one of the many castiles that punctuate the gypsum plain. Castiles are residual knolls of biogenic limestone imbedded in Castile gypsum, and provide important information regarding the role of H_2S in speleogenesis in the Guadalupe Mountains. The trip will then return to National Parks Highway and New Mexico, and at stop 3 will examine the famous State Line roadcut exposing varved and microfolded Castile gypsum. Our final stop will be in Chosa Draw, an area that provides impressive exposures of gypsum karst features such as sinkholes, skylights, resurgent springs, and caves, including Parks Ranch Cave, the second longest gypsum cave in the United States.



0.0 Assemble along the Washington Ranch road near the tufa dam across the Black River.

Proceed south, up the terrace riser. *0.1*

0.1 Crossing terrace of the Black River. In pits west of the road, exposures of the strath (base) of this deposit are 16-18 m above the Black River, and the top is 3 m higher. *0.3*

0.4 Yield sign and gate to Washington Ranch. At junction with Rattlesnake Springs road, turn left toward US 62-180. *0.1*

0.5 Rise onto higher terrace of the Black River with well-cemented limestone cobble conglomerate exposed in road cuts. This conglomerate continues east to the junction with US 62-180. *0.4*

0.9 Headquarters complex of Carlsbad Caverns National Park visible on skyline at 9:00, atop the frontal scarp of the Guadalupe Mountains. The scarp consists of the Capitan Reef facies capped by backreef Tansill limestone. *1.0*

1.9 Junction of Washington Ranch Road with 62-180. Note Historic Marker: Civilian Conservation Corps, Rattlesnake Springs Campsite. "The CCC provided employment for more than 50,000 young men in New Mexico during the Great Depression of the 1930s." **Watch for traffic. Turn right** (southwest) on US 62-180. 0.3

2.2 Milepost 10. Turnoff to Parks Ranch Cave to left. **Continue straight.** For the next 8 miles the route crosses the Yeso Hills at the northwestern edge of the Gypsum Plain. This complex surface is developed on the Castile Formation and includes a few outlying remnants of the overlying Rustler Formation (mostly Culebra Dolomite member) and underlying Salado residual deposits. Karst features of the Gypsum Plain are described in detail by Olive (1957), Sares (1984), Sares and Wells (1986), and Nance (1992). *1.1*

3.3 Dillahunty Road (CR 424) to south leads to CP Hill. **Continue straight.** Just northeast of Dillahunty Road, on the east side of the highway, is a small Lower Cretaceous outlier first reported by Lang (1947) and documented in greater detail by Kues and Lucas (1993). The outlier consists of slabs of oyster-rich limestone scattered on the Castile surface. These are not in-place outcrops but instead probably represent eroded Cretaceous debris that accumulated at the bottom of a paleo-sinkhole (Lang, 1947). The limestone slabs are very fossiliferous, and mostly yield echinoids and bivalves that indicate a middle Washita (= Albian) age (Kues and Lucas, 1993). *1.2*

4.5 Roadside park to right with view of Washington Ranch gas field, which produced dry gas from lower Pennsylvanian Morrow sands with lesser production from the Strawn and Delaware sandstones. The field (now a gas storage facility) is developed in over 180 m of structural closure on the top of the Morrow reservoir (Engwall, 1977). This site overlooks the upper Black River Valley extending to the base of the Guadalupe Escarpment. The SSE-trending Huapache monocline crosses the escarpment near the mouth of Slaughter Canyon at 2:00, and dips NE (Plate 1). Slaughter Canyon and Ogle Caves are located in the lower part of the canyon. The Huapache monocline, which is developed in late Paleozoic rocks over a Pennsylvanian thrust zone, formed during reactivation of the underlying zone of weakness and is probably Laramide in age (Hayes, 1964).

Ochoan (Lopingian) rocks in Black River Valley extend from Rattlesnake Canyon (4:00) to Big Canyon (1:00) and are deeply buried by the "Pleistocene gravel" unit of Hayes (1964, p. 38). The thickness of the upper Cenozoic alluvial deposits probably exceeds 90 m in the large fans that spread from Black, Double, and Slaughter Canyons. This valley fill could be as old as late Miocene (Hawley, 1993). *1.1*

5.6 Bottomless Lakes Road to right leads to partly flooded sinkhole complex in the Castile Formation on the edge of the Black River valley floor (Figure 2.1). The mouth of Slaughter Canyon is at 2:45 beyond the natural gas field. *0.3*

5.9 Roadcut to left is at the northwest edge of a large collapse feature in the Castile Formation. A core of dense carbonate rocks is flanked by steeply dipping conglomerate beds composed of rounded limestone clasts derived from the Guadalupe Mountains. Rubble in the roadcut includes red to orange clasts of siltstone and gypsum derived from Salado and basal Rustler dissolution residue.

Over the next mile the highway crosses three deeply weathered and very poorly exposed lamprophyre dikes trending NE-SW (Hayes, 1957). Age of the dikes is reported to be 32 MA (Calzia and Hiss, 1978). The features were originally mapped by Pratt (1954) and consist of a series of small, parallel intrusions that appear as long, narrow patches of rust-colored earthy material studded with occasional small fragments of dark-colored, fine-grained igneous rock. Although the walls of the dikes are indistinct and obscure on the ground, they are very conspicuous from the air (Figure 2.2). This dike system has been intersected by underground excavations in potash mines to the northeast.

Be advised that over the next few miles to the state line, mileposts may be missing. Watch your odometer. *0.3*

FIGURE 2.1. Bottomless Lakes sinkhole formed in alluvium overlying Castile gypsum, Black River Valley.

FIGURE 2.2. Aerial view of Pratt dike, east of National Parks Highway.

6.2 Milepost 6. Roadcut in the Castile Formation with typical interlaminated gypsum and calcite exposed to the left and biogenic replacement zone of secondary limestone in cut to right. *1.0*

7.2 Milepost 5. Crossing NE-SW trending solution valley (Olive, 1957) that extends eastward toward CP Hill. *1.7*

8.9 Outcrop of biogenic limestone on both sides of the road, trending northeast-southwest along the southern boundary of a solution valley. This linear outcrop is developed within the Castile Formation. *0.2*

9.1 Milepost 3. Crossing northwestern edge of the Yeso Hills overlooking the upper valley of the Black River drainage system. Turnout for famous "State Line gypsum roadcut" ahead. Group will return to this roadcut later in the day. *0.9*

10.0 Road descends through exposures of laminated and folded Castile Formation (Figure 2.3). Caves and sinkholes developed in gypsum on both sides of road. *0.2*

10.2 Milepost 2. Crossing valley floor at western edge of Yeso Hills. McKittrick Canyon Draw joins the upper Black River drainage system about 500 m to the north. Sulfur exploration holes in this area show about 56 m of alluvial fill. *0.7*

10.9 Ascending to surface of broad alluvial terrace, the distal part of a Pleistocene fan formed at the lower end of McKittrick Canyon Draw as mapped by P.B. King (1948). *1.2*

12.1 New Mexico-Texas state line. Texas Farm-Ranch Road 652 (Orla Road) on left. We return to this junction after stop 1 to venture into the Delaware basin area of northern Culberson and Reeves Counties, Texas. **Continue straight** on US 62-180. *1.2*

13.3 Bouldery terrace gravels of McKittrick Canyon Draw to right are capped by strong pedogenic calcretes. Route for next 8.5 miles crosses terraced apron of older (Plio-Pleistocene) alluvium deposited in the broad (dissolutional-erosional) depression between the Reef Escarpment and the Yeso Hills-Delaware Mountain area. *1.4*

14.7 Gravel pit in piedmont gravels to left. *1.9*

16.6 Road curves to right. McKittrick Canyon at 2:00. The exposed Capitan Reef (Figure 2.4) becomes progressively older toward Guadalupe Peak. The crest of the reef there is almost 300 m lower in the section than the outcrop at the mouth of Walnut Canyon at Whites City. The flat-lying beds on top of the escarpment are backreef carbonates that are more resistant to erosion than the massive reef wall. The reef facies is exposed in these canyons along the Guadalupe Escarpment. As accommodation permitted, the reef facies migrated basinward over its own steeply dipping forereef debris with thinner-bedded backreef carbonates following this basinward shift to cover older Capitan reef facies. *1.2*

17.8 Cobbly piedmont gravels exposed in old abandoned roadcut ~200 m to right. *3.4*

21.2 Highway curves at roadside rest area. Hills to left are basal Castile Formation capped by Plio-Pleistocene terrace gravel. *0.6*

21.8 Older gravel in old roadcut to right. Route ahead crosses dissected bedrock terrain at foot of the Guadalupe Mountains and from here to Guadalupe Summit is the transition zone between the Sacramento section of the Basin and Range physiographic province and the Pecos Valley section of the Great Plains (see physiographic map on back of NMGS geologic highway map). Late Cenozoic alluvial cover to Guadalupe Summit is discontinuous and relatively innocuous. *0.1*

FIGURE 2.3. Castile gypsum exposed in the famous State Line Roadcut. See Plate 7A for a color image of this outcrop.

FIGURE 2.4. Capitan Reef and forereef talus facies overlying Delaware Mountain rocks at El Capitan, southern promontory of the Guadalupe Mountains in west Texas. Distinctive sandstone unit near middle of photo is a deep water channel sand formed at the top of the Brushy Canyon Formation of the Delaware Mountain Group. See Plate 5B for a color image of this outcrop.

21.9 Optional stop: Black, thin-bedded Lamar Limestone exposed in roadcuts (Figure 2.5).

The Lamar has traditionally been designated the uppermost member of the Bell Canyon Formation of the Delaware Mountain Group, and the top of the Guadalupian Series in the Delaware Basin. Several limestone tongues of reef debris thicken toward the reef and interfinger with sandstone facies of the Bell Canyon. In descending order the interfingering tongues are the Lamar, McCombs, Rader, Pinery, and Hegler limestones. The Lamar was deposited here by low density turbidity currents in about 520 m of water, about 7 km from the reef margin. These dark, silty basinal limestones are sparsely fossiliferous, but abundant fusulinids from Lamar outcrops closer to the reef have been correlated to the Tansill Formation (uppermost unit of the Guadalupian Series in the backreef facies).

See Tyrrell et al., minipapers for an expanded discussion of shelf to basin correlations in the Guadalupe Mountains-Delaware Basin region, pages 64, 67, and 70.

A distinctive gray laminated bed containing flattened ammonoids and brachiopods on bedding planes can be traced along the upper part of the outcrop. Post-Lamar beds have recently been designated as the Reef Trail Member of the Bell Canyon Formation by Wilde et al. (1999). Unusual in marine limestone, small gypsum pseudomorphs are present near the top of the section, the result of early diagenetic replacement of carbonate below the sea floor. Note superficial folds in top of roadcut, probably caused by soft-sediment deformation near the base of a large gravity slide moving slowly downslope. *0.5*

See Brown minipaper for detailed discusssion of the Lamar limestone outcrop, p. 73.

22.4 More Lamar Limestone in roadcuts to left and right. *0.6*

FIGURE 2.5. Thin-bedded Lamar limestone member of the Bell Canyon Formation. The Lamar was deposited by low-density turbidity currents in >500 m of water, about 7 km from the reef margin. See Plate 6A for a color image of this outcrop.

23.0 Descending through Plio-Pleistocene semi-consolidated sediment along margin of McKittrick Canyon. *0.6*

23.6 Turnoff to McKittrick Canyon to right. **Continue straight.** *0.3*

23.9 Bell Canyon Sandstone capped by older piedmont deposits in canyon walls to left. *0.2*

24.1 Crossing Bell Canyon. Bell Canyon sandstones exposed in walls of lower Bell Canyon. *0.8*

24.9 Stop 1: Rader member of Bell Canyon Formation; submarine slide blocks (Figure 2.6; Plate 2). Park on shoulder.

Caution: Traffic is especially hazardous at this stop.

The following discussion has been modified from Scholle (2000). These roadcuts display a major submarine slide deposit at the base of the Rader Member of the Bell Canyon Formation. The southwest end of the outcrop consists of laminated sandstones and siltstones of the Bell Canyon. At the top of this sandstone is a conglomerate zone with limestone blocks set in a sandstone matrix. The limestone clasts are very poorly-sorted and range from pea-size pebbles to boulders the size of a car. The clasts are non-dolomitic, light-colored limestones derived from the Capitan Reef and upper fore-reef environments (Lawson, 1989). Above the zone of bouldery rubble is a thick, graded bed of similar, but finer-grained limestone clasts with carbonate matrix and cement. This is in turn capped by a series of thin-bedded, fine-grained, dark-colored limestones typical of the basinal limestone members of the Bell Canyon Formation. This slide deposit is one of several that have been identified in the Delaware Basin and represents a slender, possibly channelized tongue of transported debris

FIGURE 2.6. Rader Formation submarine debris flow. Limestone cobble conglomerate with clasts ranging from pea size to boulders the size of a car are imbedded in a sandstone matrix. The clasts are non-dolomitic, light-colored limestones eroded from the Capitan reef margin and upper fore-reef environments and transported downslope into deeper water. See Plate 6B for a color image of this outcrop.

that extends nearly 8 km into the basin from the reef crest. The deposit has been shown to thin rapidly from reef to basin. It is nearly 30 m thick at the base of the fore-reef slope but has thinned to less than 3 m at this site.

The mechanism of transport of the large limestone clasts was probably a submarine debris flow, and the volume of material involved is comparable to that of large, documented subaerial landslides (Newell et al., 1953). As with subaerial debris flows, there is relatively little disturbance of the underlying soft sediment substrate. However, the incorporation of sandstone matrix with limestone boulders, and the abruptly terminated lateral margins of the slide suggest that there was some erosion and inclusion of the underlying Bell Canyon sandstone into the slide. The event which triggered this slide also produced a turbidity current, which deposited the thick, graded bed overlying the conglomerate zone. This association of a coarse-grained debris flow overlain by finer-grained turbidity deposits is common and has also been observed in modern submarine slides (e.g., Heezen and Drake, 1964). After stop, continue southwest on 62-180 to Nickel Creek for turnaround. *0.3*

25.2 Roadcuts to left and right are Plio-Pleistocene carbonate-cemented gravels on Manzanita limestone. The Manzanita is the uppermost member of the Cherry Canyon Formation. In this roadcut and in outcrops to the south it contains waxy green bentonitic clays, which are altered volcanic ash. The Cherry Canyon Formation passes beneath the Capitan Reef and is equivalent to the older Goat Seep Reef, which was deposited north of the younger Capitan. 0.6

See Diemer et al. minipaper for an expanded discussion of bentonite in the Manzanita limestone, p. 75.

25.8 Crossing Nickel Creek. *0.1*

25.9 **Turn right** into Nickel Creek Station parking lot for turnaround. Proceed back northeast toward Orla Road. *4.6*

30.5 Rest area to right. **Turn right** into parking lot for mid-morning snacks, beverages and pit stop. After rest stop, continue northeast on 62-180. *9.0*

39.5 **Turn right** onto Texas Farm Road 652/Orla Road. Route to south and east ahead is on a Pleistocene alluvial terrace mapped as "Qao: alluvium, colluvium, caliche and gypsite on surfaces dissected by modern drainage" (Bureau of Economic Geology, 1983). This alluvium appears to be part of an abandoned fan of the McKittrick Canyon drainage. The alluvium now straddles the divide between the Black and Delaware Rivers and is pock-marked with numerous karst features. Proceed SE on Orla Rd. *1.0*

40.5 Roadcut to left exposes cobbly terrace gravels overlying gypsiferous and fine-grained older valley fill. Descend into blind valley. *1.2*

FIGURE 2.7. Castile on horizon, formed from resistant biogenic limestone imbedded in Castile gypsum.

41.7 Route enters the Yeso Hills area of the Gypsum Plain, underlain by Castile gypsum. This rolling plain continues eastward toward the Black and Pecos Rivers. *1.0*

42.7 For the next four miles, patches of gravelly calcrete (pedogenic Stage IV) form high-level terrace remnants on the Gypsum Plain. El Capitan at 3:00. *0.7*

43.4 Conical hills at various points on the horizon are castiles (Figure 2.7). *1.3*

44.7 Junction with FR 1108 to Delaware Springs. **Bear left** and continue on FR 652. About 2 km south is the north flank of an eastward-plunging anticline cored by uppermost Guadalupian Bell Canyon sandstone (King, 1948). *2.5*

47.2 Solution front of Castile gypsum to left. Road continues on extensive plain of piedmont gravels. *2.0*

49.2 Junction with FR 1165 (pipeline station to left). **Continue straight** on FR 652. *0.5*

49.7 Crossing Alligator Draw, with adjacent roadcuts in Castile gypsum. *2.2*

51.9 Roadcuts in Castile Formation continue. *0.5*

52.4 Crossing Pokorny sulfur deposit drilled by Addwest Minerals. Klemmick (1993) reports that the Pokorny sulfur deposit was inadvertently discovered by University of New Mexico researchers while studying climatic cycles and varved strata in the Castile Formation (Davis and Kirkland, 1970). A 276 m core test drilled by the University in 1969 intercepted biogenic limestone containing 3 net m of solid sulfur. All the sulfur mineralization is hosted by the Castile Formation, in which sulfur-bearing biogenic limestone has replaced anhydrite or gypsum.

Valley of the Delaware River about 1.6 km to the south follows a major N60°E trending solution-subsidence trough

(Olive, 1957). King (1949) and the Bureau of Economic Geology (1983) show a 3.2 km-wide horst with the same trend at the north edge of the valley. This feature is expressed as a salient of Bell Canyon sandstone extending outward from the eastern base of the Delaware Mountains just south of the highway at this point. *0.2*

52.6 Conical Sulfur Mine Hill at 3:00 is a castile feature 8 km to SSW on the southern margin of the Delaware River subsidence trough. Smith (1978) describes the hill as the Grant sulfur deposit, and reports that in the early 20th century, a shaft was dug between two limestone castiles and reported to be 30 m deep, lined with a 10 cm layer of crystalline sulfur, and copious volumes of hydrogen sulfide gas issued from the shaft. Stafford (2006, personal communication) recently inspected the mine and estimated the depth at only 15 m. Stafford also found no evidence of hydrogen sulfide and suggested that the H_2S reported by many workers over the years is more likely sulfur dioxide. *0.9*

53.5 Northern rim of subsidence trough and roadcut in Castile gypsum. Crossing valley of Delaware River. *0.2*

53.7 Roadcut in Pleistocene terrace deposits, primarily gypsiferous sand with thin pebble interbeds, capped with well-developed gypsic soil containing a thin, indurated gypcrete. *0.4*

54.1 Bridge over Delaware River. Since at least late Miocene time (past 12 million years), this and other tributaries to the ancestral Pecos River were part of a surface and subsurface discharge system that shaped much of the topography along this route. Past intervals of cooler and moister climate, such as occurred during Pleistocene glacial-pluvial stages, supported much larger sustained discharges than those during the Holocene (past 10,000 years). The old Butterfield Stagecoach Line followed the Delaware River. The first way station east of the Pinery at Guadalupe Pass (west of Stop 1) was at Delaware Springs about 13 km west of this crossing. *0.1*

54.2 About 6 m of terrace material is exposed in this roadcut, consisting primarily of structureless gypsite and gypsiferous silt, with discontinuous, finely laminated zones exposed near the base of the cut. Age of the deposit is probably mid-Pleistocene and it forms a prominent intermediate-level terrace on the south side of the Delaware River in this area. *1.2*

55.4 Castile gypsum exposed on shoulder of road. *0.1*

55.5 Crossing south rim of solution-subsidence trough. To the south is disrupted surface drainage along Sinkhole Flat of lower Castile Draw. **Prepare to stop on right shoulder of highway. Follow flaggers' directions! Once out of vehicles, watch for traffic.** *0.5*

56.0 Stop 2: Castile on south side of Orla Road.

The following discussion is modified from Brown and Loucks (1988).The gypsum plain east of the Guadalupe Mountains forms a rolling topography punctuated by isolated hills or clusters of hills known as castiles (Figure 2.8) (Adams, 1944). These features result from differential erosion of gypsum around areas of more resistant secondary limestone. This castile is formed by a V to U-shaped, near-vertical, dike-like body of steeply-dipping

FIGURE 2.8. Cluster of castiles on Orla Road.

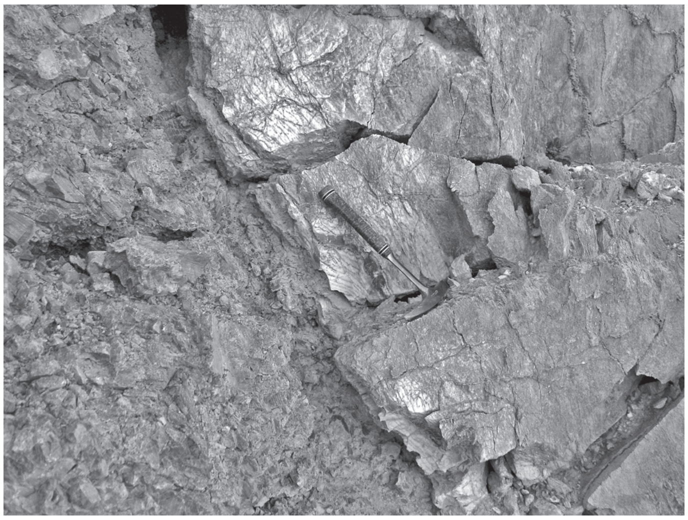


FIGURE 2.9. Sharp contact between Castile gypsum and biogenic limestone in castile.

laminated and brecciated secondary limestone enclosing and surrounded by steeply-dipping gypsum of the Castile Formation and by gypsum breccia. The roadcut exposes a sharp contact between the gypsum breccia and secondary limestone dipping steeply toward the road (Figure 2.9). Small amounts of secondary limestone occur within the gypsum breccia. Structure in the outcrop and around the hill is not consistent with a simple breccia pipe.

