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1992 SASGS FIELD TRIP
Geology of the Southern High Plains

CAPROCK CANYONS STATE PARK
GUIDEBOOK FOR THE
GEOLOGY OF THE SOUTHERN HIGH PLAINS
AT
CAPROCK CANYONS STATE PARK
TEXAS

prepared for
FALL 1992 FIELDTRIP
OF THE SOUTHWESTERN ASSOCIATION OF
STUDENT GEOLOGICAL SOCIETIES
(S.A.S.G.S.)
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hosted by
Department of Geosciences, Texas Tech University
Lubbock, Texas 79409-1053

Tom Lehman and John Schnable, editors
James Browning, fieldtrip coordinator

Contributing Authors:
Craig Bunting
Terri Husband
Tom Lehman
John Schnable
Susan Tomlinson
Mike Williams

Cover Art by Susan Tomlinson
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INTRODUCTION

Welcome to the 1992 SASGS fieldtrip hosted by the Department of Geosciences at Texas Tech University. This year we will examine the geology of the Southern High Plains as exposed at Caprock Canyons State Park in Briscoe County, Texas, and in the surrounding vicinity.

REMEMBER, WHILE YOU ARE IN CAPROCK CANYONS STATE PARK, DO NOT COLLECT ROCK, MINERAL, OR FOSSIL SPECIMENS, AND STAY ON THE MARKED HIKING TRAILS.

Caprock Canyons State Park includes several canyons, cut by tributaries of the Little Red River, into the Caprock Escarpment. This escarpment is the boundary between the Rolling Plains of north central Texas, underlain mostly by Permian "red beds", and the Southern High Plains of west Texas, underlain by Triassic, Tertiary, and Quaternary strata. On this fieldtrip, we will follow a hiking trail up the Caprock Escarpment where we can examine this stratigraphy.

This guidebook is divided into two parts. The first is a trail log, with brief descriptions of points of interest along the hiking trail. The second part includes more detailed discussions of each of the stratigraphic units, along with references for further information. A great deal of information on this region is found in three recent collections of papers published by the Bureau of Economic Geology and edited by Thomas Gustavson (1986, 1990, 1990).
EXPLANATION

- Channel and terrace deposits
- Lingos Formation
- Blackwater Draw Formation
- Tertiary: Ogallala Formation
- Triassic: Dockum Group
- Permian: Quartermaster Formation and Whitehorse Sandstone

GENERALIZED GEOLOGIC MAP OF THE FIELD TRIP AREA (from Caran & Baumgardner,1990)
Protoavis texensis

Texas Tech Was The First To Give You The BIRD
On this fieldtrip we will follow the Upper Canyon Trail (5.5 mi.) and the Canyon Loop Trail (1.5 mi.) in Caprock Canyons State Park. This trail will take you upstream along the valley of the South Prong of the Little Red River, then up the Caprock Escarpment to the top of the High Plains on Haynes Ridge. This ridge is the drainage divide between the North and South Prongs of the Little Red River. The trail then leads down the escarpment into the valley of the North Prong of the Little Red River. From there, you proceed downstream, and eventually cross over the divide, back to the trailhead where you began.

In the following trail log, features of geological interest are pointed out for the first part of the hike--up the South Prong to the top of the High Plains. For the second (return) part of the hike--down the North Prong and back, we invite you to enjoy finding the stratigraphic contacts and features pointed out on the trip up!
STOP 1. Lower Part of the Quartermaster Formation

At this locality, the lower part of the Permian Quartermaster Formation is well exposed in a cut-bank on the South Prong of the Little Red River. This part of the Quartermaster Formation is thinly to very-thinly bedded siltstone and sandstone. A variety of original depositional features can be observed. Several kinds of soft-sediment deformation (convolute lamination, load casts) are abundant, and these indicate that the sediments were originally water-saturated. Wave ripples, current ripples, graded bedding, and cross-lamination also indicate that the sediments were deposited subaqueously. The Quartermaster Formation, like other Permian "red bed" deposits surrounding the Permian Basin, is thought to represent an arid region coastal tidal flat environment (see Chapter 3).

Try to find some examples of some of these sedimentary structures.

Several interesting secondary (post-depositional) features are also evident at this outcrop. Notice the pale green color blotches and color bands that generally follow bedding in these otherwise red strata. This illustrates a phenomenon known as "reduction spotting." Reduction spots are a diagenetic feature common in red-bed deposits. They are thought to form where groundwater causes the iron in the red sediment—normally in the ferric (oxidized) state—to undergo reduction to the ferrous (reduced) state. This causes the color to change from red to green. Small bits of organic material distributed along bedding planes may have set up locally reducing micro-environments to produce the restricted spots.