The limestone is assumed to have formed by metabolysis (associated with sulfate-reducing bacteria) of the original calcium sulfate and migrating hydrocarbons during Cenozoic uplift, erosion and salt dissolution. The calcium is reincorporated into calcite, the hydrocarbons are oxidized to carbonate which is incorporated into the calcite, sulfate is reduced to H_2S, and energy is yielded to the bacteria. Locally, economic quantities of sulfur are produced by further reaction of the H_2S with oxidizing groundwater. Highly depleted $\delta^{13}C$ values in the Castile carbonates (often < -28‰) are consistent with a hydrocarbon origin for the castiles

FIGURE 2.10. Castile microfolds and laminae preserved in biogenic limestone of castile breccia.

(Kirkland and Evans, 1980). The generalized reaction in the presence of sulfate-reducing bacteria was probably:

$$CaSO_4 + CH_4 \rightarrow H_2O + H_2S + CaCO_3 + \text{energy}$$

Most calcite was locally formed as a volume-for-volume reaction, with laminae and microfolds originally present in the precursor anhydrite perfectly preserved in small areas of secondary limestone (Figure 2.10). On a larger scale, porosity is produced because a 20% volume reduction occurs when anhydrite is converted to calcite. This volume difference is accommodated by formation of secondary vugs and by brecciation. After stop, turn vehicles around to northwest and return on FM 652 to highway 62-180. *16.6*

72.6 **Turn right** and proceed northeast on US 62-180. Route ahead is on Pleistocene alluvial terrace. *1.3*

73.9 Descend to valley floor at eastern edge of alluvial terrace. McKittrick Canyon Draw joins the upper Black River drainage system about 500 m to the north and continues northeast to Washington Ranch. Yeso Hills erosional escarpment visible to right (Figure 2.11). *0.7*

74.6 Route ascends Yeso Hills. **Prepare to park as directed by flaggers.** *0.1*

74.7 Stop 3: Castile gypsum state line roadcut.

Park on shoulder and prepare to walk along roadcut to examine exposures.

Caution! Traffic is particularly hazardous at this stop.

The text that follows has been modified from Scholle (2000).

The Castile gypsum is the oldest Ochoan (or Lopingian; see Lucas, 2006) unit in southeastern New Mexico, and conformably overlies the Guadalupian Bell Canyon Formation. The Castile is entirely confined to the Delaware Basin and does not extend onto the adjacent Northwest Shelf. The formation is reported to reach a maximum thickness of ~550 m in the subsurface of the northern Delaware Basin, the bulk of which consists of laminated anhydrite with intervals of laminated halite. The Castile grades conformably into massive salt beds of the overlying Salado Formation, the host rock for the WIPP Site and for the potash minerals that are extensively mined east of Carlsbad.

See Holt et al. minipaper for a discussion of depositional cycles in the Salado Formation, p. 78.

Onset of Castile evaporite deposition coincided with termination of reef growth around the margins of the Delaware Basin. Eustatic fall in sea level, reef growth, or tectonic uplift may have inhibited influx of normal marine waters into a semi-restricted basin. These factors, combined with extreme aridity and high evaporation rates in an equatorial setting would have led to drastic increases in salinity of waters in the basin, and extinction of reef-building organisms. Marine influx continued in a restricted form, supplying the salts that now make up the Castile and Salado Formations, and it appears likely that a shift in marine water supply vs. evapo-

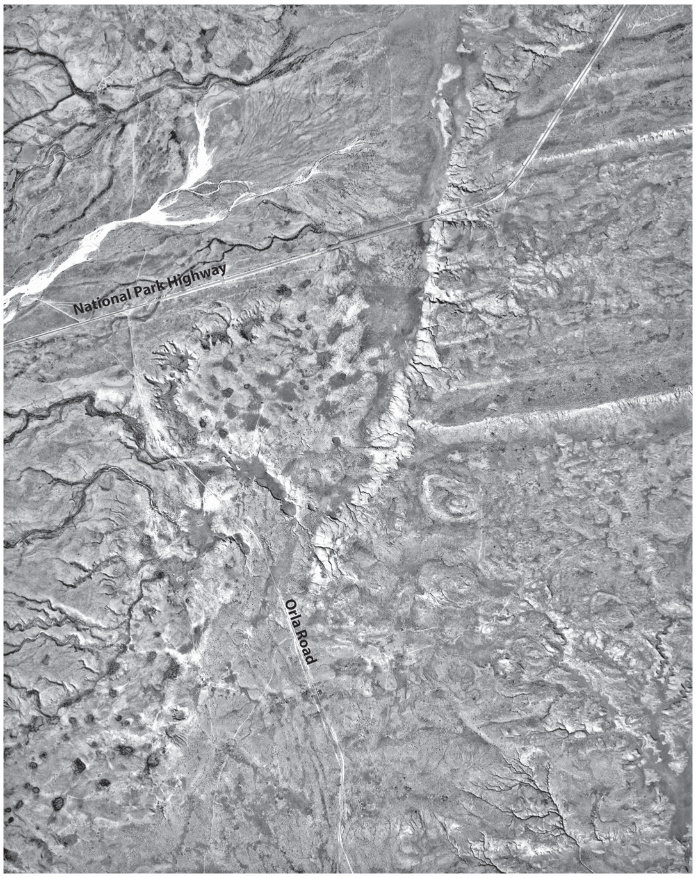


FIGURE 2.11. Areal photograph of the gypsum plain near the Texas-New Mexico state line. Note sinkholes and trend of the Castile gypsum solution front that trends from northeast to southwest.

ration led to an abrupt transition from carbonate to evaporite sedimentation when a critical salinity level was reached.

Although some fall in water levels in the Basin may have occurred during Ochoan time, the fine-scale lamination of the Castile and absence of other sedimentary structures indicate that most of the unit was deposited in deep water. The Castile Formation has thus become identified as a classic example of a deep-water, marine evaporite deposit, and represents evaporative filling of a restricted basin with bathymetric relief of ~550 m at the end of Guadalupian time.

Where the Castile is exposed at the surface the anhydrite component has been largely altered to gypsum. At this roadcut the section consists of couplets of white to gray gypsum laminae 1 – 5 mm in thickness, alternating with sub-millimeter, dark-brown laminae containing a mixture of calcite and organic matter (Figure 2.12). Larger-scale cycles are also evident, especially in the upper part of the outcrop. The gypsum-calcite couplets have been interpreted as seasonal varves (Kirkland, 2003; Anderson et al., 1972), in part because of their remarkable lateral continuity; individual laminae have been traced across the basin for over 110 km. The zone with disrupted bedding near the middle of the section has been correlated to the lowest halite bed of the halite III zone, about half way through the Castile section (Anderson and Kirkland, 1987).

The calcite-organic matter layers are thought to represent annual freshening of the water and development of plankton blooms. The anhydrite/gypsum layers represent more restricted, evaporitic conditions. If the seasonal varves model is correct, it implies that the entire 550 m of the Castile Formation were deposited in about 220,000 years.

See Anderson minipaper for an expanded discussion of paleoclimatic implications of the Castile varves, p. 80.

One of the most striking features of the state line roadcut is the presence of microfolding in concentrated zones or pods in the lower part of this exposure (Figure 2.13). Zones of microfolding a few cm thick are sandwiched between unfolded laminae. The folding is clearly post-depositional and has been interpreted to represent volume changes due to hydration and conversion to gypsum during uplift and exposure of the Castile at the surface. However, more recent studies (e.g., Kirkland and Anderson 1970; Anderson and Kirkland 1987; Watkinson and Alexander, 1993) show that the fold axes generally trend N30-50°W, parallel with Tertiary faulting in the area, suggesting that the folds may be of Tertiary age and tectonic in origin. After stop 3, continue northeast on US 62-180. *0.9*

FIGURE 2.12. Castile gypsum exposed at State Line Roadcut. Note closely-spaced, light and dark couplets of gypsum and calcite/organic matter. These couplets have been interpreted to represent seasonal varves. Individual couplets have been correlated across the Delaware Basin for >110 km.

FIGURE 2.13. Microfolded varves in Castile gypsum at State Line Roadcut. See Plate 7B for a color image of this outcrop.

75.6 Milepost 3. Physiographic features of the Yeso Hills-Gypsum Plain developed on the Castile Formation continue to the east and northeast to the Black and Pecos Rivers. *1.0*

76.6 Milepost 4. Outcrops of biogenic limestone on both sides of highway, trending NE-SW along the southern boundary of a solution valley (Olive, 1957). This outcrop is imbedded in Castile gypsum in a linear fracture that crosses the highway. The solution valley continues to Milepost 5. *1.0*

77.6 Optional stop: Yeso Hills selenite mine and K Hill ahead to right. Milepost 5. 0.1

77.7 Gravel road to right crosses cattle guard and gate and leads to Yeso Hills selenite quarry and K Hill paleosink. Selenite quarry is 0.5 miles down road and to left, and is easily visible on sunny days because of sunlight reflected from large selenite crystals exposed in quarry face.

The Yeso Hills selenite occurence is a high-purity lenticular gypsum deposit covering an area of ~13,000 m^2. The deposit consists of interlocking, cleavable masses of selenite exposed across the northern boundary fault of a NE-trending solution-subsidence depression formed in the Castile gypsum. About 240 m^3 of selenite were excavated from a 30 m long trench during WWII and used as a substitute raw material for the manufacture of capacitors. The largest crystal in the deposit is 5.5 m long by 2 m wide, and forms the northwest wall of the trench. Although now abandoned, the quarry is still visited by local collectors who use the selenite as decorative stone for landscaping. Crawford (1993) proposed that the selenite may have formed by local hydration of anhydrite along a solution-subsidence boundary fault by means of which upwelling waters from the underlying Delaware Mountain Group contacted Castile evaporites.

K Hill, 1.2 miles further east beyond the selenite deposit, is a 17 m high paleokarst erosional remnant formed in a 1.6 km wide solution-subsidence trough within the Castile Formation. K Hill is covered with cobble to boulder-size breccia clasts embedded in gypsum and selenite of the upper Rustler and possibly the Salado Formations. Crawford (1993) interprets the breccia clasts as karstic slump debris from the Culebra and Magenta dolomite Members of the Rustler Formation. The eastern part of the hill consists of a rotated, downdropped karst block that preserves red-brown, fine-grained sandstones of the uppermost Permian Dewey Lake Formation in contact with red-brown, coarse-grained, cross-bedded sandstones and conglomerates of the basal Cretaceous Cox Formation. *0.9*

78.6 Milepost 6. Roadcut in Castile Formation, with typical interlaminated gypsum and calcite exposed to right and biogenic replacement zone of secondary limestone in cut to left. *0.6*

80.2 Roadside park to left overlooks Washington Ranch gas field. *1.2*

81.4 Dilahunty Road to right. *1.2*

82.6 Milepost 10. Slow. **Turn right** across cattle guard onto BLM road to Parks Ranch Cave. *0.1*

82.7 Fork in road. **Bear left** onto road parallel to National Parks Highway. Contact between terrace gravels and Castile Formation to right. *0.4*

83.1 Gate. **Open gate and turn right** (east) parallel to fence. Blocks of Castile gypsum in road ahead. (Be sure to close gate.) *0.8*

83.9 BLM fences dead ahead around two sinkhole entrances to Parks Ranch Cave. *0.1*

84.0 Stop 4a. Parks Ranch Cave. Half of group parks here and prepares to enter cave.

Remainder of group stays in vans and continues down road. *0.6*

84.6 Stop 4b. Chosa Draw.

Road swings north at fence line.

Park vans and prepare for hike along upper reaches of Chosa Draw to view gypsum caves, resurgent springs and other karst features developed in the Castile gypsum.

The Chosa Draw area contains some of the finest examples of epigenetic gypsum karst in the Gypsum Plain region of southeastern New Mexico and west Texas. The area is characterized by intensive karst development in the Castile Formation, with up to 50 sinkholes/km2. Castile gypsum in Chosa Draw is the host rock for Parks Ranch Cave, the second-longest documented gypsum cave in North America. Parks Ranch Cave has multiple entrances, most of which are formed in sinkholes (Figure 2.14). Other features to note include selenite crystals armoring the beds of some

FIGURE 2.14. Sinkhole entrance to Parks Ranch Cave, Chosa Draw.

arroyos in the area (Figure 2.15); skylights (Figure 2.16); well-developed rillenkarren (Figure 2.17); resurgent springs (Figure 2.18); and large gypsum slump blocks (Figure 2.19).

FIGURE 2.15. Selenite gypsum armoring the bed of an arroyo in Chosa Draw.

FIGURE 2.16. Skylight in ceiling of cave, Chosa Draw.

FIGURE 2.17. Gypsum rillenkarren in Chosa Draw.

FIGURE 2.18. Resurgent spring downstream from Parks Ranch Cave.

See the Stafford minipaper for an expanded discussion of karst geology and hydrology of the Chosa Draw area, p 82.

After Stop 4, turn vehicles around and return to National Parks Highway (62-180). *2.0*

86.6 Junction with National Parks Highway. **Turn right.** *0.3*

86.9 Junction with Washington Ranch Road. **Turn left** and return to Washington Ranch. *1.5*

88.4 Gate to Washington Ranch. **Turn right** and enter ranch. *0.4*

88.8 Crossing tufa dam at WR.

End of Day 2 road log.

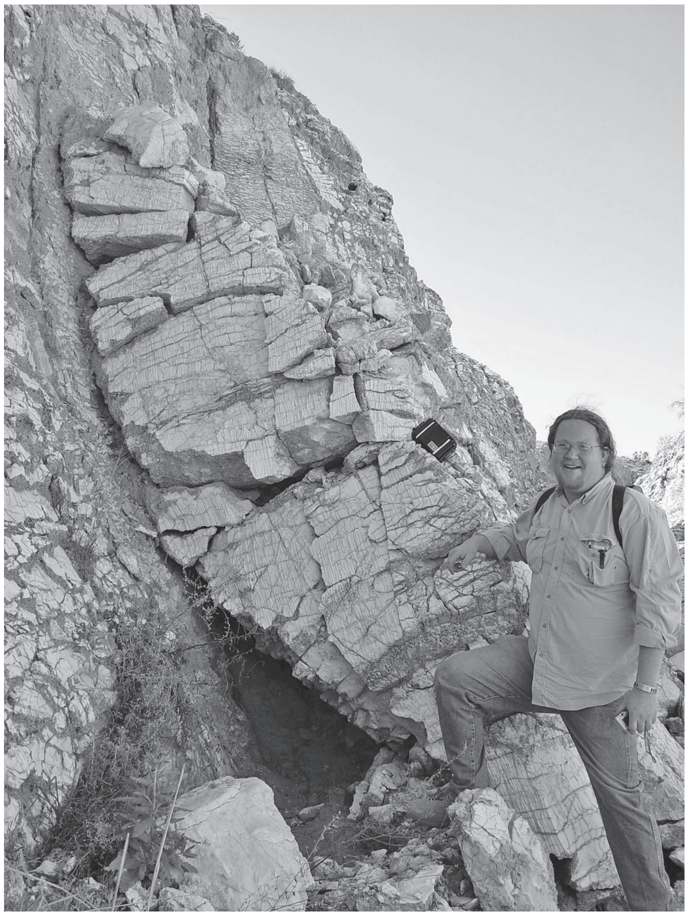


FIGURE 2.19. Gypsum slump block in Chosa Draw.

THREE PERMIAN SERIES

Spencer G. Lucas
New Mexico Museum of Natural History, 1801 Mountain Road N.W., Albuquerque, NM 87104

In 1840, British geologist Roderick Murchison (1792-1871) visited Russia as a guest of the Czar. East of Moscow, he examined strata in the Perm region of the western Urals, and applied the name Permian System to a "vast series of beds of marls, schists, limestones, sandstones and conglomerates" (Murchison, 1841, p. 419) that overlie the Carboniferous strata in a great arc that extends from the Volga River on the west to the Ural Mountains on the east, and from the White Sea on the north to Orenburg on the south. Thus was born the Permian System, and in western Europe it soon came to be equated to the British Magnesian Limestone and New Red Sandstone, and to the German Rotliegendes and Zechstein. This became the basis for the longstanding division of the Permian System into two series (epochs), Lower (Early) and Upper (Late).

A Permian System composed of two series won international acceptance in the twentieth century. This is well reflected by Harland et al. (1982, chart 2.7), who divided the Permian into Early and Late epochs to which they later applied the names Rotliegendes and Zechstein (Harland et al., 1990, fig. 3.7). However, by the 1990s, a group of earth scientists was undertaking a serious overhaul of the Permian timescale as the work of the Subcommission on Permian Stratigraphy, which is part of the International Commission on Stratigraphy of the International Union of Geological Sciences. The Subcommission thus advocated changing the twofold subdivision of the Permian into a threefold subdivision (Glenister et al., 1992). A global view of large scale Permian depositional cycles and biotic change was thus marshalled to support recognition of three Permian series, which were named the Cisuralian (Early), Guadalupian (Middle) and Lopingian (Late). The Cisuralian is based on the section in the Russian Ural foreland basin, although these Lower Permian strata were not part of Murchison's original Permian; he considered them Carboniferous, though by the 1940's Russian and Soviet stratigraphers had moved them into the Permian.

The strata that typify the stages of the type Guadalupian Series are in the area of this field conference—the Road Canyon, Word and Capitan formations of the Permian Basin of west Texas and southeastern New Mexico. In southern China, Upper Permian strata of the Wuchiapingian and Changxingian stages define the Lopingian Series. The three Permian series encompass nine stages that have their boundaries (bases) defined by conodont biostratigraphy (Fig. 2.20).

Thus, a "new" Permian timescale has come into use (Wardlaw et al., 2004; Henderson, 2005). Critical to the formulation of this scale has been the great Guadalupian reef complex and associated strata that provided the basis for recognition of a Middle Permian Series. All timescales should evolve towards more detailed subdivision, because this provides greater precision. I thus see the division of the Permian into three series instead of two as a step forward in efforts to provide a more precise chronology for Permian time.

REFERENCES

Glenister, B. F., Boyd, D. W., Furnish, W. M., Grant, R. E., Harris, M. T., Kozur, H., Lambert, L. L., Nassichuk, W. W., Newell, N. D., Pray, L. C., Spinosa, C., Wardlaw, B. R., Wilde, G. L. and Yancey, T. E., 1992, The Guadalupian: Proposed international standard for a Middle Permian series: International Geology Review, v. 34, p. 857-888.

Harland, W. B., Armstrong, R. L., Cox, A. V., Craig, L. E., Smith, A. G. and Smith, D. G., 1990, A geologic time scale 1989. Cambridge, Cambridge University Press, 263 p.

Harland, W. B., Cox, A. V., Llewellyn, P. G., Pickton, C. A. G., Smith, A. G. and Walters, R., 1982, A geologic time scale. Cambridge, Cambridge University Press, 131 p.

Henderson, C. M., 2005, Inter

Murchsion . 19, p. 417-422.

Wardlaw, B. R., Davydov, V. and Gradstein, F. M., 2004, The Permian Period; in Gradstein, F. M., Ogg, J. G. and Smith, A. G., eds., A geologic time scale 2004. Cambridge, Cambridge University Press, p. 249-270.

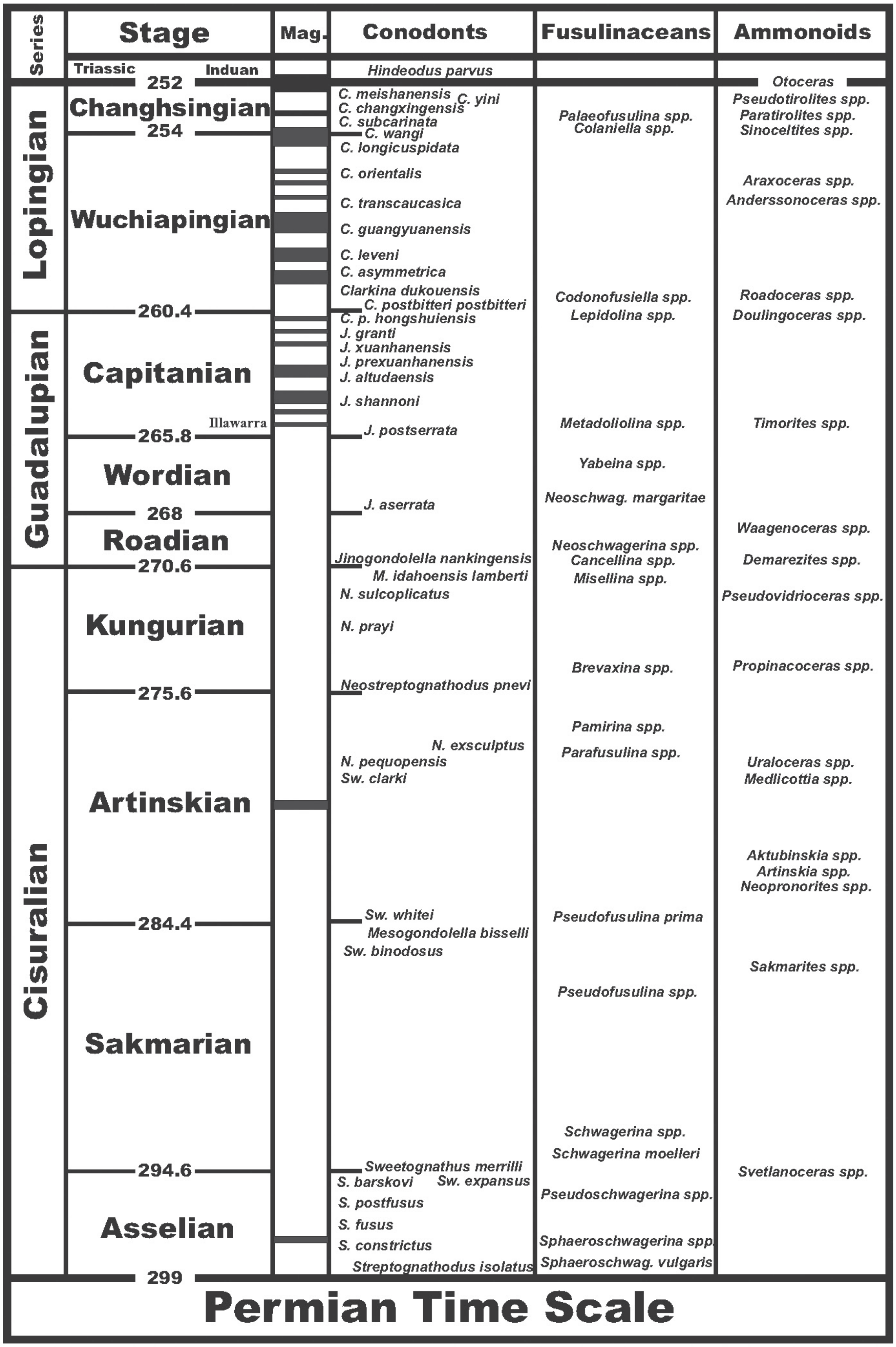


FIGURE 2.20. The current Permian timescale (from Henderson, 2005).

OCHOA GROUP, NOT SERIES OR STAGE, UPPER PERMIAN OF WEST TEXAS AND SOUTHEASTERN NEW MEXICO

Spencer G. Lucas
New Mexico Museum of Natural History, 1801 Mountain Road N.W., Albuquerque, NM 87104

The Permian basin of west Texas and southeastern New Mexico has long provided the standard Permian section for North America (Adams et al., 1939). This means that the stratotypes of the North American marine stages for Permian time are located in the Permian basin. Thus, commonly used chronostratigraphic terms such as Wolfcampian and Leonardian are based on Permian basin strata and fossils.

In subdividing the Permian strata of the Permian basin, Adams et al. (1939; also see Adams, 1935, 1944) named the uppermost subdivision the Ochoan Series (Stage). They took the name from the (now defunct) Ochoa Post Office in T24S, R34E, Lea County, New Mexico. This series, considered equivalent to Late Permian time, was based on the (ascending order) Castile, Salado, Rustler and Quartermaster (= Dewey Lake) formations, a section with a maximum thickness of approximately 1700 m (Fig. 2.21). These are lithostratigraphic units composed of extensive gypsum and/or halite, some carbonate rocks and siliciclastic red beds and that have very sparse age control (Lucas and Anderson, 1993, 1994, 1997).

The base of the Castile Formation overlies the Lamar Limestone Member of the Bell Canyon Formation, so it is no older than late Guadalupian (Capitanian) at its base. Counting repetitive laminae ("varves") in the Castile Formation suggests its entire deposition took only 200-300,000 years (Anderson and Kirkland, 1966; Anderson, 1982). There are no age indicators in the Salado Formation. Brachiopods and conodonts from the lower part of the Rustler Formation (Virginia Draw Member) indicate a Late Permian age, but are not more precise age indicators (Donegan and DeFord, 1950; Walter, 1953; Croft, 1978; Wardlaw and Grant, 1992). The Quartermaster Formation has yielded a few Permian bivalves, and magnetostratigraphic studies indicate it is of Late Permian age (Molina et al., 1989, 2000; Steiner, 2001). However, Ar/Ar ages on ash beds in the upper Quartermaster (though only published in an abstract: Renne et al., 1996) suggest it may span the Permian-Triassic boundary. This seems unlikely, though, given the magnetostratigraphy, and a Late Permian age for the entire Quartermaster Formation best fits the data (but see Retallack, 2005).

The Ochoan was long considered a series or stage equivalent to the Upper Permian. However, the Ochoan strata generally lack fossils or other data by which their precise ages can be determined. Furthermore, estimates of the duration of Ochoan deposition, and intra-Ochoan unconformities, suggest that the Ochoan strata represent a very incomplete record of Late Permian (Lucas and Anderson, 1993, 1994, 1997). Therefore, Ochoan should be dropped from the chronostratigraphic hierarchy and not treated as a series (epoch) or stage (age). Instead, Ochoa is best considered a lithostratigraphic group (Fig. 2.21). Thus, the Ochoa Group encompasses the Castile, Salado, Rustler and Quartermaster formations. It is an evaporite-dominated succession of mostly anhydrite and halite with minor carbonate and siliciclastic rocks up to 1700 m thick and present primarily in west Texas and southeastern New Mexico. Ochoa Group is a valid and useful lithostratigraphic concept for Permian basin strata of Late Permian age.

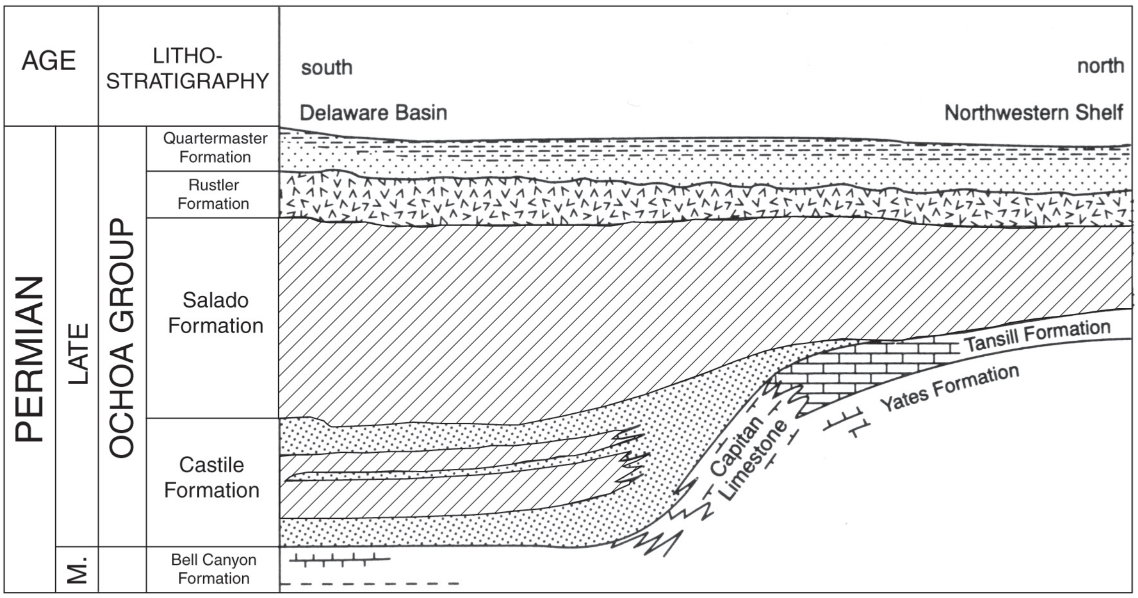


FIGURE 2.21. Generalized lithostratigraphy and age relationships of Ochoan strata in southeastern New Mexico and west Texas.

REFERENCES

Adams, J. E., 1935, Upper Permian stratigraphy of West Texas Permian basin. American Association of Petroleum Geologists Bulletin, v. 19, p. 1010-1022.

Adams, J. E., 1944, Upper Permian Ochoa Series of Delaware basin, West Texas and southeastern New Mexico: American Association of Petroleum Geologists Bulletin, v. 28, p. 1596-1625.

Adams, J. E., Cheney, M. G., DeFord, R. K., Dickey, R. I., Dunbar, C. O., Hills, J. M., King, R. E., Loyd, E. R., Miller, A. K. and Needham, C. E., 1939, Standard Permian section of North America: American Association of Petroleum Geologists Bulletin, v. 23, p. 1673-1681.

Anderson, R. Y., 1982, A long geoclimatic record from the Permian: Journal of Geophysical Research, v. 87, p. 7285-7294.

Anderson, R. Y. and Kirtkland, D. W., 1966, Intrabasin varve correlation: Geological Society of America Bulletin, v. 77, p. 241-256.

Croft, J. S., 1978, Upper Permian conodonts and other microfossils from the Pinery and Lamar Limestone Members of the Bell Canyon Formation and from the Rustler Formation, West Texas [M. S. thesis]: Columbus, Ohio State University, 176 p.

Donegan, B.and DeFord, R. K., 1950, Ochoa is Permian: American Association of Petroleum Geologists Bulletin, v. 34, p. 2356-2359.

Lucas, S. G. and Anderson, O. J., 1993, Stratigraphy of the Permian-Triassic boundary in southeastern New Mexico and west Texas: New Mexico Geological Society, Guidebook 44, p. 219-230.

Lucas, S. G. and Anderson, O. J., 1994, Ochoan (Late Permian) stratigraphy and chronology, southeastern New Mexico and West Texas: New Mexico Bureau of Mines and Mineral Resources, Bulletin 15, p. 29-36.

Lucas, S. G. and Anderson, O. J., 1997, Ochoa as a lithostratigraphic unit, not a chronostratigraphic unit of the Permian: West Texas Geological Society Bulletin, v. 36, no. 9, p. 5-10.

Molina-Garza, R. S., Geissman, J. W. and Lucas, S. G., 2000, Paleomagnetism and magnetostratigraphy of uppermost Permian strata, southeast New Mexico, USA: Correlation of the Permo-Triassic boundary in non-marine environments: Geophysical Journal International, v. 141, p. 778-786.

Molina-Garza, R. S., Geissman, J. W. and Van der Voo, R., 1989, Paleomagnetism of Dewey Lake Formation (Late Permian), northwest Texas: End of Kiaman Superchron in North America: Journal of Geophysical Research, v. 94, p. 17881-17888.

Renne, P. R., Steiner, M. B., Sharp, W. D., Ludwig, K. R. and Fanning, C. M., 1996, $^{40}Ar/^{39}Ar$ and U/Pb SHRIMP dating of latest Permian tephras in the Midland basin, Texas: EOS, v. 77, suppl. 46, p. 794.

Retallack, G. J., 2005, Permian greenhouse crises: New Mexico Museum of Natural History and Science Bulletin 30, p. 256-269.

Steiner, M. B., 2001, Magnetostratigraphic correlation and dating of West Texas and New Mexico Late Permian strata: New Mexico Geological Society, Guidebook 52, p. 59-68.

Walter, J. C., Jr., 1953, Paleontology of Rustler Formation, Culberson County, Texas: Journal of Paleontology, v. 27, p. 679-702.

Wardlaw, B. R. and Grant, R. E., 1992, Detailed conodont biostratigraphy of the North American regional stratotype of the Permian (West Texas) and its international correlation: 29th International Geological Congress, Abstracts, v. 2, p. 270.

OCCURRENCES OF THE FUSULINID *YABEINA TEXANA* IN THE BASAL PARTS OF THE TANSILL FORMATION AND LAMAR LIMESTONE MEMBER IN THE GUADALUPE MOUNTAINS AREA, WEST TEXAS AND NEW MEXICO

Willis W. Tyrrell, Jr.[1], Gorden L. Bell, Jr.[2], John A. Diemer[3] and Merlynd K. Nestell[4]
[1]Consulting Geologist, 5718 Bentway Dr., Charlotte, NC 28226, wtyrrell@carolina.rr.com
[2]Guadalupe Mountains National Park, HC 60, Box 400, Salt Flat, TX 79847, gorden_bell@nps.gov
[3]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, jadiemer@uncc.edu
[4]Department of Geology, University of Texas at Arlington, Arlington, TX 76019, nestell@uta.edu

INTRODUCTION

The small but distinctive fusulinid *Yabeina texana* is restricted to the basal part of the Tansill Formation on the shelf and the basal part of the Lamar Limestone Member of the Bell Canyon Formation in the Delaware Basin (Fig. 2.22). Its distribution also is restricted to the outer shelf, shelf margin and the basin margin facies. Thus its use as an inner shelf, middle shelf, shelf crest or central basin guide fossil is geographically limited. Like other fusulinids its depositional environment is considered to be shallow water and normal marine. It is most abundant in the basal part of the Tansill Formation, but it was first described from age equivalent deep water basinal limestone beds of the Lamar Limestone Member where specimens are considered to have been redeposited down slope at the basin margin. *Yabeina texana* has been found in the lower part of the Tansill Formation as well as in the basal part of the Lamar Limestone Member in McKittrick Canyon (Fig. 2.22; Tyrrell, 1962, 1969). It also may be present in coeval reef and reef talus facies of the upper part of the Capitan Formation but we are not aware of such occurrences.

Yabeina texana was first described from the basal 15 feet of the Lamar Limestone Member in Bear Canyon located about 1.5 miles southwest of the mouth of McKittrick Canyon (Skinner and Wilde,1955; see locality 1 on Fig. 2.22). Unlike most fusulinids, *Yabeina texana* is almost spherical as shown in thin sections on Figure 2.22 reproduced from Skinner and Wilde (1955). Its test is small (coarse sand size) compared to the large and slightly older *Polydiexodina* that characterizes most of the Capitanian Stage of the Guadalupian Series (Middle Permian). Skinner and Wilde (1955) report it is the smallest representative of the family Neoschwagerinidae known, and the most primitive species of the genus *Yabeina* which is late Middle Permian in age in the Pacific Northwest and Eurasia. In the Guadalupe Mountains area it has a very restricted vertical range and Wilde (1990, 2000) places it in his subzone PG-6-A which overlies his PG-5 zone (*Polydiexodina* zone).

YABEINA TEXANA LIES ABOVE A SIGNIFICANT FAUNAL BREAK

The easily recognized, long ranging, large *Polydiexodina* that characterizes most of the lower and middle Capitanian Stage in the Permian Basin, became extinct prior to deposition of the upper Yates "C" on the shelf and the McKittrick Limestone Member of the Bell Canyon Formation in the Delaware Basin. Wilde (2000) states that the "abrupt demise of *Polydiexodina"* ---- "constitutes an extremely important change in late Guadalupian fusulinacean evolution". This change may follow a significant sea level lowstand. When normal marine conditions returned to the shelf areas a new group of minute "Tethyan" fusulinids characterize the upper Yates and Tansill formations on the shelf and Lamar and Reef Trail members of the Bell Canyon Formation. These include *Codonofusiella* in the upper Yates; *Yabeina* in the basal Tansill and basal Lamar; *Paradoxiella* and *Richelina* in the overlying lower Tansill and the Lamar, and *Paraboultonia* in the Ocotillo Silt Member, upper Tansill and the Reef Trail Member (Skinner and Wilde, 1954, 1955; Tyrrell, 1969; Tyrrell, et al., 1978; Wilde, 1999). The post-*Polydiexodina* faunal break is a good place to put the top of the Guadalupian composite sequence CS 13 and high frequency sequence HFS G-24 of Kerans and Tinker (1999). This faunal break occurs within the Yates Formation (top of Yates "B" of Newell, et al., 1953; top of the Hairpin Dolomite; see Brown, 1996) on the Northwest Shelf and top of the McCombs Limestone Member in the Delaware Basin. No previous lowstand in the upper Guadalupian Capitanian Stage in the Permian Basin resulted in extinction of the long-ranging *Polydiexodina*. The short range zones of the Tethyan related fusulinids in the overlying upper parts of the Yates and Tansill formations suggest short-lived lowstand and highstand HFS for the uppermost Guadalupian Series prior to deposition of Ochoa Group evaporites (Upper Permian, Lopingian Series).

YABEINA OCCURRENCES IN THE GUADALUPE MOUNTAINS AREA

We have identified specimens of *Yabeina texana* in the Guadalupe Mountains area from the numbered localities shown on Figure 2.22 and listed below. One of the authors (Nestell) recently found *Yabeina texana* at several localities in coeval strata in the Apache Mountains, but the genus is yet to be found in coeval strata in the Glass Mountains.

1. Basal 15 feet of the Lamar Limestone Member, south side of Bear Canyon, NW1/4 SE 1/4 sec. 14, Blk. 65, Twp. 1 S, T. & P. R. R. Survey, Culberson Co., TX, Location is 31°57'32.43"N, 104°45'57.023"W (WGS 1984). See also Skinner and Wilde (1955).
2. Basal part of the Lamar Limestone Member, south wall of McKittrick Canyon, at base of cliff just northwest of the visitor center. A single *Yabeina* specimen was found in a thin, pyrite-bearing, silty packstone bed at the base of the Lamar (sample WT-61-G111, section 7; Tyrrell, 1962). Locality is in sec. 12, Blk. 65, Twp 1, Culberson Co., TX; approximate location is 31°58'45"N,

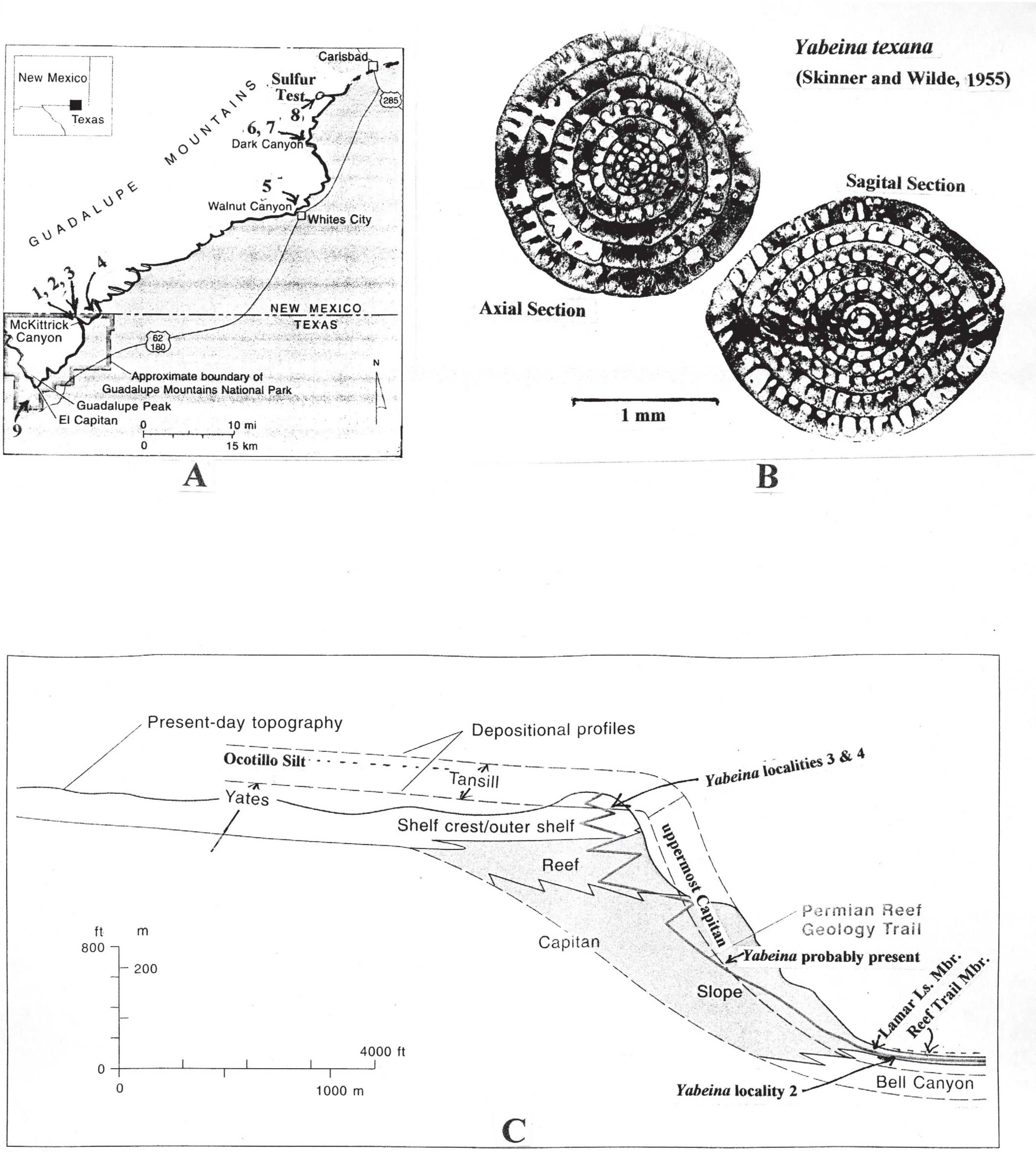


FIGURE 2.22. (A) Index map modified from Bebout, et al. (1993) showing localities where *Yabeina texana* has been found in the basal parts of the Tansill Formation and Lamar Limestone Member. The Capitan Reef trend (solid black line), separates shallow water shelf deposits on the Northwest Shelf from deep water Delaware Basin deposits to the southeast. The numbered localities are listed and described in the text. Localities 3, 5, 8 and 9 are reported here for the first time.
(B) The tiny *Yabeina texana* is a coarse sand-sized, almost spherical fusulinid first described from the basal part of the Lamar Limestone Member in Bear Canyon (our locality 1). The axial and sagittal sections of the paratypes are reproduced from Skinner and Wilde (1955).
(C) Cross section of stratigraphic relationships on the north wall of McKittrick Canyon (modified from Bebout, et al., 1993). *Yabeina texana* has been found in the basal part of the Lamar Limestone Member on the south wall (locality 2) and in the basal part of the Tansill Formation near the top of the Permian Reef Geology Trail (locality 3) and nearby in New Mexico on the third spur northeast of the mouth of McKittrick Canyon (locality 4).