The most striking post-depositional features shown at this outcrop are the gypsum-filled veins. These are common in the lower part of the Quartermaster Formation, but are rarely formed higher in the section. Fibrous crystals of "satinspar" gypsum fill fractures that run parallel to the bedding and cut across bedding at angles of 30-60° to horizontal. There is also brecciation and small-scale faulting associated with the veins. Notice that most of the veins have a discolored zone running down the midline, with fibrous crystals on either side. This medial "scar" is thought to mark where the earliest crystals began forming. As the fractures continued to open, the mineral crystal fibers grew
toward the vein/rock contact (Collins, et al., 1986).

**What do you think?**

Most of the crystal fibers are straight and the crystals merge between horizontal and inclined veins—indicating that opening of the fractures and faulting occurred before mineralization. Some veins have curved crystal fibers, suggesting that shear may have occurred along faults during or after crystallization (Collins, et al., 1986).

The fracturing, faulting, and gypsum mineralization is thought to reflect the subsidence and collapse of these strata as groundwater dissolves salt beds in the underlying Permian section. As the salt beds dissolve, the overlying rock undergoes vertical extension, splitting along bedding planes, and forming conjugate fractures at angles to the bedding. Saline groundwater then precipitates gypsum in the open fractures.
1. ORIGINAL DEPOSITS
- thinly-bedded ss/sh
- gypsum/salt

2. HORIZONTAL COMPRESSION
- vertical fractures form

3. VERTICAL EXTENSION
- horizontal & inclined fractures form
- fractures fill with "satin spar" gypsum
- salt beds removed in solution by groundwater

STOP 1
STOP 2. Holocene Alluvium and Terraces of the South Prong of the Little Red River

At this locality, some of the Recent alluvial sediments in the valley of the South Prong of the Little Red River are exposed. The South Prong has cut a picturesque canyon about 200 m (660 feet) deep with steep valley walls and a relatively flat valley floor. Most or all of this erosion was accomplished during Holocene time (the last 10,000 years) and the floor of the valley is mantled with alluvium (river sediments) of Holocene age. Holocene colluvium (material deposited by sheetwash and mass wasting), is present locally along the flanks of the valley. The broad, flat valley floor was probably produced by lateral migration of the meandering South Prong. At this locality, Holocene alluvium is exposed in a cut-bank incised into the lower of two stream terraces in this valley. A higher stream terrace can be seen farther up the valley, at the next stop. Radiocarbon dates from soil organic material in this alluvium, at a site about four miles downstream, indicates that the valley was cut prior to about 2000 years ago (Baumgardner, 1986). As much as 50 feet of sediment then accumulated on the valley floor until about 1000 years ago. At that time, the stream incised down into its older deposits, forming the high terrace. Sediment then accumulated again in the lower incised floodplain. Recently, (perhaps no more than a few hundred years ago) the stream has incised to a lower level, forming the lower terrace we see at this locality.

Recent work (Hall, 1990) indicates that the period of channel down-cutting that formed the high stream terraces about 1000 years ago, occurred over much of Texas and Oklahoma at the same time. This coincides in time with a change to an arid climate over this region. There has been much debate about the cause of more recent episodes of channel trenching. Some authors believe that overgrazing and deforestation of the valley by man is the cause; others suggest continuing climatic change.

The Holocene alluvium exhibits crude horizontal bedding, and imbrication of gravel. Similar features can be seen in the modern stream sediments. Primary current lineation, current ripples, and sets of small antidunes are also evident in the present stream bed.

Things to see: Imbrication, primary current lineation, ripples, antidunes.
Geologic map of the upper Little Red River drainage basin (from Baumgardner, 1986). Arrow indicates STOP 2.
STOP 2

SEQUENTIAL DEVELOPMENT OF THE VALLEY OF THE LITTLE RED RIVER

(from Baumgardner, 1986)
STOP 3. Upper Part of the Quartermaster Formation

At this locality, a view of several features in the upper part of the Permian Quartermaster Formation is provided. Unlike the uniform, thin, and horizontally bedded lower part of the formation, the upper part exhibits large sets of lenticular, inclined bedding and very thick "massive" bedding. In the canyon walls on both sides of the valley, large sets of inclined layers or cross-bedded intervals can be seen. These inclined beds are well exposed in several thin buttresses projecting out of the canyon wall. The interpretive exhibit at the trailhead suggests that these inclined beds formed as deltas built seaward depositing sediment on the inclined front of the delta ("delta front" deposits). However, we believe that the inclined beds may instead be deposits formed by lateral (sideways) migration of meandering tidal channels, with sediment deposited on the inclined inner bank ("point bar"). This is suggested by the fact that: 1) the inclined beds occupy lenticular channels; 2) the base of sets of inclined beds rests on an erosional surface; 3) sets of inclined beds dip in different directions.