104°45'25"W.

3. Basal Tansill along the Permian Reef Geology Trail, McKittrick Canyon. Numerous *Yabeina texana* were noted but not collected by Tyrrell on top of a ledge about 10 feet above the base of the Tansill Formation, close to trail stop 27. *Yabeina texana* also was found in float nearby and deposited in collections of Gorden Bell, National Park Service. The approximate location is 31°59"50"N, 104°45'35"W.

4. Numerous specimens of *Yabeina texana* were collected in the basal 10 feet of the Tansill Formation on the third spur northeast of McKittrick Canyon (sample WT-61-154, section 5, Tyrrell, 1962). This New Mexico locality is just north of the park boundary and in thick bedded ledges of lime grainstone and packstone in SE1/4 sec. 36, T26S, R21E, Eddy Co., NM. Approximate location is 32°00'10"N. and 104°45'20"W.

5. Specimens of *Yabeina texana* were identified by Gorden Bell in the basal part of the Tansill Formation near stream level along the northeast wall of Walnut Canyon near the entrance to Carlsbad National Park. These strata are in the lower sub-tidal portion of the highstand system tract in HFS G-25 of Rush (2002). Rush did not find *Yabeina texana* in his collections and identified this HFS as coincident with the appearance of *Codonofusiella extensa*. This occurrence is in SW1/4 SE1/4 sec. 27, T24S, R25E., Eddy Co., NM. Location is 32°10'55.483"N, 104°22'56.981"W (WGS 1984).

6. Numerous *Yabeina texana* occur in the basal beds of the Tansill Formation on the north wall of Dark Canyon in yellowish grainstone. Although dolomitized, the *Yabeina texana* here are well preserved and commonly have algal (?) coatings (sample WT-61-271, section 3, Tyrrell, 1962) and were identified by Amoco paleontologist Don Lokke in 1961. In weakly cemented portions, unbroken specimens locally have the appearance of minute basketballs because of roundness and surface indentations of septa. Located in SE1/4 sec. 23, T23S, R25E, Eddy Co., NM.; approximately at 32°17'10"N, 104°21'42"W.

7. The Amoco # 2 Dark Canyon research core drilled in 1968 and now deposited at New Mexico Bureau of Geology and Mineral Resources, Socorro, NM, includes *Yabeina texana* in the basal part of the Tansill Formation from 390 to 399 feet. This site is about 1.0 mile west of the mouth of Dark Canyon on top of the north wall near the jeep road and 0.5 mile north of measured section 3 of Tyrrell (1962). Location is SE1/4 NE1/4 sec. 23, T23S, R25E at approximately 32°17'34"N and 104°21'40"W, Eddy Co., NM (see cross section in Tyrrell, et al., 1978; Tyrrell, 2002). Fusulinids and lithology were studied by G. Verville, G. Sanderson, K. Mesolella, and W. Tyrrell (Amoco Research). Lithology of the core also was examined by Parsley (1988). The fusulinids and nonfusulinid foraminifers were studied recently by Nestell and Nestell (2002, 2006).

8. Tyrrell identified one specimen of *Yabeina texana* in cuttings from the basal part of the Tansill Formation at 470 feet in the Jefferson Lake Sulfur Co. # 28-1 Windham, a sulfur test located at the mouth of Little McKittrick Draw SE1/4 SW1/4 NW1/4 sec. 28, T22S, R26E, Eddy Co., NM, at approximately 32°21'50"N, 104°17'40"W.

9. In January, 2006, Merlynd Nestell found scarce *Yabeina texana* in three superposed beds from the basal 3 feet of the Lamar Limestone Member in the eastern edge of the Patterson Hills. They are associated with abundant *Codonofusiella extensa* and occur in thin, iron-stained, silty grainstone debris flows separated by thin siltstone layers. Outcrop samples were collected by M. Nestell, G. Nestell, and G. Bell. Location is in Culberson Co., TX at 31°49'51.304"N, 104°52'05.705" W (WGS 1984).

REFERENCES:

Bebout, D.G., Kerans, C. and Harris, P.M., 1993, Introduction, *in* Bebout, D.G. and Kerans, C. eds., Guide to the Permian Reef Geology Trail, McKittrick Canyon, Guadalupe Mountains National Park, West Texas: Bureau of Economic Geology, University of Texas at Austin, Guidebook 26, p. 1-4.

Brown, A., 1996, Position of the *Polydiexodina* last-occurrence datum in Guadalupian Strata: revised correlations at McKittrick Canyon, *in* DeMis, W.D., and Cole, A.G., eds., The Brushy Canyon Play in Outcrop and Subsurface: Concepts and Examples: Permian Basin Section, Society of Economic Paleontologists and Mineralogists, Guidebook, Publication 96-38, p. 75 -83.

Kerans, C., and Tinker, S.W., 1999, Extrinsic stratigraphic controls on development of the Capitan Reef Complex, *in* Saller, A.H., Harris, P.M., Kirkland, B.L., and Mazzulo, S.J., eds., Geologi□

King, P.B., 1948, Geology of the Southern Guadalupe Mountains, Texas: U. S. Geological Survey Professional Paper 215, 183 p.

Nestell, M.K. and Nestell, G.P., 2002, Late Capitanian foraminifers from Dark Canyon, Guadalupe Mountains, New Mexico and their correlative potential, *in* Hunt, T.J. and Lufholm, P.H., eds.; The Permian Basin: Preserving Our Past – Securing Our Future: West Texas Geological Society Publication 02-111, p. 9-10.

Nestell, G.P. and Nestell, M.K., 2006, Middle Permian (Late Guadalupian) foraminifers from Dark Canyon, Guadalupe Mountains, New Mexico: Micropaleontology, (in press)

Newell, N.D., Rigby, J.K., Fischer, A.G., Whiteman, A.J., Hickox, J.E. and Bradley, J.S., 1953, The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico: San Francisco, W.F. Freeman, 236 p.

Parsley, M.J., 1988, Deposition and diagenesis of a Late Guadalupian barrier-island complex from the middle and upper Tansill Formation (Permian), east Dark Canyon, Guadalupe Mountains, New Mexico [M.S. thesis]: University of Texas at Austin, 245 p.

Rush, J.W., 2001, Sequence architecture of a Late Guadalupian carbonate rimmed shelf, Walnut Canyon, New Mexico [M.S. Thesis]: University of Texas at Austin, 97p.

Skinner, J.W., and Wilde, G.L., 1955, New fusulinids from the Permian of West Texas: Journal of Paleontology, v. 29, p. 927-941.

Tyrrell, W.W., 1962, Petrology and stratigraphy of near-reef Tansill -Lamar strata, Guadalupe Mountains, Texas and New Mexico; *in* West Texas, Roswell and Hobbs Geol. Societies; Permian of the Central Guadalupe Mountains Guidebook: West Texas Geological Society Publication 62-48, p. 59-69 (Reprinted 1964 *in* Geology of the Capitan Reef Complex of the Guadalupe Mountains: Roswell Geological Society Guidebook, p. 66 -82).

Tyrrell, W.W., 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* Friedman, G. M., ed.; Depositional Environments in Carbonate Rocks: Society of Economic Paleontologists and Mineralogists, Special Publication Number 14, p. 80-97.

Tyrrell, W.W., 2002, Atlas of well log cross sections helps relate Permian sequences, Guadalupe Mountains Area, New Mexico and West Texas, p. 217 -233, *in* Hunt, T.J. and Lufholm, P.H., eds., The Permian Basin: Preserving Our Past-Securing Our Future: West Texas Geological Society, Publication 02-111, p. 217-233.

Tyrrell, W.W., Lokke, D.H., Sanderson, G.A. and Verville, G.J., 1978, Late Guadalupian correlations, Permian Reef Complex, West Texas and New Mexico; *in* Austin, G.S.

(compiler), Geology and Mineral Deposits of Ochoan Rocks in Delaware Basin and Adjacent Areas: New Mexico Bureau Mines and Mineral Resources, Circular 159, p. 84-85. (a 15 page 1988 WTGS field trip handout by these authors has an expanded text and includes logs of the two Dark Canyon Amoco Research cores)
Wilde, G.L., 1990, Practical fusulinids zonation: the species concept; with Permian Basin emphasis: West Texas Geological Society Bulletin, v. 29, p. 5-13, 15, 28-34.
Wilde, G.L., 2000, Formal Middle Permian (Guadalupian) Series: a fusulinacean perspective: *in* Wardlaw, B. R., Grant, R. E. and Rohr, D. M., eds., The Guadalupian Symposium: Smithsonian Contributions to the Earth Sciences Number 32, p. 89-100.
Wilde, G.L., Rudine, S.F., and Lambert, L.L., 1999, Formal designation: Reef Trail Member, Bell Canyon Formation, and its significance for recognition of the Guadalupian-Lopingian Boundary, *in* Saller, A.H., Harris, P.M., Kirkland, B.L., and Mazzullo, S.J., eds., Geologic Framework of the Capitan Reef: Society of Economic Paleontologists and Mineralogists, Special Publication Number 65, p. 63-83.

THICKNESS VARIATIONS IN THE LAMAR LIMESTONE AND REEF TRAIL MEMBERS OF THE BELL CANYON FORMATION, NORTHWESTERN DELAWARE BASIN, NEW MEXICO AND WEST TEXAS.

Willis W. Tyrrell, Jr.[1], John A. Diemer[2], Gorden L. Bell, Jr[3], and Richard J. Bichsel[4]
[1]Consulting Geologist, 5718 Bentway Dr., Charlotte, NC 28226, wtyrrell@carolina.rr.com
[2]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, jadiemer@uncc.edu
[3]Guadalupe Mountains National Park, HC 60, Box 400, Salt Flat, TX 79847, Gorden_Bell@nps.gov
[4]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, rjbichsel@uncc.edu

INTRODUCTION

In the Delaware Basin, the uppermost Guadalupian Series (Capitanian Stage) includes the Lamar Limestone Member and the overlying Reef Trail Member of the Bell Canyon Formation. They underlie the Castile Formation (Ochoa Group) which is traditionally considered the oldest unit of the Lopingian Series. The Lamar Limestone and Reef Trail Members are included in two high frequency sequences (HFS) designated G-27 and G-28 by Kerans and Tinker (1999). We prefer to call the G-27 HFS the **Lower Tansill-Lamar HFS**, and G-28 HFS the **Upper Tansill-Reef Trail HFS**. Especially fine fusulinid zonation establishes correlation of the lower Tansill Formation to the Lamar Limestone Member of the Bell Canyon Formation and of the upper Tansill Formation (including the Ocotillo Silt Member) on the shelf to the basinal Reef Trail Member (Tyrrell, 1969; Tyrrell, et al., 1978; Wilde, 1990). In the subsurface both the Lamar Limestone Member and the Reef Trail Member generally are easily identified on wireline logs. This short paper concerns thickness variations of the wireline log-defined limestone interval of the Lamar Limestone Member between the Lamar Escarpment outcrop and the Carlsbad area. We also emphasize that most of the so-called "Lamar Lime" or "Delaware Lime" in the central Delaware Basin is equivalent to the Reef Trail Member.

LAMAR LIMESTONE MEMBER

The outcropping limestone part of the Lamar HFS has been well studied (e.g., King, 1948; Newell, et al., 1953; Tyrrell, 1962, 1969; Babcock, 1977; Brown and Loucks, 1993). The basinal Lamar Limestone Member is considered the highstand equivalent of the lower Tansill Formation (below the Ocotillo Silt Member) on the shelf. The Lamar Limestone Member contains *Yabeina* at the base and *Paradoxiella* and *Reichelina* in the main part of the unit (Skinner and Wilde, 1955; Tyrrell, 1969; Tyrrell, et al., 1978; Wilde, 1999) but these small "Tethyan" fusulinids are rare or absent away from the basin margin. The lowstand siliciclastics between the Lamar Limestone Member and the underlying McKittrick Canyon Limestone Member include the locally productive Ramsey sandstone of the Delaware Basin. The Lower Tansill-Lamar HFS includes all the lowstand sandstone below the Lamar Limestone Member and above the McCombs Limestone Member. We tentatively interpret the thin McKittrick Canyon Limestone Member as the top of a cycle within this HFS.

Northeast of the Lamar Escarpment the basinal Lamar Limestone and Reef Trail members have been penetrated by more than the 400 control wells shown on Figure 2.23. In addition, we had access to a few Culberson Co., TX wells. The Lamar Limestone Member gradually thins basinward (Fig. 2.23) and its characteristic wireline log-defined limestone portion pinches out approximately 20 miles basinward of the basin margin (Tyrrell, 1969).

"Clean", low radioactive limestone beds in the Lamar Limestone Member can be identified on Gamma Ray logs and the interval containing these limestone beds greater than one foot thick is mapped on Figure 2.23. At the basin margin the senior author examined cable tool cuttings from the Turner and DeVito #1 DeVito (SW1/4 SE1/4 sec.12, T23S, R25E). Here the Lamar Limestone Member is 355 feet thick and it consists mostly of lime mudstone and skeletal wackestone interpreted as turbidite deposits. In the central basin where distal limestone beds have pinched out, their equivalent is a thin, organic-rich siltstone or impure limestone condensed section, the log character of which is a radioactive kick on Gamma Ray logs (see Tyrrell, 1969, fig. 17). An example of this is in University of New Mexico #1 Phillips core (Culberson Co., TX, Sec. 3, Blk. 110, PSL) where Anderson, et al. (1972) found this condensed section at the base of their Claystone II unit. Here it consists of only 2 feet of laminated organic-rich limey claystone with only a few inches of lime mudstone. Thus, the equivalent of the Lamar Limestone Member in the central basin comprises only the lowermost part of the so-called "Lamar Lime". Published descriptions of this more distal

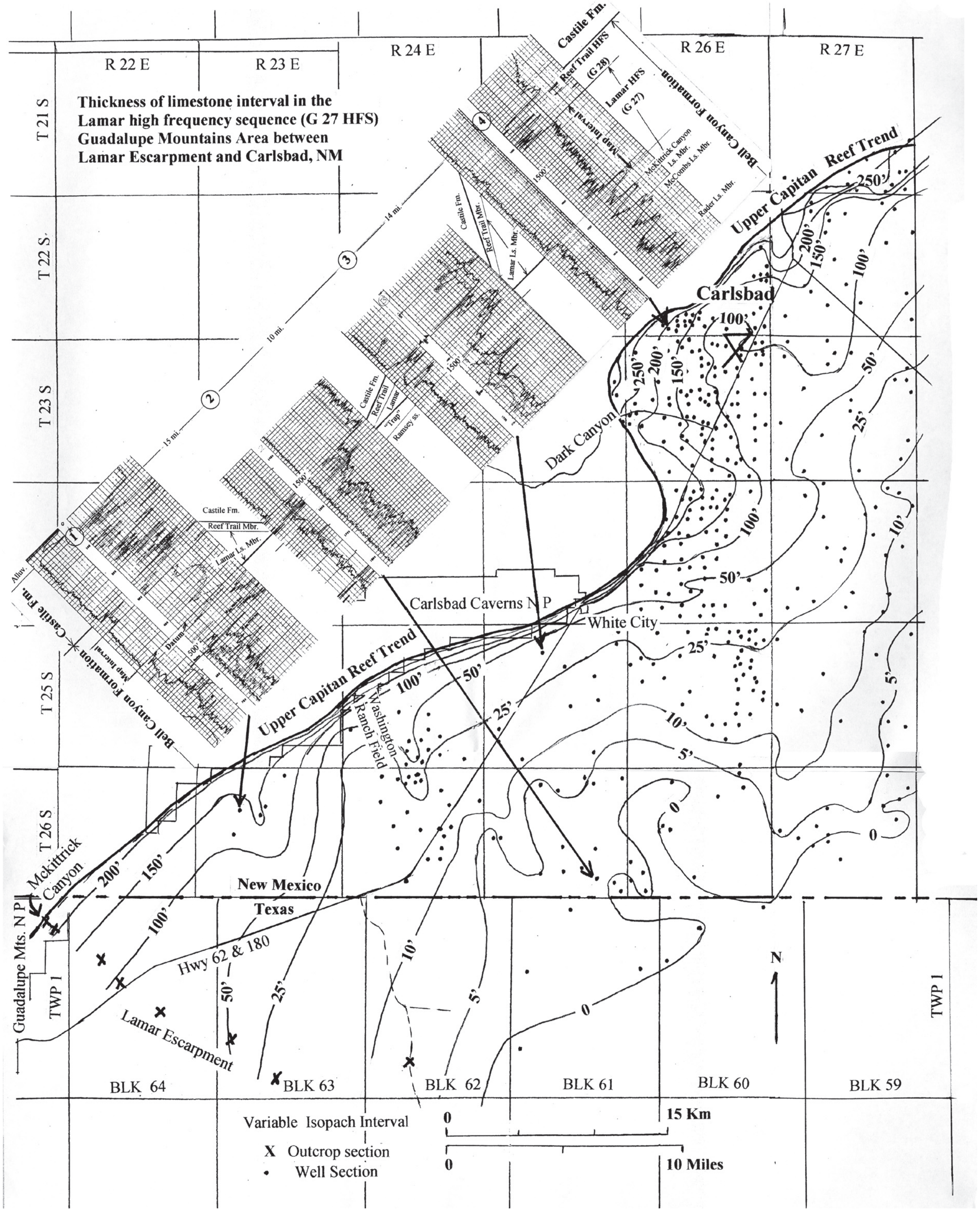


FIGURE 2.23. Thickness variations of the Lamar limestone interval in the area southwest of Carlsbad, NM and a cross-section showing typical wireline log character of the Reef Trail and Lamar Limestone members. Locations of wells 1 to 4 are indicated by arrows. The large "X" south of Carlsbad is runways at the Cavern City airport. Note the variation in thickness along strike between Carlsbad and the Lamar Escarpment where the more resistant Lamar Limestone Member forms a cuesta. The Lamar limestone interval pinches out about 20 miles from the basin margin. It is thinner in the Whites City area than in the embayment between Dark Canyon and Carlsbad. See discussion in text for probable reasons. The four representative wireline logs show the mapped interval, interpreted stratigraphy and typical log character of the Reef Trail and Lamar Limestone members of the Bell Canyon Formation in the northwestern Delaware Basin. Datum is the base of the Lamar Limestone Member. Wells used are: (1) Amoco # 1 Federal "BH" (API 30-15-23355); (2) Amoco # 1 Federal "BQ" (API 30-015-23442); (3) Cabal Energy # 1 High Hog "9" Federal (API 30-015-33462), and (3) Dreyfus Natural Gas # 2 State "EV" (API 30-015-27559). These wireline logs, as well as many other New Mexico logs, can be viewed full scale online at the New Mexico Energy, Minerals, and Natural Resources Department; Division of Oil Conservation website.

basinal equivalent to the Lamar Limestone Member are in papers describing oil fields producing from the underlying Ramsey sand (cf. Watson, 1979; Dutton, et al., 1999). We are not aware of any oil fields producing from the Lamar Limestone Member.

The isopach map (Fig. 2.23) includes only the “clean”, low radioactivity, Gamma Ray log-defined limestone interval in the Lamar Limestone Member. No doubt there were variations in the amount of carbonate debris shed from the prograding Capitan Reef during lower Tansill time. Furthermore, reef growth probably was not uniform along strike. The Lamar limestone interval is relatively thick near McKittrick Canyon, the Washington Ranch Field area and between Dark Canyon and Carlsbad. In the latter area, thickening of the Lamar limestone interval may be due to the embayment of the Capitan Reef trend such that more carbonate debris and turbidites could be derived from a range of directions. In contrast, the Lamar limestone interval thins rapidly basinward between the Washington Ranch Field and Whites City, possibly due to distance from thicker lobes of reef-derived carbonate turbidites.

REEF TRAIL MEMBER

The Reef Trail Member was named by Wilde, et al. (1999) to replace the “post-Lamar beds” of King (1948). Generally, in the central Delaware Basin, the Reef Trail Member makes up most of what is commonly called the “Delaware Lime”, “Lamar Lime”, “Black Lime”, or “Lamar Shale”. At the basin margin the Reef Trail Member includes limestone debris or thin limestone beds in its upper part (see wells 1 and 4, Fig. 2.23) but these thin rapidly basinward. Away from the basin margin the Reef Trail Member is relatively uniform in thickness and generally consists of an upper, thin, organic-rich siltstone or impure limestone unit that produces a radioactive kick on Gamma Ray logs and this is underlain by 20 to 30 feet of nonporous to low porosity, very fine-grained sandstone or siltstone (Watson, 1979). Therefore we did not map the Reef Trail Member which has a relative uniform thickness in the central basin. We interpret the thin, uppermost, radioactive unit to be the highstand condensed section equivalent to basin margin limestone and to the upper Tansill Formation carbonate above the Ocotillo Silt Member on the shelf. The main siltstone-sandstone portion of the Reef Trail Member in the central basin is considered a lowstand deposit equivalent to the Ocotillo Silt or its basal bypass surface. The upper Tansill Formation on the shelf and the Reef Trail Member in the basin are in the zone of *Paraboultonia* (Skinner and Wilde, 1954) which traditionally has marked the top of the Guadalupian Series. Wilde, et al., (1999) recommended placing the Reef Trail Member in the basal part of the overlying Lopingian Series but this has been questioned by other workers (e. g. Nestell and Nestell, 2002). We are not aware of any oil production from the Reef Trail Member which makes up most of the so called “Lamar Lime” in the basin.

CONCLUSIONS

The highstand Lamar Limestone Member thins irregularly basinward from over 300 feet at the basin margin to less than one foot about 20 miles into the basin. Over most of the central basin it is represented by a thin condensed section of radioactive, organic-rich siltstone with local impure limestone at the base of the so called “Lamar Lime”. The Reef Trail Member is relatively uniform in thickness (30–20 feet) away from the basin margin and makes up most of the “Lamar Lime” in the central Delaware Basin.

REFERENCES

Anderson, R.Y., Dean, W.E., Kirkland, D.W., Snider, H.L., 1972, Permian Castile varved evaporite sequence, west Texas and New Mexico: Geological Society of America Bulletin, v. 83, p. 59-86.

Babcock, L.C., 1977, Life in the Delaware Basin-the paleoecology of the Lamar Limestone; *in* Hileman, M.E. and Mazzullo, S.J., eds., Upper Guadalupian Facies, Permian Reef Complex: Permian Basin Section, Society of Economic Paleontologists and Mineralogists, Publication 77-16, p. 357-389.

Brown, A. and Loucks, R.G., 1993, Toe of slope and geology loop trail; *in* Bebout, D.G. and Kerans, C., eds.; Guide to the Permian Reef Geology Trail, McKittrick Canyon, Guadalupe Mountains National Park, West Texas: Bureau Economic Geology, University of Texas at Austin, Guidebook 26, p. 5-13.

Dutton, S.P., Barton, M.D., Asquith, G.B., Malik, M.A., Cole, A.G., Gogas, J., Guzman, J.I., and Clift, S.J., 1999, Geologic and engineering characterization of turbidite reservoirs, Ford Geraldine Unit, Bell Canyon Formation, West Texas: Bureau Economic Geology, University of Texas at Austin, Report of Investigations Number 255, 88p.

Kerans, C., and Tinker, S.W., 1999, Extrinsic stratigraphic controls on development of the Capitan Reef Complex, *in* Saller, A.H, Harris, P.M., Kirkland, B.L. and Mazzulo, S.J., eds., Geologic

Nestell, M.K. and Nestell, G.P., 2002, Late Capitanian foraminifers from Dark Canyon, Guadalupe Mountains, New Mexico and their correlative potential, *in* Hunt, T. J. and Lufholm, P. H., eds., The Permian Basin: Preserving Our Past–Securing Our Future: West Texas Geological Society, Publication 02-111, p. 9 -10.

Newell, N.D., Rigby, J.K., Fischer, A.G., Whiteman, A.J., Hickox, J.E. and Bradley, J.S., 1953, The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico: San Francisco, W. F. Freeman, 236 p.

King, P.B., 1948, Geology of the southern Guadalupe Mountains, Texas: USGS Professional Paper 215, 183 p.

Skinner, J.W., and Wilde, G.L., 1954, The fusulinid subfamily Boultoniinae: Journal of Paleontology, v. 28, p. 434-444.

Skinner, J.W., and Wilde, G.L., 1955, New fusulinids from the Permian of West Texas: Journal of Paleontology, v. 29, p. 927-941.

Tyrrell, W.W., 1962, Petrology and stratigraphy of near-reef Tansill-Lamar strata, Guadalupe Mountains, Texas and New Mexico; *in* West Texas, Roswell and Hobbs Geological Societies; Permian of the Central Guadalupe Mountains Guidebook, West Texas Geological Society, Publication 62–48, p. 59-69 (Reprinted 1964 *in* Geology of the Capitan Reef Complex of the Guadalupe Mountains: Roswell Geol. Soc., Guidebook, p. 66-82).

Tyrrell, W.W., 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* Friedman, G. M., ed., Depositional Environments in Carbonate Rocks: Society of Economic Paleontologists and Mineralogists, Special Publication Number 14, p. 80 -97.

Tyrrell, W.W., Lokke, D.H., Sanderson, G.A., and Verville, G.J., 1978, Late Guadalupian correlations, Permian Reef Complex, West Texas and New Mexico; *in* Austin, G.S. (compiler), Geology and Mineral Deposits of Ochoan Rocks in Delaware Basin and Adjacent Areas: New Mexico Bureau of Mines and Mineral Resources, Circular 159, p. 84-85 (an expanded text and logs of the Dark Canyon Amoco Research Tansill Formation-Capitan Formation cores are in a 15 page handout prepared for the 1988 West Texas Geological Society fieldtrip).

Watson, W.G., 1979, Inhomogeneities of the Ramsey Member of the Permian Bell Canyon Formation, Geraldine Ford Field, Culberson and Reeves Counties, Texas, *in* Sullivan, N.M., ed., Guadalupian Delaware Mountain Group: Permian Basin Section, Society of Economic Paleontologists and Mineralogists, Publication 79-18, p. 2-38.

Wilde, G.L., 1990, Practical fusulinids zonation: the species concept; with Permian Basin emphasis: West Texas Geological Society Bulletin, v. 29, p. 5-13; 15; 28-34.

Wilde, G.L., Rudine, S.F., and Lambert, L.L., 1999, Formal designation: Reef Trail Member, Bell Canyon Formation, and its significance for recognition of the Guadalupian-Lopingian Boundary; *in* Saller, A.H., Harris, P.M., Kirkland, B.L., and Mazzullo, S.J., eds., Geologic Framework of the Capitan Reef: Society of Economic Paleontologists and Mineralogists, Special Publication Number 65, p. 63-83.

REGIONAL MAP OF HIGHSTAND PHASE OF LOWER SEVEN RIVERS - HEGLER HIGH FREQUENCY SEQUENCE (HFS) IN SOUTHEAST NEW MEXICO

Willis W. Tyrrell, Jr.[1], John A. Diemer[2], Gorden L. Bell, Jr.[3], and Richard J. Bichsel[4]

[1]Consulting Geologist, 5718 Bentway Dr., Charlotte, NC 28226, wtyrrell@carolina.rr.com

[2]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, jadiemer@uncc.edu

[3]Guadalupe Mountains National Park, HC 60, Box 400, Salt Flat, TX 79847, gorden_bell@nps.gov

[4]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, rjbichsel@uncc.edu

INTRODUCTION

Outcrop studies in the Guadalupe Mountains over the last 70 years have established many shelf-to-basin correlations in the upper Wordian and Capitanian Stages of the Guadalupian Series (e. g. Garber, et a., 1989). In addition, outcrop sequence analysis recognized 3 composite sequences (CS-12, CS-13, and CS-14) and 12 high frequency sequences (HFS G-17 through G-28) equivalent to the Capitan Formation (Kerans and Tinker,1999). This paper concerns only the lower of these: the HFS G-17 in outcrop and subsurface. We prefer to call it the **Lower Seven Rivers-Hegler HFS** which includes the lower Seven Rivers Formation on the Northwest shelf, the lowermost Capitan Formation on the shelf margin, and the lowstand uppermost sandstone of the Cherry Canyon Formation together with the overlying highstand Hegler Limestone Member of the Bell Canyon Formation in the Delaware Basin (see outcrop sections A and D and logs B and C of Fig. 2.24). Because of good wireline log character, we are able to tie published outcrop-defined HFS in CS 12 to wireline log-defined equivalent HFS in the subsurface across the Northwest Shelf and into the northern Delaware Basin. Control consists of over 1000 well logs and 25 published measured sections of outcrops. Thus this paper presents, for the first time, a regional map of the highstand phase of the Lower Seven Rivers-Hegler HFS. A map of the lowstand phase of this HFS was presented elsewhere (Tyrrell and Diemer, 2004).

On the Northwest Shelf there is minor oil production from highstand carbonates that make up most of the Seven Rivers Formation. In the Delaware Basin the highstand carbonates and siltstone/sandstone of the Hegler Limestone Member are rarely productive but locally (i. e. Indian Draw Field near Carlsbad) there is production from channel sandstones in the uppermost Cherry Canyon Formation which is the lowstand portion of the Lower Seven Rivers-Hegler HFS.

On the west face of the Guadalupe Mountains the shelfal Lower Seven Rivers Formation grades into the lower Capitan reef and talus which in turn grades into the basinal Hegler Limestone Member of the Bell Canyon Formation (King, 1948, Newell, et al., 1953). These units are the highstand phase of the Lower Seven Rivers-Hegler HFS and they mark the base of the zone of *Polydiexodina* (Newell, *et al*., 1953; Wilde, 1990; 2000). Thus the correlation of these units is well established.

The Northwest Shelf and the Delaware Basin cross sections on Figure 2.24 show our interpretation of the stratigraphic relationships of the Lower Seven Rivers-Hegler HFS. On the Northwest Shelf the Lower Seven Rivers is overlain by the Bowers Sand zone, an established subsurface marker (Tyrrell and Diemer, 2003). Although not named as such, the Bowers Sand zone also can be recognized in published measured sections in North McKittrick Canyon (Hurley, 1978; Tinker,1998), in Bear Canyon (Hayes, 1964), and in Rocky Arroyo (Sarg, 1989). In the basin the outcropping Hegler Limestone Member can be correlated to the widespread subsurface “Two Finger Limestone” marker (Tyrrell and Diemer, 2003). These outcrop-to-subsurface relationships permit regional mapping of the highstand portion of the Lower Seven Rivers- Hegler HFS (Fig. 2.24). The siliciclastic lowstand part of this HFS in the northern Delaware Basin also can be mapped but is not shown here (cf. Tyrrell and Diemer, 2004).

HIGHSTAND THICKNESS MAP

Our well control is concentrated in New Mexico where the New Mexico Bureau of Geology and Mineral Resources supplied some well logs. Considerable well data is now available online from the Oil Conservation Division, New Mexico Energy, Minerals and Natural Resources Department. As shown by the control points on the map we have concentrated our mapping in Eddy County and have done only reconnaissance mapping in Lea County. We have access to relatively few wells in Texas.

Figure 2.24 shows the regional trends and thickness variations of the highstand portion of the Lower Seven Rivers-Hegler HFS. On the Northwest Shelf the mapped interval is the lower Seven Rivers Formation which is underlain by the Shattuck Sand Member (“Artesia Red Sand”) and overlain by the Bowers Sand zone (see outcrop section A and well B). The lower Seven Rivers

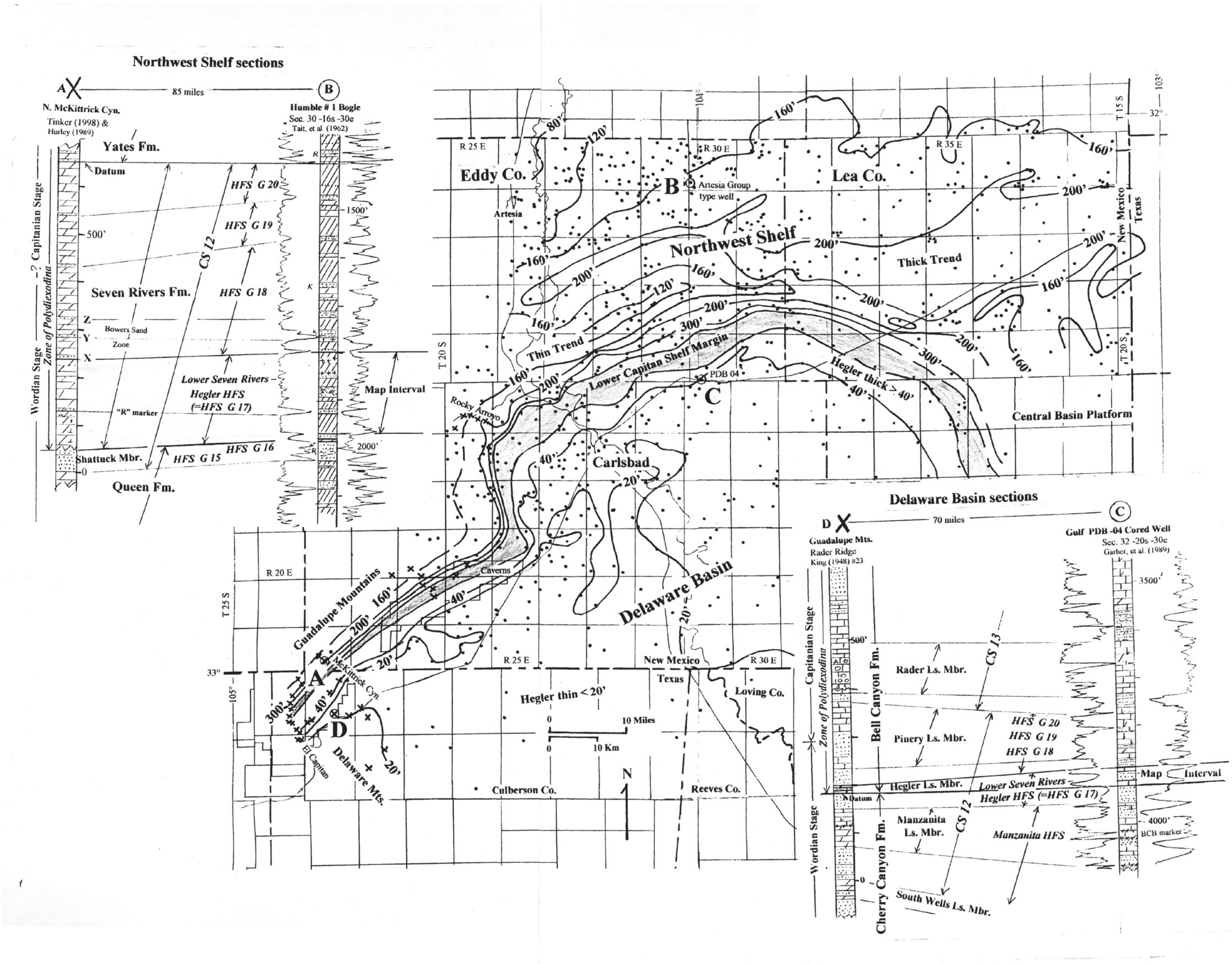


FIGURE 2.24. Isopach map of the highstand phase of the Lower Seven rivers-Hegler HFS in southeastern New Mexico. Note the thickening toward the shelf margin of the lower Seven Rivers Formation and the rapid thinning of the coeval Hegler limestone Member into the basin. Cross section A-B shows typical outer shelf (A) and inner shelf (B) stratigraphy of the lower Seven Rivers Formation. Outcrop section A in North McKittrick Canyon is from Tinker (1998) and Hurley (1989) who named siliciclastic beds above the lower Seven Rivers zones X, Y, and Z (Bowers Sand zone of subsurface). The Shattuck Member of the Queen Formation underlies the Seven Rivers Formation and its uppermost part may be the lowstand and transgressive phases of the Lower Seven Rivers-Hegler HFS.

Delaware Basin cross section D-C shows typical outcrop and log character of the basinal Hegler Limestone Member of the Bell Canyon Formation. Well C is the PDB 04 cored research well (Garber, et al., 1989) and measured section D is from Rader Ridge, Guadalupe Mountains (section 23 of King, 1948). In the Delaware Basin the Lower Seven Rivers-Hegler HFS includes the lowstand uppermost sandstone/siltstone of the Cherry Canyon Formation which overlies the regional Mz 5 cycle of the Manzanita HFS (Tyrrell, et al., 2004). Note the typical "Two Finger Limestone" signature of the highstand Hegler Limestone Member in outcrop and the subsurface.

Formation is mostly dolomite in the middle shelf, shelf crest and part of the outer shelf facies. It is mostly evaporite in the inner shelf facies (see well B). Well B is the subsurface type well for the Artesia Group (Tait, et al., 1962) which includes the Seven Rivers Formation. The carbonate-to-evaporite transition in the lower Seven Rivers Formation has been documented in Rocky Arroyo outcrops west of Carlsbad (Bates, 1942; Sarg, 1989). Note the regional "R" marker of Tyrrell and Diemer (2003) which subdivides the lower Seven Rivers HFS into two cycles. The siliciclastics of the Bowers Sand zone and the "R" marker form more vegetated layers in outcrop (for example see Jerry Lucia's photo of the north end of North McKittrick Canyon, fig. 6 in Bebout, et al., 1993; fig. 6 in Tyrrell and Diemer, 2003).

In the mapped area, thickness of the lower Seven Rivers Formation on the shelf ranges from 80 feet north of Artesia, NM to over 300 feet near its transition into the lower Capitan Formation. Note the thick and thin trends which may reflect depositional topography, older structure, or facies change and differential compaction.

In most Delaware Basin outcrops and wells the Hegler Limestone Member ("Two Finger Limestone") consists of two thin limestone beds separated by equally thin siltstone/sandstone (see outcrop D and well C). As shown on Figure 2.24, thickness of the Hegler Limestone Member exceeds 40 feet only along the basin margin. In the northwestern part of the basin, south of the Guadalupe Mountains, the Hegler Limestone Member ("Two Finger Limestone") still consists of limestone-siltstone-limestone but is thin - generally less than 20 feet thick. In the more distal basin the limestone beds of the Hegler Limestone Member commonly grade into radioactive, organic-rich silty condensed sections.

CONCLUSIONS

Regional mapping of some Guadalupian HFS from the shelf to the basin is possible (1) where regional siliciclastic lowstand markers on the shelf can be recognized on wireline logs and used to define HFS, and (2) where correlative basinal limestones or equivalent condensed sections can be recognized on wireline logs and overlie lowstand siliciclastic units. The Lower Seven Rivers–Hegler HFS is ideal and both the highstand and lowstand portions can be mapped regionally. Such maps add much to the regional understanding of detailed outcrop-defined sequence stratigraphy. However, well control in the basin generally is not close enough to map narrow channels such as the oil reservoir in the Indian Draw Field west of Carlsbad, NM in the lowstand portion of the Lower Seven Rivers-Hegler HFS (see Cromwell, 1979).

REFERENCES:

Bates, R.L., 1942, Lateral gradation in the Seven Rivers Formation, Rocky Arroyo, Eddy County, New Mexico: Bulletin of American Association of Petroleum Geologists, v. 26, p. 80 -99.

Bebout, D.G., Kerans, C., and Harris, P.M., 1993, Introduction, *in* Bebout, D.G. and Kerans, C., eds., Guide to the Permian Reef Geology Trail, McKittrick Canyon, Guadalupe Mountains National Park, West Texas: Texas Bureau of Economic Geology, University of Texas at Austin, Guidebook 26, 48p.

Cromwell, D.W., 1979, Indian Draw Delaware Field, a model for deeper Delaware sand exploration, in Sullivan, N. M., ed., Guadalupian Delaware Mountain Group of West Texas and So

Garber, R.A., Grover, G.A. and Harris, P.M., 1989, Geology of the Capitan shelf margin - subsurface data from the northern Delaware Basin, *in* Harris, P.M. and Grover, G.A., eds., Subsurface and Outcrop Examination of the Capitan Shelf Margin, Northern Delaware Basin: Society of Economic Paleontologists and Mineralogists, Core Workshop 13, 269 p.

Hayes, P.T., 1964, Geology of the Guadalupe Mountains, New Mexico: U. S. Geological Survey Professional Paper 446, 69 p.

Hurley, N.F., 1978, Facies mosaic of the Lower Seven Rivers Formation (Permian), North McKittrick Canyon, New Mexico, *in* Harris, P.M. and Grover, G.A., eds., Subsurface and Outcrop Examination of the Capitan Shelf Margin, Northern Delaware Basin: Society of Economic Paleontologists and Mineralogists, Core Workshop 13, p. 325-346.

Kerans, C., and Tinker, S.W., 1999, Extrinsic stratigraphic controls on development of the Capitan Reef Complex, *in* Saller, A.H., Harris, P.M., Kirkland, B.L., and Mazzulo, S.J., eds., Geologi

King, P.B., 1948, Geology of the Southern Guadalupe Mountains, Texas: U.S. Geological Survey Professional Paper 215, 183 p.