What do you think?

Above the interval with inclined bedding, are very thick sandstone beds that hold up nearly vertical cliffs. These beds appear, from a distance, to be structureless ("massive"). However, on close inspection, you will be able to see climbing cross-lamination and parallel-lamination on fresh surfaces. Farther up the canyon, and on the back part of the trail down the North Prong, you will have a chance to get a close look at these beds.

Note: On the south side of the canyon you can see the high Holocene terrace level mentioned at Stop 2.
DELTA PROGRADATION?

- coarsening upward
- gradational base

OR

LATERAL CHANNEL MIGRATION?

- fining upward
- erosional base

ORIGIN OF INCLINED BEDDING IN UPPER QUARTERMASTER
FRACTURE ORIENTATIONS FROM SURFACE LOCATIONS (white) AND TEST WELLS (dark) in the Texas Panhandle and eastern New Mexico (from Gustavson & Finley, 1985).
STOP 4. Regional Fractures in the Quartermaster Formation

At this locality, you can view examples of an extensive regional fracture system found over the entire Southern High Plains. The massive brittle sandstones in the upper part of the Quartermaster Formation display these fractures particularly well, although similar fractures also penetrate overlying Triassic strata. These fractures are, for the most part, vertical (perpendicular to bedding) and they display a strong northeasterly orientation. A major systematic set is about N30°E, with lesser sets at angles to that. The horizontal and inclined gypsum-filled fractures observed in the lower part of the Quartermaster Formation (Stop 1) cut across these vertical fractures, suggesting that the regional vertical fractures formed before widespread salt dissolution (Collins et al., 1986). The systematic vertical fractures do not run parallel to the canyon walls, suggesting that pressure release and rock expansion ("unburdening") following erosion could not have caused the fractures. The best explanation for the formation of these sort of fractures is horizontal tectonic compression. Regional extension fractures are thought to form at low stresses, parallel to the direction of maximum compressive stress (Lorenz, et al., 1991). The High Plains fracture system thus may have formed when the region experienced northeastward compression during the Laramide Orogeny (Late Cretaceous-Early Tertiary).

When lighting conditions are good, "plumose structure" or frondescent "arrest lines" can be seen on the fracture surfaces. These are curved steps on the fracture surfaces that radiate from points (on bedding planes or fracture intersections) in the direction of fracture propagation.

Can you see them?

(from Lorenz et al., 1991)

load-parallel extension fractures
(from Collins et al., 1986)
STOP 5. Overview of the Triassic Section

At this locality, a good view of the Triassic stratigraphy of the High Plains is provided in a tributary valley north of the trail. Triassic strata are subdivided into three formations, the Tecovas, Trujillo, and Cooper Canyon formations. Together these comprise the Dockum Group, which is entirely Late Triassic in age. Although the Dockum Group is very thin in this area, all three formations are present (see Chapter 4). As you continue up the trail, you will have a chance to see each of these stratigraphic units more closely.

The contact between multicolored shale and sandstone of the Triassic Dockum Group, and the underlying red-orange sandstone of the Permian Quartermaster Formation occurs at the break in slope of the valley walls. Resting unconformably on the cliff-forming Quartermaster Formation, is a soft, slope-forming, pink-weathering, sandstone of uncertain age. No fossils have been found in this sandstone, and it could be either Late Permian or Early Triassic in age. We are uncertain whether to place it in the Quartermaster Formation or overlying Dockum Group. We refer to it here as the "P/Tr aeolian sandstone." Resting unconformably on this aeolian sandstone is the Tecovas Formation—the lowermost part of the Dockum Group. In this area, the Tecovas Formation is composed mostly of red-weathering, slope-forming shale. A thin, white-weathering, highly quartzose, buried soil (paleosol) is present along the base of the Tecovas Formation. Resting unconformably on the Tecovas Formation are tan-colored, cliff-forming, sandstones of the Trujillo Sandstone. The Trujillo Sandstone holds up the second major set of cliffs along the Caprock Escarpment. Gradationally overlying the Trujillo Sandstone is red, slope-forming, shale of the Cooper Canyon Formation. The Cooper Canyon Formation is very thin in this area. It was mostly eroded away prior to deposition of the overlying Ogallala Group.