Newell, N.D., Rigby, J.K., Fischer, A.G., Whiteman, A.J., Hickox, J.E. and Bradley, J.S., 1953, The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico: San Francisco, W. F. Freeman, 236 p.

Sarg, J.R., 1989, Stratigraphy and sedimentology of the back-reef upper Queen-lower Seven Rivers Formation strata, Goat Seep-Capitan reef complexes (Middle-Late Guadalupian, Permian) southeast New Mexico, *in* Harris, P.M., and Grover, G.A., eds., Subsurface and Outcrop Examination of the Capitan Shelf Margin, Northern Delaware Basin: Society of Economic Paleontologists and Mineralogists, Core Workshop 13, p. 347-352.

Tait, D.B., Ahlen, J.L., Gordon, A., and Scott, G.L., 1962, Artesia Group of New Mexico and Texas: Bulletin of American Association of Petroleum Geologists, v. 46, p. 504-517.

Tinker, S.W., 1998, Shelf-to-basin facies distribution and sequence stratigraphy of a steep-rimmed carbonate margin: Capitan depositional system, McKittrick Canyon, New Mexico and Texas: Journal of Sedimentary Research, v. 68, p. 1146-1174.

Tyrrell, W.W. Jr. and Diemer, J.A., 2003, Relationship of two widespread Guadalupian subsurface marker units (Bowers Sand and Two Finger Limestone) to outcrop-defined sequences in the Guadalupe Mountains, New Mexico and west Texas, *in* Hunt, T.J., and Lufholm, P.H., *eds*. The Permian Basin: Back to Basics, West Texas Geological Society, Publication 03-12, p. 97-118.

Tyrrell, W.W., and Diemer, J.A., 2004, First regional lowstand-highstand maps of a Guadalupian high frequency sequence (Lower Seven Rivers HFS = G 17), northern Delaware Basin area, New Mexico and West Texas (abs): Program with Abstracts, AAPG Southwest Section Meeting, El Paso, TX, March 6-9, 2004.

Wilde, G.L., 1990, Practical fusulinids zonation: the species concept; with Permian Basin emphasis: West Texas Geological Society Bulletin, v. 29, p. 5-3; 15; 28-34.

Wilde, G.L., 2000, Formal Middle Permian (Guadalupian) Series: a fusulinacean perspective: *in* Wardlaw, B.R., Grant, R.E. and Rohr, D.M. eds., The Guadalupian Symposium: Smithsonian Contributions to the Earth Sciences, Number 32, p. 89-00.

DELAWARE (LAMAR) LIMESTONE ROADCUT

Alton Brown
Consultant, 1603 Waterview Drive, Richardson, TX 75080, Altonabrown@yahoo.com

This long roadcut exposes the uppermost carbonate and siltstone units of the Bell Canyon Formation, commonly referred to as the Delaware Lime or Lamar Limestone by subsurface workers (e. g. Tyrell et al., 1978). This location is about 4.5 miles into the basin from the contemporaneous shelf edge. Despite the dark, monotonous appearance of the outcrop, there are a number of geologically significant features to be seen here.

Lithostratigraphy and depositional environments. The basal 20 ft of the Delaware Lime is not exposed in this roadcut. The exposed section is informally divided into 4 units (Figure 2.25). The lower two units are exposed at the western end of the north side of the cut, whereas the upper two units are best exposed and accessible towards the middle of the roadcut.

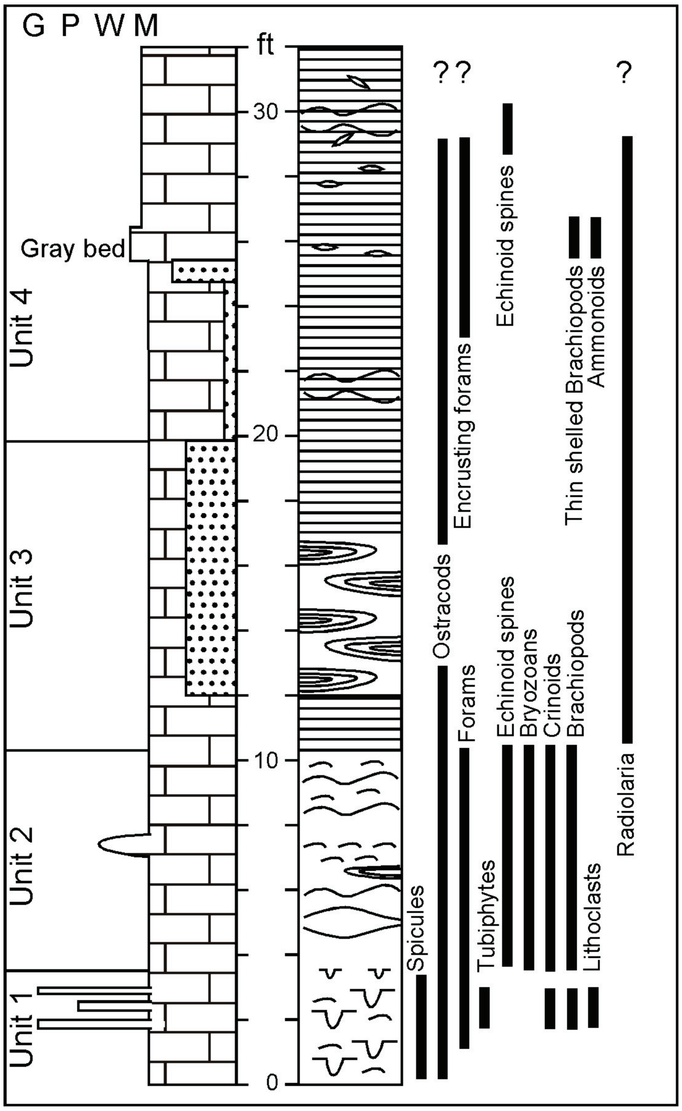


Figure 2.25. Measured section of the Lamar roadcut section.

The lower two units are predominantly homogeneous, sparsely fossiliferous lime mudstones with a relict peloidal fabric. Unit 1 is a medium-bedded, sparsely fossiliferous, burrowed lime mudstone with thin silicified grainstone beds containing lithoclasts, Tubiphytes and fragmented shell debris. Unit 2 is medium- to thin-bedded with no trace fossils. Transported shelly fossils are locally common on bedding planes in Unit 2. Unit 2 also shows significant small-scale contortion consistent with syndepositional incipient slumping. Both Units 1 and 2 are interpreted as basinal sedimentation dominated by low-density turbidity current deposition of carbonate mud and shells derived from the middle slope. Unit 1 deposition was dysaerobic (low-oxygen at seafloor), whereas Unit 2 deposition was anaerobic (lacking oxygen on the sea floor). Grainstones in Unit 1 are turbidites derived from the upper slope.

Units 3 and 4 are predominantly anaerobic, laminated basinal suspension deposits with almost no benthonic fauna and little evidence for turbidity current deposition. Medium silt-sized fossils are common, so thin sections have a grainy look while maintaining a wackestone to mudstone fabric. Unit 3 is a silty limestone to calcareous siltstone with distinctive isoclinal folding (discussed below). Unit 4 contains some calcareous siltstone beds, but it is predominantly suspension-laminated lime wackestone. A distinctive gray laminated bed can be traced along the upper part of the outcrop. It contains flattened ammonoids and brachiopods on bedding planes. Small radiolaria are seen in thin section throughout Units 3 and 4, but they are especially abundant in the gray bed. Small echinoid (?) spines are evident on some bedding planes near the top of the section.

Small gypsum pseudomorphs are present near the top of the section. Although these pseudomorphs have been cited as evidence for high salinity before the onset of Castile deposition, their petrographic characteristics indicate early diagenetic replacement of carbonate below the seafloor instead of seafloor precipitation (Figure 2.26). The base of the Castile Formation projects to a horizon a few meters above the top of this section, so replacement probably occurred soon after initial Castile deposition as evaporitic brines sank through the unconsolidated sediment. Some pseudomorphs contain anhedral celestite.

Correlation to the basin margin. There are two lithostratigraphic datums that may correlate to the shelf-margin section. The burrows in Unit 1 are strikingly similar to the zone of burrows in the upper Lamar limestone exposed near McKittrick and Big Canyons. The burrowed zone was first correlated into the basin by Babcock (1977). The siliciclastic silt in Unit 3 may correlate to the major pulse of basinal siliciclastic siltstone deposition in the basal Reef Trail Member exposed in and just north of McKittrick Canyon (Wilde et al. 1999). This correlation is consistent with Tyrell's wireline log correlations. Both lithostratigraphic markers indicate that Units 1 and 2 correlate to the upper

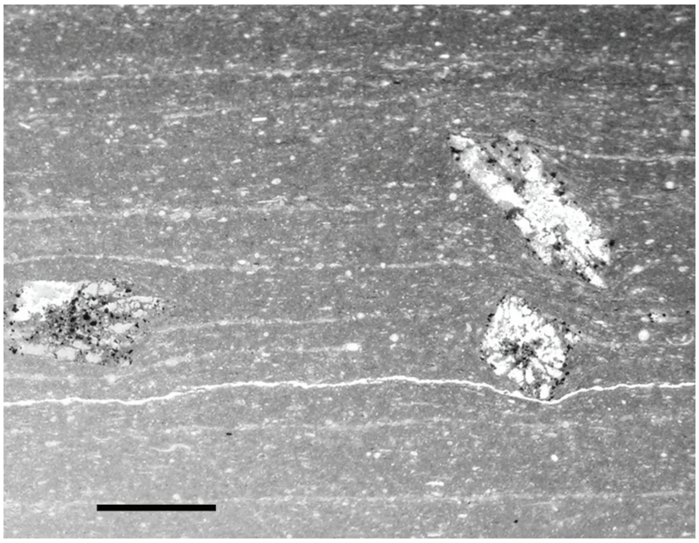


Figure 2.26. Photomicrograph of gypsum pseudomorphs from the upper part of Unit 4. Crystals lack seafloor growth or transport morphology. Pseudomorphs replace layers near the middle of the crystals and preserve an epitaxial growth morphology. Growth is both upwards and downwards. Beds near pseudomorphs wrap around the crystals, indicating pre-compaction origin. Scale bar is 2 mm long.

Lamar Limestone and Units 3 and 4 correlate with the Reef Trail Member of the Bell Canyon Formation.

Structure. Syndepositional and Laramide-aged structures are well exposed on outcrop. The most obvious soft-sediment deformation is the folding in Unit 3. Folds are predominantly isoclinal recumbent folds with distinct upper and lower decollement surfaces. Fold axes are oriented to the SE, parallel to inferred paleoslope. Some axial planes dip slightly basinward, but axial planes are themselves locally deformed by gentle cylindrical folding with a nearly vertical axial surface. The most likely origin for Unit 3 folding is deformation near the base of a large gravity slide moving slowly down slope. Shear near the decollement surface stretches early folds into sheath folds elongated parallel to the downslope transport direction. Absence of complete disruption or liquefaction in the decollement zone may indicate slow slide movement. If this interpretation is correct, then Unit 4 may be transported from a short distance upslope.

Tectonic stylolites and small thrusts indicate about 0.5% to 1 % shortening in a NE-SW direction. These are clearly post-lithification features. The tectonic stylolites are near-vertical planes striking NW with small "teeth" nearly perpendicular to the stylolite surface. Tectonic stylolites are particularly abundant in Unit 4. The thrusts are small listric features that show less than an inch of NE or SW offset. Thrusts are most evident in the middle and upper parts of the outcrop. Strain on some thrusts is accommodated by shortening at the tectonic stylolites, so these features are contemporaneous.

Tectonic stylolites are evident in limestones throughout the Guadalupe Mountains, and both tectonic stylolites and small thrusts with NE shortening occur widely across the Trans-Pecos area. This orientation is strikingly different from interpreted Tertiary basin-range stresses and is most consistent with early Laramide deformation (Price and Henry 1985). Tectonic stylolites showing NE shortening affect middle Cretaceous limestones near Kent, supporting a Late Cretaceous or Early Tertiary (Laramide) origin.

Petroleum potential. Although rocks from the outcrop have a strong petroliferous smell, they all have low organic carbon content (< 0.3 % TOC) and low petroleum source potential. During Bell Canyon deposition, facies containing coarse-grained gravity-flow deposits formed a very narrow fairway close to the basin margin (Brown and Loucks 1993). This outcrop lies basinward of the grain-dominated gravity-flow belt, so almost all carbonates are mud dominated and reservoir potential in the basinal carbonate facies is almost nil. The overlying Castile Formation acts as a regional seal for Bell Canyon petroleum accumulations.

REFERENCES

Babcock, L, 1977, Life in the Delaware Basin: the paleoecology of the Lamar Limestone: in Upper Guadalupian facies, Permian Reef Complex Guadalupe Mountains, N.M. and West Texas; PBS-SEPM 1977 field conference guidebook, PB # 77-16, pp. 357-390.

Brown, A., and R. Loucks, 1993, Influence of Sediment Type and Depositional Processes on Stratal Patterns in the Permian Basin-Margin Lamar Limestone, McKittrick Canyon, Texas: in: *Carbonate Sequence Stratigraphy*: AAPG Memoir 57, p. 133-156.

Price, J. G., and C. D. Henry, 1985, Summary of Tertiary stress orientation and tectonic history of Trans-Pecos Texas: in Structure and Tectonics of Trans-Pecos Texas, West Texas Geological Society publication # 85-81, pp. 149 - 151.

Tyrell, W. W., D. Lokke, G. A. Sanderson, and G. Verville, 1978, Late Guadalupian correlations, Permian reef complex, West Texas and New Mexico,: in: Austin, G. S., ed., Geology and Mineral deposits of Ochoan rocks in Delaware basin and adjacent areas: New Mexico Bureau of Mines and Mineral Resources, Circular 159, p. 84-85.

Wilde, G. L., S. F. Rudine, and L. Lambert, 1999, Formal designation: Reef Trail Member, Bell Canyon formation, and its significance for recognition of the Guadalupian-Lopingian boundary: in Geologic framework of the Capitan Reef, SEPM Special Publication 65, p. 63-83.

THE BENTONITE-BEARING MANZANITA LIMESTONE MEMBER, CHERRY CANYON FORMATION, EXPOSED IN A PATTERSON HILLS ROAD CUT, CULBERSON COUNTY, TEXAS

John A. Diemer[1], Willis W. Tyrrell, Jr.[2], Gorden L. Bell, Jr.[3], David H. Griffing[4]
[1]Department of Geography and Earth Sciences, UNC Charlotte, Charlotte, NC 28223, jadiemer@uncc.edu
[2]Consulting Geologist, 5718 Bentway Dr., Charlotte, NC 28226, wtyrrell@carolina.rr.com
[3]Guadalupe Mountains National Park, HC 60, Box 400, Salt Flat, TX 79847, gorden_bell@nps.gov
[4]Department of Geology, Hartwick College, Oneonta, NY 13820, GriffingD@hartwick.edu

INTRODUCTION

A nearly complete section of the bentonite-bearing Manzanita Limestone Member of the Cherry Canyon Formation is exposed in a road cut in the Patterson Hills six miles south-southwest of El Capitan of the southern Guadalupe Mountains, west Texas (Fig. 2.27). The road cut is located at 31°46’39”N and 104°53’20.7”W near milepost 118 on US Highway 62/180, about 3 miles west of the intersection with State Road 54, and 2 miles east of the Hudspeth County line. In addition to the Manzanita Limestone Member, the road cut exposes the overlying uppermost sandstone of the Cherry Canyon Formation and the lowermost unit of the Bell Canyon Formation, the Hegler Limestone Member. We correlate this section with other outcrop and subsurface sections using distinctive Manzanita and Lower Seven Rivers-Hegler high frequency sequences (HFS) in the northern Delaware Basin (Fig. 2.27b; Tyrrell et al., 2004, 2006).

The type section of the Manzanita Limestone Member is at Nipple Hill about ten miles northeast of the Patterson Hills road cut (King, 1948). A reference section for the Manzanita Limestone Member is located on Rader Ridge at Ligon Ranch (Hampton, 1983) near the partial section of the Manzanita Limestone Member on Highway 62/180 near Nickel Creek station (Fig.2.27A). The sedimentology of the outcropping Manzanita Limestone Member in the Guadalupe and Delaware Mountains has been summarized by earlier workers (King, 1948; Newell, et al., 1953; Hampton, 1983; Harms and Pray, 1985). More regional subsurface relationships, sequence stratigraphy and regional recognition of five cycles (Mz-1 through Mz-5) in the Manzanita HFS have been presented elsewhere (Tyrrell et al., 2004, 2006).

MANZANITA LIMESTONE MEMBER IN THE PATTERSON HILLS ROAD CUT

This road cut did not exist when King (1948) mapped the Patterson Hills, nor was it studied by Hampton (1983, 1989). The four bentonite beds in the Patterson Hills section have been analyzed and compared to those in two other outcrops (Nicklen, 2003). This report builds on Nicklen’s measured section by adding more sedimentologic and stratigraphic detail. Although the base of the Manzanita Limestone Member is not exposed here, its top and the contact between the uppermost sandstone of the Cherry Canyon Formation and the base of the Hegler Limestone Member of the Bell Canyon Formation are present (Fig. 2.27). This road cut probably is bounded by normal faults, however only minor faulting is present in the exposed section. This permits us to interpret the sequence stratigraphy of the Manzanita Limestone Member in the Patterson Hills road cut and to correlate this section to Guadalupe and Delaware Mountains outcrops and the subsurface of the northern Delaware Basin (Fig. 2.27; Tyrrell et al., 2004, 2006).

Sedimentologic and stratigraphic features in the west-dipping beds of the Patterson Hills road cut are presented in Figure 2.27. The positions of the Mz-2, Mz-3, Mz-4 and Mz-5 cycles are indicated. Each cycle has a lower siliciclastic component grading up into an upper carbonate component. The Mz-1 cycle and the lower part of the Mz-2 cycle are not exposed in this road cut. Much of the Mz-3 cycle is dolomite whereas most of the Mz-4 cycle is limestone. Although most of the carbonate beds are mudstone to fine-grained packstone, there are distinctive conglomeratic beds within the Mz-3 cycle comprising intraclast-rich skeletal-pelloidal packstones (at 5.5 m and 9.2 m). No fusulinids have been found in our samples collected at one meter intervals throughout the Manzanita Limestone Member. Some fossils (e.g., the ammonoid *Paraceltites*) have been found in the skeletal packstones and wackestones of the Hegler Limestone Member at the western end of this road cut.

The outcrop photograph of Figure 2.27B portrays the lower part of the road cut section and it records the evenly bedded carbonates of the Manzanita Limestone Member interbedded with very fine grained sandstones, siltstones and four deeply weathered, greenish, shaley bentonite beds. Note that the thicker bentonite beds (bentonites 1 and 4) together with interbedded carbonate comprise the distinctive unit we call the “BCB” (bentonite-carbonate-bentonite) marker which generally can be recognized on wireline logs throughout the northern Delaware Basin and thus is an excellent regional marker. At the road cut, bentonite 1 contains pale green “chert” and bentonite 4 contains biotite.

The Manzanita Limestone Member has been interpreted as a low energy, basinal marine unit deposited in water depths of 200 to 400 meters (Hampton 1983, 1989). Sequence stratigraphic analysis suggests that sea level fluctuated during the time that the Manzanita Limestone Member accumulated. Taken together the Mz-1 through Mz-5 cycles comprise the Manzanita HFS (Tyrrell et al., 2004). Each cycle in the complete Manzanita HFS in the eastern Delaware Basin is interpreted as the product of lowstand (siliciclastic) deposition followed by highstand (carbonate) deposition accompanying sea level fluctuations (Tyrrell et al., 2004, 2006). The siliciclastic deposits are interpreted as the product of turbidity currents supplied by eolian sands and silts bypassing the shelf during lowstands. The carbonates formed during highstands and they are interpreted as the product of deposition from both suspension and turbidity currents. In particular, the intraclast pebble conglomerates in Mz-3 are interpreted as coarse grained turbidites derived from the platform margin or upper slope during

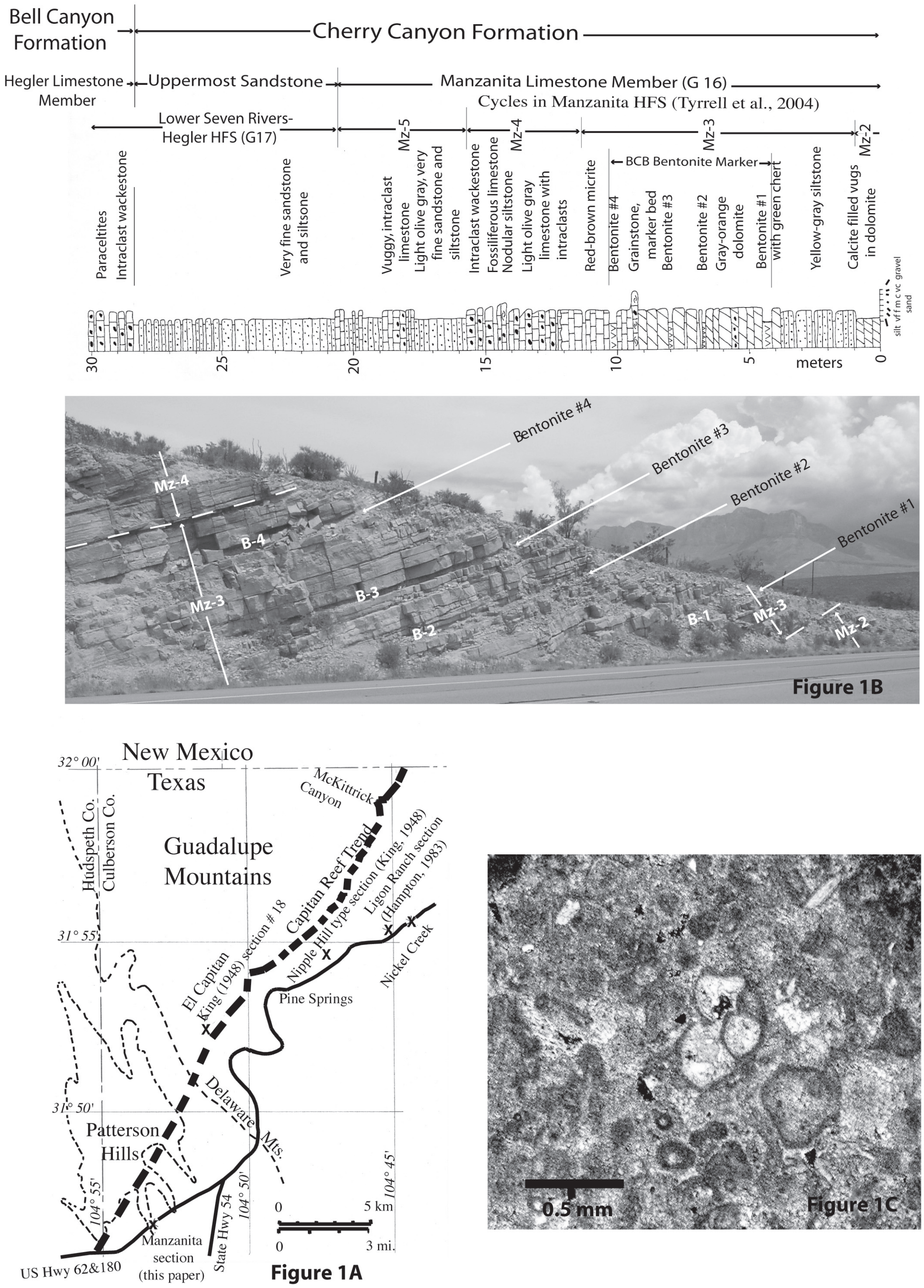


Figure 2.27. (A) Location map of the study site and other measured sections of the Manzanita Limestone Member; (B) Measured section of the entire west-dipping Patterson Hills road cut and an outcrop photograph of its eastern (lower) part. The interpreted sequence stratigraphy for the Manzanita HFS, including four of its five cycles, and the overlying Lower Seven Rivers-Hegler HFS is indicated; (C) Cross-polarized view of a peloidal-intraclast-skeletal packstone from 5.5 m above the base of the section. Though not shown here, this same unit also contains pebble-size intraclasts derived from an upper slope and platform environment.

a relative highstand. The bentonite beds are interpreted as altered volcanic ash derived from volcanic arcs in what is now northern Mexico (King, 1948; Nicklen, 2003). The distinctive "BCB" marker in the Mz-3 cycle is recognized in many wells penetrating the Manzanita Limestone Member in the Delaware Basin and has long been used as a basin-wide time marker by explorationists. Unlike most basin margin limestones in the Cherry Canyon and Bell Canyon Formations, the limestones in the Manzanita Limestone Member are fusulinid poor. Nonetheless, the Manzanita Limestone Member generally is interpreted to be in the zone of *Parafusulina* because of its stratigraphic position (G. Wilde, personal commun. 1999).

UPPERMOST CHERRY CANYON FORMATION AND LOWER HEGLER LIMESTONE MEMBER

At the western end of the road cut, the Manzanita Limestone Member is overlain by eight meters (24 feet) of yellowish, very fine-grained sandstone and siltstone of the uppermost Cherry Canyon Formation interpreted to be the lowstand sandstone at the base of the Lower Seven Rivers-Hegler high frequency sequence (HFS G-17 of Kerans and Tinker, 1999). This sandstone unit locally contains oil reservoirs in the Delaware Basin such as the channel sandstone reservoir in the Indian Draw Field east of Carlsbad (Cromwell, 1969). Also exposed in this outcrop are the lower two meters (6 feet) of the Hegler Limestone Member of the Bell Canyon Formation. The Hegler Limestone Member is interpreted as the highstand carbonate portion of the Lower Seven Rivers-Hegler high frequency sequence. It is the lowest basinal limestone tongue equivalent to the lower Capitan Reef. We did not find fusulinids in the Patterson Hills road cut section, but *Polydiexodina* has been found in other outcrops of the Hegler Limestone Member near its basin margin in the Guadalupe Mountains area (Newell et al., 1953).

REFERENCES

Cromwell, D.W., 1979, Indian Draw Delaware Field: a model for deeper Delaware sand exploration, *in* Sullivan, N.M, ed., Guadalupian Delaware Mountain Group of West Texas and S

Hampton, B.D., 1983, Carbonate sedimentology of the Manzanita Member of the Cherry Canyon Formation (Middle Guadalupian, Permian), Guadalupe Mountains [M.S. thesis]: Madison, University of Wisconsin, 178 p.

Hampton, B.D., 1989, Carbonate sedimentology of the Manzanita Member of the Cherry Canyon Formation, *in* Harris, P.M. and Grover, G.S., eds., Subsurface and Outcrop Examination of the Capitan Shelf Margin, Northern Delaware Basin: Society of Economic Paleontologists and Mineralogists, Core Workshop 13, p. 431-439.

Harms, J.C., and Pray, L.C., 1985, Day 3 road log and locality guide, *in* Cunningham, B.K., and Hedrick, C.L., eds., Permian Carbonate/Clastic Sedimentology, Guadalupe Mountains: Permian Basin Section, Society of Paleontologists and Mineralogists, Guidebook, Publication 85-24, p. 56-57.

Kerans, C., and Tinker, S.W., 1999, Extrinsic stratigraphic controls on development of the Capitan Reef Complex, *in* Saller, A.H., Harris, P.M., Kirkland, B.L., and Mazzulo, S.J., eds., Geologi

King, P.B., 1948, Geology of the Southern Guadalupe Mountains, Texas: US Geological Survey Professional Paper 215, 183 p.

Newell, N.D., Rigby, J.K., Fischer, A.G., Whiteman, A.J., Hickox, J.E., and Bradley, J.S., 1953, The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico: San Francisco, W.F. Freeman, 236 p.

Nicklen, B.L., 2003, Middle Guadalupian (Permian) Bentonite Beds, Manzanita Member, Cherry Canyon Formation, West Texas: Stratigraphic and Tectonomagmatic Applications [M.S. thesis]: Cincinnati, University of Cincinnati, 66 p.

Tyrrell, W.W., Diemer, J.A., Bell, G.L, and Griffing, D.H., 2004, Subsurface sequence stratigraphy of the Manzanita Limestone Member, Cherry Canyon Formation, northern Delaware Basin, New Mexico and West Texas, *in* Trentham, R.C., ed., Banking on the Permian Basin: West Texas Geological Society Publication 04-112, p. 125-155.

Tyrrell, W.W., Diemer, J.A., Bell, G.L, and Griffing, D.H., 2006, The Manzanita Limestone Member, Cherry Canyon Formation, northern Delaware Basin, New Mexico and West Texas, *in* Entzminger, D.J., ed., Basinal Facies of the Uppermost Guadalupian: 2006 Permian Basin Section Fieldtrip Guidebook, Society of Economic Paleontologists and Mineralogists, *in press*.

HALITE DEPOSITIONAL CYCLES IN THE UPPER PERMIAN SALADO FORMATION

[1]Robert M. Holt, [2]Dennis W. Powers, and [3]Tim K. Lowenstein
[1]University of Mississippi, Department of Geology and Geological Engineering, 118 Carrier Hall, University, MS, 38677, rmholt@olemiss.edu
[2]Consulting Geologist, 140 Hemley Road, Anthony, TX, 79821, dwpowers@evaporites.com
[3]Binghamton University, Department of Geological Sciences and Environmental Studies, P.O. Box 6000, Binghamton, NY, 13902, lowenst@binghamton.edu

The Upper Permian Salado Formation of southeastern New Mexico and west Texas is the middle of three Ochoan evaporite-bearing formations in the Delaware Basin and is underlain by the Castile Formation and overlain by the Rustler Formation. About 85 to 90 percent of the Salado is halite with the remainder consisting of anhydrite, polyhalite and minor amounts of other potassium-bearing minerals (Jones et al., 1973). Beds of anhydrite and polyhalite alternate with thicker beds of halite throughout the Salado section. The Salado is nearly 600 meters thick in the subsurface of the eastern part of the Delaware Basin and is only a few tens of meters of brecciated insoluble material at outcrops in the western part of the basin. The Salado is subdivided into three informal members: an unnamed upper member, a middle member locally designated the McNutt potash zone, and an unnamed lower member. Individual beds within the Salado are often traceable for large distances. These areally persistent beds allow the middle and upper Salado to be subdivided on a much finer scale. A system of numbering areally extensive beds of anhydrite and polyhalite as markerbeds was introduced by geologists of the U. S. Geological Survey (Jones et al., 1960). This markerbed system is used extensively by mining companies in the Carlsbad potash district and by researchers at the Waste Isolation Pilot Plant (WIPP) for smaller scale stratigraphic control.

Early descriptions of the Salado Formation showed repetitive vertical variations in mineralogy. Schaller and Henderson (1932) described sequences of clay–anhydrite–polyhalite–halite and minor amounts of polyhalitic halite, and Jones (1954, 1972) reported cyclical units consisting of clay–magnesite–anhydrite, polyhalite or glauberite–halite–argillaceous halite capped with mudstone. These authors interpreted these mineralogical changes to reflect simple evaporation of marine and terrestrial waters.

Using a sedimentological approach, Lowenstein (1982, 1988) began to reveal the depositional complexity of the Salado. He identified two types of Salado depositional cycles: a Type I cycle consisting of a basal mixed siliciclastic and carbonate (magnesite) mudstone – laminated to massive anhydrite or polyhalite – halite – halite with mud and a Type II cycle consisting of halite grading to muddy halite. The Type I cycle was interpreted to record upward desiccation beginning with carbonate and sulfate deposition in a perennial lake of marine origin and ending with halite deposition in a saline lagoon/salt pan environment. Type II cycles were differentiated from Type I cycles because they are not underlain by sulfate units and display no features indicative of prolonged subaqueous deposition from perennial lagoons. Type II cycles were interpreted to be deposited in shallow lake or lagoon and salt pan environments following terrestrially-derived flooding.

Based on detailed mapping of underground exposures and shafts at the WIPP site, Holt and Powers (1990 a, b) and Holt (1993) extended the sedimentological work of Lowenstein and

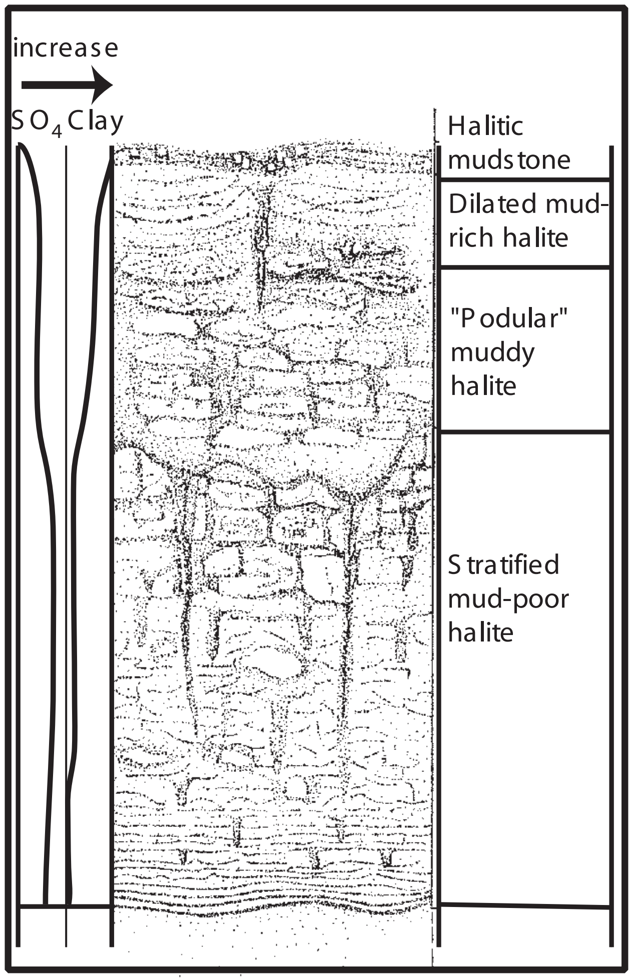


Figure 2.28. The idealized Salado halite cycle consists of four lithofacies. The stratified mud-poor-halite lithofacies is characterized by a well-developed to crude sense of horizontal stratification and halite with chevron and cornet fluid inclusion banding and cumulate fabrics. "Podular" muddy halite consists of irregular pods and lenses of finely- to medium-crystalline halite intercalated with irregular stringers of sulfate or clay bounded by argillaceous halite containing coarsely-crystalline halite cements. The dilated mud-rich halite lithofacies consists of argillaceous halite exhibiting abundant displacive halite cement fabrics. Halitic mudstone contains isolated crystals and aggregates of fine to coarse displacive halite crystals and poikilotopic halite cements. Dissolution pipes, macropores, and solution lags become more abundant upward through the cycle.

recognized additional complexities in Salado depositional cycles. They observed little textural difference between halite sequences with or without underlying sulfate beds, and they documented subaerial exposure on the tops of some sulfate interbeds, indicating complete desiccation prior to halite accumulation. On this basis, they separated halite sequences from underlying units and recognized variations on only one cyclical pattern in Salado halite.

Holt and Powers (1990 a, b) and Holt (1993) identify four distinct halite lithofacies within a complete Salado halite cycle (Fig. 2.28), from the base upward: 1) stratified mud-poor halite, 2) podular muddy halite, 3) dilated mud-rich halite, and 4) halitic mudstone. The textures and features found within each lithofacies can be attributed to subaqueous (e.g., chevron and cornet halite or planar dissolution) conditions, subaerial exposure and vadose-zone alteration (e.g., hygroscopic alteration, efflorescent crusts, desiccation cracks, solution pipes and macropores), and phreatic-zone cements (e.g., overgrowths and displacive cements). Sedimentary structures at the base of the cycles mainly reflect subaqueous halite accumulation and sedimentary structures higher in the cycle indicate increasingly longer periods of subaerial exposure and more complete vadose-zone dissolution. Mud contents and solution pipe depths generally increase upward suggesting longer periods of exposure and deeper water table positions. Displacive phreatic-zone cements also become more abundant upward through the cycles.

Lithofacies in “complete” Salado halite cycles record the vertical progression of the following depositional environments: 1) mud-poor saltpan, 2) “hummocky” saltpan (e.g., Devil’s Golf Course, Death Valley, California), 3) mud-rich saltpan, and 4) saline mudflat. Mud-poor salt pan halite was deposited following a flooding event. Evaporation and reworking of existing halite increased the salinity to halite saturation. Halite accumulated subaqueously until the basin was completely desiccated. When the water table dropped below the sediment surface, vadose-zone dissolution produced solution lags, dissolution pipes and macropores. Small-scale flooding events occurred episodically, creating subaqueous conditions that allowed additional halite to accumulate. Vadose zone alteration increased with time as flooding events became less frequent. Long periods of deep water-table conditions cycled with shallow water table and episodic saline lagoon conditions producing and preserving podular muddy halite in a “hummocky” salt pan environment. Vadose zone alteration intensified with time, reducing the overall relief across the basin. This allowed more efficient transport of clastic materials into the basin, and a mud-rich salt pan developed. Saline mudflat environments moved laterally toward the Salado depocenter, as desiccation continued.

Understanding the depositional, hydrologic, and climatic conditions that lead to the deposition of the Salado Formation have become increasingly important. Recent geobiological studies of the Salado have reported the oldest known life on Earth, viable halotolerant bacteria, trapped in halite fluid inclusions (Vreeland et al., 2000; Powers et al., 2001). Geochemical studies of fluid inclusion chemistries from the bacteria sampling site show that fluid inclusions are consistent with Permian seawater and support a Permian age for the bacteria (Satterfield et al., 2005). These studies and earlier studies of Salado fluids (e.g., Stein and Krumhansl, 1988) suggest a complicated origin for Salado halite.

In addition, recent age dates from the Salado (251 ± 0.2 Ma) and stratigraphic equivalents to overlying rocks (249-250 Ma) suggest that the Permian-Triassic boundary may lie within or just above the Salado (Renne et al., 1998, 2001; Sharp et al., 1997). The record of Salado rocks deposited under repeated variations in depositional environments and the fluid inclusions trapped within the salt may contain an unprecedented record of the climatic, environmental, and geobiological conditions surrounding or leading up to the end-Permian extinction. Unlike most sediments, evaporites are readily dissolved and altered within their depositional environment. Dissolution and alteration within the vadose zone and later cementation within the phreatic zone can juxtapose evaporite minerals derived from very different fluids. Future geobiological and paleoclimatic studies of the Salado require a detailed understanding of Salado depositional cycles and environments.

REFERENCES

Holt, R.M., 1993, Sedimentary textures, structures, and lithofacies in the Salado Formation: A guide for recognition, classification, and interpretation: U.S. Department of Energy, DOE/WIPP93-056, 129 p.

Holt, R.M., and Powers, D.W., 1990a, Geotechnical activities in the air intake shaft (AIS): DOE/WIPP 90-051, U.S. Department of Energy, Carlsbad, NM.

Holt, R.M., and Powers, D.W., 1990b, Halite sequences within the Late Permian Salado Formation in the vicinity of the Waste Isolation Pilot Plant, *in* Powers, D.W., Holt, R.M., Beauheim, R.L., and Rempe, N., eds., 1990, Geological and Hydrological Studies of Evaporites in the Northern Delaware Basin for the Waste Isolation Pilot Plant (WIPP): Guidebook 14, Geological Society of America (Dallas Geological Society), P. 45-78.

Jones, C.L., 1954, The occurrence and distribution of potassium minerals in southeastern New Mexico: New Mexico Geological Society, 5th Field Conference, Guidebook, p. 107-112.

Jones, C.L., 1972, Permian basin potash deposits, southwestern United States, *in* Geology of Saline Deposits, UNESCO, Earth Science Series, No. 7, p. 191-201.

Jones, C.L., Bowles, C.G., and Bell, K.G., 1960, Experimental drill hole logging in potash deposits of the Carlsbad District, New Mexico: U.S. Geological Survery, Open-file Report 60-84, 22 p.

Jones, C.L.,, Cooley, M.E., and Bachman, G.O., 1973, Salt deposits of Los Medaños area, Eddy and Lea Counties, New Mexico: U.S. Geological Survey, Open-file Report 4339-7, 67 p.

Lowenstein, T.K., 1982, Primary features in a potash evaporite deposit, the Permian Salado formation of west Texas and New Mexico, *in* Handford, C.R., Loucks, R.G., and Davies, G.R., eds., Depositional and Diagenetic Spectra of Evaporites – A Core Workshop: Society of Economic Paleontologists and Mineralogists, Core Workshop 3, p. 276-304.

Lowenstein, T.K., 1988, Origin of depositional cycles in a Permian “saline giant”: The Salado (McNutt zone) evaporites of New Mexico and Texas: Geological Society of America, Bulletin, v. 100, p. 592-608.

Powers, D.W., Vreeland, R.H., Rosenzweig, W.D., 2001, How old are bacteria from the Permian age?–Reply: *Nature*, v. 411, p. 155-156.

Renne, P.R., Sharp, W.D. & Becker, T.A., 1998, $^{40}Ar/^{39}Ar$ dating of langbeinite $[K_2Mg_2(SO_4)_3]$ in late Permian evaporites of the Salado Formation, Southeastern New Mexico, USA: Mineralogy Magazine, v. 62A, p. 1253-1254.

Renne, P.R., Sharp, W.D., Montañez, I.P., Becker, T.A., and Zierenberg, R.A., 2001, $^{40}Ar/^{39}Ar$ dating of Late Permian evaporites, southeastern New Mexico, USA: Earth and Planetary Letters, v. 193, p. 539-547.
Satterfield, C.L., Lowenstein, T.K., Vreeland, R.H., Rosenzweig, W.D., and Powers, D.W., 2005, New evidence for 250 Ma age of halotolerant bacterium from a Permian salt crystal: Geology, v. 33, p. 265-268.
Schaller, W.T., and Henderson, E.P., 1932, Mineralogy of drill cores from the potash field of New Mexico and Texas: U.S. Geological Survey, Bulletin 833, 124 p.
Sharp, W.D., Ludwig, K.R., Renne, P.R., Steiner, M.B., and Fanning, C.M., 1997, $^{40}Ar/^{39}Ar$ and U-Pb SHRIMP dating of latest Permian tephras in the Midland Basin, Texas: Proceedings, South American Symposium on Isotope Geology, v. 1, p. 293-294.
Stein, C.L., and Krumhansl, J.L., 1988, A model for the evolution of brines in salt from the lower Salado Formation, southeastern New Mexico: Geochimica et Cosmochimica Acta, v. 52, p. 1037-1046.
Vreeland, R.H., Rosenzweig, W.D., and Powers, D.W., 2000, Isolation of a 250-million-year-old halotolerant bacterium from a primary salt crystal: *Nature*, v. 407, p. 897-900.

PANGEAN MONSOON AND CLIMATIC CYCLES IN NM-TEXAS STATE-LINE OUTCROP

Roger Y. Anderson
Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM 87131, ryand@unbm.edu

Thin, dark layers of calcium carbonate, alternating with thicker, light-colored layers of calcium sulfate in the Upper Permian Castile evaporite, as exposed in the highway cut near the NM-Texas state line, are an important source of information about monsoonal climate. The coastal basin, which now contains the banded Castile evaporite, was almost on the equator, but the west coast of the giant landmass was desert, with little if any moisture reaching the area from the Tethyian sea (Fig. 2.29). During dry seasons (fall-winter-spring), layers of pure calcium sulfate were precipitated from a brine pool within the newly isolated Delaware (evaporite) Basin (technically a lake, Kirkland, 2003; Anderson and Dean, 1995). In contrast, a strong monsoon marked the onset of the “wet” season as warm air rising above the huge Pangean landmass drew marine air in over the coastal basin, and possibly a cloud layer, cutting off or greatly reducing sulfate precipitation and allowing calcium carbonate and organic material to accumulate in thin, discrete layers (Kirkland, 2003). Carbonate layers generally have a sharp lower contact that marks a sudden onset of the annual monsoon. Thin-sections show that a rain of calcite crystals, settling on the sediment interface, was overtaken and diluted by sulfate particles as evaporative concentration and sulfate precipitation increased toward the end of the monsoon season.

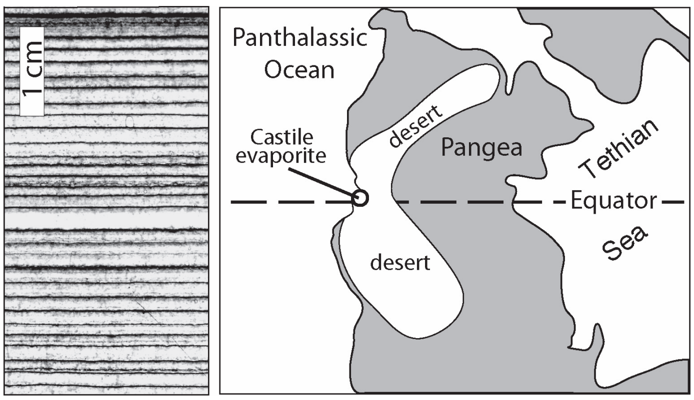


Fig. 2.29. Approximate location of Castile evaporite in relation to Pangean equator. Dark- and light-colored Castile laminations (right) formed annually as insolation, probably over the larger southern hemisphere of Pangea, heated the continental interior, drawing clouds and monsoonal moisture inland over the coastal basin, thereby reducing evaporation and sulfate precipitation and allowing dark layers of calcium carbonate to accumulate. Sulfate layers accumulated during drier seasons.

In the 1960’s and 70’s, several cores were collected of the banded Castile Formation and a series of 209,000 pairs (couplets) of dark-light layers was measured and assembled into a continuous time series (Anderson et al, 1972; Anderson, 1982). Some features of the series, as well as microfolding of laminations, were considered in a previous guidebook (Anderson and Kirkland, 1987). Since then, improved analytical methods have permitted re-examination of climatic features represented by Castile laminations, as illustrated in Fig. 2.30, which is a raw-data plot, year by year, of couplet thickness, trimmed to ~10 mm. Average thickness of couplets is ~1.8 mm in core samples of anhydrite, but layers are thicker in outcrop after expansion as gypsum.

The thinly laminated limestone near the midpoint of the series in Fig. 2.30 (~T_o +106-108 ky) corresponds to a ~0.5 m limestone unit that forms a low topographic ridge and caps the exposed Castile section in the previous State-line road cut. This unit contains almost no calcium sulfate and represents an extreme freshening event near the base of the Halite 3 Member of the Castile Fm. that occurred during a precession cycle (Fig. 2.30). About 7.5 such orbital cycles are resolved in a 162 kyr segment of the series, for an average cycle length of ~22 kyrs, or almost the same period found in the late Pleistocene.

Most of the exposed Castile corresponds to the Anhydrite 3 Member (Fig. 2.30), and changes in couplet thickness on the scale of several meters, as seen along the face of the road cut, correspond to strong, millennial climate cycles in the range of ~1500 to ~2300 years (Fig. 2.30). Such climatically driven salinity cycles usually commence with a laminated limestone and progress through an increase in sulfate thickness until precipitation of sulfate is greatly reduced during the next freshening event. In many cases the salinity cycles contain anomalously thick layers of

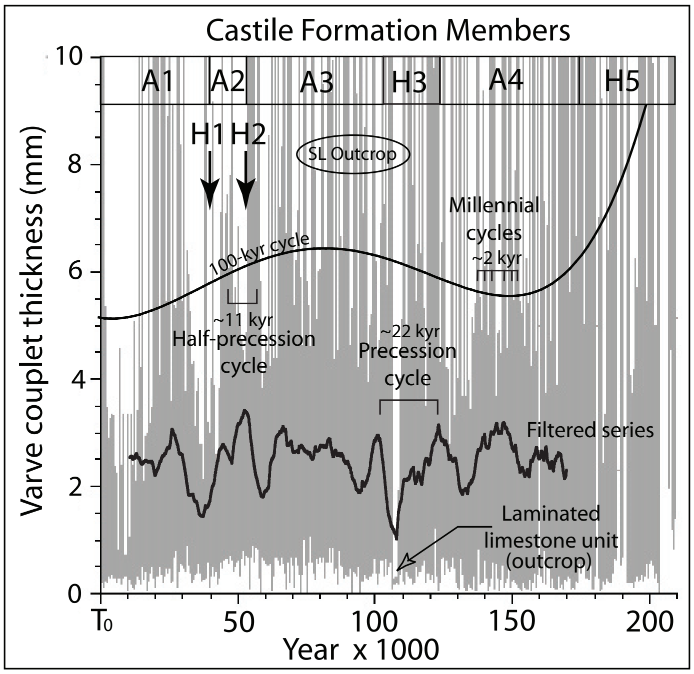


Fig. 2.30. Plot of raw thickness for pairs of light (sulfate) and dark (carbonate) layers in Castile time series, trimmed to 10 mm. Series has a length and resolution comparable to high-latitude ice cores, and provides a detailed record of changes in tropical climate and monsoonal variability. Solid shade of gray across mid-range of thickness values is a result of closely spaced data. Exceptionally thick layers of calcium sulfate in outcrop sometimes develop a nodular structure, and reflect high rates of evaporative concentration of brine during millennial-scale climate (monsoonal) oscillations. Widest trace on graph (black) is couplet thickness after detrending and applying a 590-year weighted filter, which shows roughly 7.5 cycles in 162 kyrs (~22 ky per cycle), attributed to a monsoonal response to changes in insolation during precession of the equinox. The exposure, however, uncovers only parts of Anhydrite III and Halite III members of Castile Fm. (oval) and precession cycles are not easily recognized in outcrop, although a 0.5 m unit of laminated limestone near the top of the previous road cut represents a profound, ~2000-year freshening event during a precession cycle.

nearly pure calcium sulfate (visible in outcrop) that were deposited after a high rate of evaporative concentration increased the density of the surface brine layer in the basin to the point of overturn, resulting in the mixing of brine and a greatly increased rate of sulfate precipitation (Kirkland, 2003). In many cases, thick layers of calcium sulfate developed a nodular structure, probably related to increased salinity, but rarely did concentration increase to the point of halite precipitation. All halite has been dissolved in the area of the outcrop, leaving residues of dissolution breccia. The position of Halite I and Halite II members in the Castile time series are noted in Fig. 2.30. Both Halite members accumulated during single millennial-scale salinity cycles, with the 140-foot-thick Halite I member having only 1063 annual sulfate-halite couplets (plotted only as sulfate thickness in Fig. 2.30). Strong climatic oscillations on the millennial time scale include glacial stadial-interstadial cycles of the late Pleistocene and powerful Dansgaard-Oeschger (DO) events, and the occurrence of strong and similar-scale cycles in the Castile suggests a persistent climatic feature with tropical and monsoonal associations.

Other climate oscillations found in the Castile include the enigmatic 100-kyr cycle (Fig. 2.30), which is the dominant glacial-interglacial cycle of the late Pleistocene, and cycles of half-precession (~11 kyr). The presence of half-precession climate cycles in the Castile confirms that the Delaware Basin was close to the paleoequator because land on both sides of the equator must be alternately heated during the precession cycle in order to develop cycles of half-precession. Actually, the Delaware Basin may have been closer to the equator than depicted in some reconstructions of the Pangean continent, and a location on, or just south of the equator cannot be ruled out (Fig. 2.29). Still being investigated is an unexpected finding that the period of precession recorded in the Castile is nearly the same as modern precession (Anderson, in preparation). In summary, the remarkable laminations in the Castile Fm. appear to document an exceptionally stable planetary system and also confirm the orbital-monsoon theory of climatic change (Kutzbach and Gallimore, 1989).

REFERENCES

Anderson, R. Y. , 1982, A long geoclimatic record from the Permian. Journal of Geophysical Research 87, C9, 7285-7294.

Anderson, R. Y., in preparation, Pangean Monsoon as Timepiece for Orbital Stability.

Anderson, R. Y., Dean, W. E., Jr., Snider, H. I., and Kirkland, D. W., 1972, Permian Castile varved evaporite sequence, West Texas and New Mexico: Geol. Soc. Amer. Bull., v. 83, pp. 59-86.

Anderson, R. Y. & Dean, W. E., 1995, Filling of the Delaware Basin: Hydrologic and climatic controls on the Upper Permian Castile Formation varved evaporite *in* Scholle, P. A., Peryt, T. M., and Ulmer-Scholle, D. S., (eds), *The Permian of Northern Pangea, Volume 2: Sedimentary Basins and Economic Resources*, Berlin, Springer-Verlag, 61-78.

Anderson, R. Y., Kietzke, K. K., and Rhodes, D., (1978), Development of dissolution breccias, northern Delaware basin, New Mexico and Texas: New Mexico Bur. Mines and Mineral Resources Circular 159, p. 47-52.

Anderson, R. Y., and Kirkland, D. W., 1987, Banded Castile evaporites, Delaware Basin, New Mexico, Geol. Society America Centennial Field Guide, v. 2, 455-458.

Kirkland, D. W., 2003 An explanation of the varves of the Castile evaporates (Upper Permian), Texas and New Mexico, USA. Sedimentology, 50, 899-920.

Kutzbach, J. E. & Gallimore, R. G.. 1989, Pangean climates: megamonsoons of the megacontinent. Jour. of Geophysical Research, 94, No. D3, 3341-3357.

GYPSUM KARST OF THE CHOSA DRAW AREA

Kevin W. Stafford
Cave and Karst Studies, Dept. of Earth and Environmental Science, New Mexico Institute of Mining and Technology, Socorro, NM 87801
kwstafford@juno.com

The Chosa Draw area contains the best documented region of epigenetic gypsum karst development of the Gypsum Plain, including epikarst, sinkholes and cavernous porosity. The Gypsum Plain covers an area of ~2600 km^2 (Hill, 1996) and is comprised of Ochoan rocks that are dominated by outcrops of the Castile Formation (~1600 km^2) (Figure 1) (Kelley, 1971; Dietrich et al., 1995). The Chosa Draw Area of Critical Environmental Concern includes an area of approximately 10 km^2 that was delineated by the New Mexico Bureau of Land Management in order to protect the extremely well developed gypsum karst within the Chosa Draw drainage area (BLM, 1997). The area is characterized by intense karst development with up to fifty sinkholes per square kilometer and contains the second longest documented gypsum cave in North America (Parks Ranch Cave) (Klimchouk, 1996a).

Epigenetic karst forms when rocks are exposed to the direct effects of meteoric weathering, as opposed to hypogenetic karst which forms in confined settings (*e.g.* Carlsbad Caverns) (White, 1988). When gypsum is exposed at the surface it can produce dramatic karst landscapes as a result of the high dissolution kinetics of gypsum. In pure water, gypsum's solubility is approximately four orders of magnitude greater than calcium carbonate; however, in the presence of carbon dioxide, the solubility of calcium carbonate is substantially increased, while gypsum remains effectively unchanged (Klimchouk, 1996b). Therefore, gypsum solubility is generally 10 to 30 times that of calcium carbonate when exposed to meteoric waters. Because of its high solubility, dissolution in exposed gypsum rocks often produces well-developed interface karst features (*i.e.* sinkholes, dolines) that focus water into small cave networks (Klimchouck, 1996c).

The Chosa Draw area is dominated by gypsum outcrops that are sparsely mantled with gypsite and Cenozoic gravel deposits (Sares, 1984). Although the Castile Formation is generally described as a calcite / anhydrite (gypsum) laminated unit (*e.g.* Anderson and Kirkland, 1966; Anderson et al., 1972), other fabrics are common (*e.g.* massive, tabular and nodular) (Dean et al., 1975). Exposures at the surface and in caves within the Chosa Draw area are primarily massive, sucrosic gypsum, which is likely a significant factor in the intensekarst development within the area. Standard tablet studies within the Chosa Draw area have found that dissolution rates are not uniform (Nance, 1993). Instead, studies of relative rates of dissolution imply that surface denudation and karst formation is most rapid in caves and arroyos, while areas mantled with gravel and gypsite exhibit minimal dissolution due to surface armoring (Nance, 1993).

Throughout the Chosa Draw area, well-developed karren can be observed where gypsum rock is not mantled by soils or gravel deposits. On vertical rock faces within arroyos, rillenkarren development forms linear features centimeters to tens of centimeters in length that can be laterally extensive. On flat surfaces, rain pits and small kaminitza-like features with limited lateral extent occasionally develop. In areas that exhibit jointing, cleft karren may form, but it is often associated with sinkhole development. Although common in some regions of the Gypsum Plain, no blister caves (*i.e.* tumuli) have been observed within the local area.

Two distinct morphologies of sinkholes have been identified in the region based on width to depth ratios (Sares, 1984). Wide, shallow sinkholes generally form elliptical features that are frequently filled with soil and dense vegetation. Deep, narrow sinkholes are generally associated with incised arroyos and contain swallets that focus water into the subsurface. Most deep sinks are associated with cave development, however the majority of the caves are too small for human entry. The Chosa Draw area is unique since many karst features form integrated cave systems that are large enough to enter. The origin of these caves has been attributed to a regional base level drop of the Black River during the late Pleistocene, which resulted in deep incision of many arroyos and the formation of caves that function as hydrologic bypass features (Sares, 1984).

Numerous caves have developed within the Chosa Draw area, but the dominant feature is Parks Ranch Cave. The Parks Ranch Cave system is the longest gypsum cave in New Mexico and the Delaware Basin, with a surveyed length of 6269 meters (Figure 1) (Knapp, 1992; Klimchouk, 1996a). The cave was originally explored in the 1950's, while surveying continued intermittently through the 1990's. During the regional drought of the 1980's, significant portions of the cave were discovered and mapped as local lowering of base level opened previously flooded portions to exploration. Numerous passages that have not been fully explored remain to be surveyed , suggesting this cave system is even more complex than currently realized (Knapp, 1992).

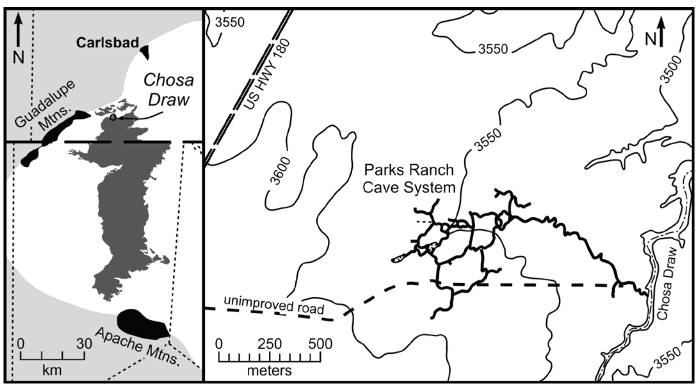


Figure 1. Location of the Chosa Draw area in relation to outcrops of the Castile Formation (dark grey) of the Gypsum Plain and the western edge of the Delaware Basin (white), with enlarged view of the Chosa Draw area showing relationship between topography (elevation in feet), Chosa Draw and Parks Ranch Cave (line plot of cave) (adapted from Nance, 1993).

Passage morphology ranges from elliptical phreatic tubes with diameters up to three meters, to incised, meandering canyons of variable width and height. In general, the cave system forms an anastomosing fluivial-like pattern. At least two primary horizons of cave passage development are formed within the Parks Ranch Cave-Entrance #1 sinkhole (the northern fenced sinkhole within the Chosa Draw Area of Critical Environmental Concern). Many sections of the cave system exhibit structural controls, with passages distinctly aligned along fractures associated with regional jointing oriented at ~N35°E and ~N15°W (Nance, 1993). However, numerous fractures can be observed intersecting passages at acute angles with minimal dissolutional widening and passages frequently change direction abruptly with no associated fractures to provide pre-existing controls for passage development. These abrupt changes in passage orientation suggest additional factors beyond structural deformation are significantly effecting development of gypsum karst; although the exact mechanism controlling cave development not associated with fractures is not clear at this time.

Regional hydrology shows that normal flow through Parks Ranch Cave supplies local groundwater recharge that discharges along the Black River as a series of small seeps and springs (Sares, 1984). However, during intense flood events, groundwater recharge rates are exceeded and Parks Ranch Cave becomes a bypass feature for flood waters that discharge into Chosa Draw at Resurgence Cave (Sares, 1984). It is likely that many other caves within the region exhibit similar hydrologic patterns.

The Chosa Draw area, and Park's Ranch Cave in particular, provide an exceptional example of epigenetic gypsum speleogenesis. Although the Chosa Draw area is the best documented region of gypsum karst within the Gypsum Plain, it is not unique. Gypsum cave and karst development is common throughout the entire region with several hundred individual features documented elsewhere in Eddy County, New Mexico and Culberson County, Texas.

REFERENCES

Anderson, R.Y., Dean, W.E., Kirkland, D.W. and Snider, H.I., Permian Castile varved evaporite sequence, west Texas and New Mexico: Geologic Society of America Bulletin, v. 83, p. 59-86.

Anderson, R.Y. and Kirkland, D.W., 1966, Intrabasin varve correlation: Geological Society of America Bulletin, v. 77, p. 241-256.

Bureau of Land Management, 1997, Carlsbad, New Mexico: United State Department of the Interior, Bureau of Land Management, 1 sheet.

Dean, W.E., Davies, G.R. and Anderson, R.Y., 1975, Sedimentological significance of nodular and laminated anhydrite: Geology, v. 3, p. 367-372.

Dietrich, J.W., Owen, D.E., Shelby, C.A. and Barnes, V.E., Geologic atlas of Texas: Van Horn-El Paso sheet: University of Texas Bureau of Economic Geology Publication, Austin, TX, 1 sheet.

Hill, C.A., 1996, Geology of the Delaware Basin, Guadalupe, Apache and Glass Mountains: New Mexico and West Texas: Permian Basin Section/Society of Economic Paleontologists and Mineralogists Publication 96-39, 480 p.

Kelley, V.C., 1971, Geology of the Pecos Country, Southeastern New Mexico: New Mexico Bureau of Mines and Mineral Resoureces, Memoir 24, 78 p.

Klimchouk, A., 1996a, Speleogenesis in gypsum: International Journal of Speleology, v. 25, no. 3-4, p. 61-82.

Klimchouk, A., 1996b, Dissolution and conversion of gypsum and anhydrite: International Journal of Speleology, v. 25, no. 3-4, p. 21-36.

Klimchouk, A., 1996c, Dissolution of gypsum from field observations: International Journal of Speleology, v. 25, no. 3-4, p. 37-48.

Knapp, R., 1992, Park's Ranch Cave System, in Belski, D. ed., GYPKAP Report #2, Southwestern Region of the National Speleological Society: Adobe Press, Albuquerque, NM, p. 34.

Nance, R.G., 1993, Application of the Standard Tablet Method to a Study of Denudation in Gypsum Karst, Chosa Draw, Southeastern New Mexico: MS Thesis, University of Northern Colorado, Greeley, CO, p. 82.

Sares, S.W., 1984, Hydrologic and Geomorphic Development of a Low Relief Evaporite Karst Drainage Basin, Southeastern New Mexico: MS Thesis, University of New Mexico, Albuquerque, NM, p. 123.

White, W.B., 1988, Geomorphology and Hydrology of Karst Terrains: Oxford University Press, New York, NY, 464 p.

Science team exiting Lower Cave, Carlsbad Cavern, New Mexico. Paticipants ((left to right); Leslie Melim (Western Illinois University), Mike Spilde (University of New Mexico), Penny Boston (New Mexico Tech), and Michael Queen (Pomona College). Photo by Kenneth Ingram.