Note: As you go up the trail to the top of the Quartermaster Formation, notice the fallen boulders in the colluvium on the slope—some of which are "pedestalled" by erosion, and now sit perched on spires.
STOP 5

TO CANYON FORMATION

TRUJILLO SANDSTONE

TECOVAS FORMATION

COOPER CANYON FORMATION

QUARTERMETER FORMATION
STOP 6. **Permo-Triassic Aeolian Sandstone**

At this locality, we will inspect an unusual sandstone unit resting between *bona fide* Permian strata (Quartermaster Formation) below, and Triassic strata (Tecovas Formation) above. This pink-weathering, fine-grained sandstone forms a slope above the cliffs held up by the Quartermaster Formation. In most areas it is about 10 meters thick, and has thin interbeds of red shale. In some places, however, thick beds of red shale, similar to that in the overlying Tecovas Formation, separate the sandstone layers. This unit is cut out by erosion both to the north and south, where the Tecovas Formation rests directly on undisputed Quartermaster Formation.

The sandstone beds exhibit very large-scale cross-bedding (sets over 1 meter thick) produced by aeolian dunes and parallel-lamination produced by wind ripple migration. This unit was probably deposited in an aeolian dune field. The red shale interbeds may represent pond deposits that formed between dunes. The dip direction of dune cross-bedding indicates that the paleo-wind direction was to the north.

No fossils have been found in this unit but we are inclined to believe that it may be Early Triassic in age because it is interbedded with red shale similar to that of the Tecovas Formation.

What do you think?

**Note:** The unusual pea-sized "knots" on weathered surfaces of this sandstone are produced by patches of resistant calcite cement (poikilotopic) that weather in relief.
TRIASSIC STRATIGRAPHY on the Upper Canyon Trail, showing the P/Tr aeolian sandstone (P/Tr), Tecovas Formation (Tr1), Trujillo Sandstone (Tr2), and Cooper Canyon Formation (Tr3). The base of the Ogallala Group (To) is also indicated.
TRIASSIC STRATIGRAPHY

STOP 9
COOPER CANYON FORMATION

STOP 8
TRUJILLO SANDSTONE

STOP 7
TECOVAS FORMATION
paleosol

STOP 6
P/Tr aeolian sandstone

TOP OF QUARTERMASTER FORMATION

METERS
0
10
20

BASE OF OGA LALA GROUP

DOCKUM GROUP
STOP 7. Paleosol at the Base of the Tecovas Formation

At this locality, we will inspect a buried soil horizon (paleosol) that marks the base of the Tecovas Formation. In most areas, the base of the Tecovas Formation has a thin, white-weathering, coarse-grained, very quartzose sandstone bed with scattered chert pebbles. This sandstone bed has purple, red, and yellow color mottling and root-like columnar structures suggesting that it has been significantly modified by soil-forming processes. In places, the top of the sandstone is impregnated with nodules and layers of calcite (paleo-caliche) that are locally replaced by silica ("silcrete"). This buried soil probably records an extended period of weathering and non-deposition that took place in mid-Triassic time. In the Colorado Plateau region of Arizona and Utah, a strikingly similar series of buried soils known as the "mottled strata" generally separates the Moenkopi Formation (mid-Triassic) from the Chinle Formation (Late Triassic).

How can you recognize a buried ancient soil (paleosol)?

Note: Above this buried soil horizon are red shales characteristic of the Tecovas Formation, which is very thin in this area. The base of the Trujillo Sandstone can also be seen from here.
STOP 8. The Trujillo Sandstone

At this locality, we will examine features of the Trujillo Sandstone, which forms the second set of cliffs in the canyon. Most of these features are nicely illustrated in the fallen boulders along the trail.

The Trujillo Sandstone consists of thick beds of fine-grained sandstone and conglomerate, with interbeds of red shale. Unlike the clean quartzose sands in the base of the Tecovas Formation, the Trujillo sandstones are rich in rock fragments and mica flakes. The conglomerate beds are composed mostly of carbonate nodules (reworked paleo-caliche nodules) and pebbles of older sedimentary rocks.

In contrast to the uniform horizontal bedding in the Quartermaster Formation, the Trujillo Sandstone consists of thick, lenticular, channel-like beds that pinch-out laterally and cut across one another. Trough cross-bedding and current ripple cross-lamination is common, particularly in the upper parts of sandstone beds. The Trujillo Sandstone represents the deposits of a major river system that flowed to the northwest through the High Plains region in Late Triassic time.

Note: Nice examples of carbonate nodule conglomerate and cross-bedded micaceous sandstone can be seen in fallen Trujillo boulders next to the trail.
STOP 9. Overview from the Top of the Trail

From the scenic rest-stop and overlook at the top of the trail, you can get a better look at several interesting features pointed out on the way up. From here, you can see the wide, relatively flat, valley floor of the South Prong and the higher and lower stream terraces mentioned at Stop 2. Notice the nice meander loops in the South Prong. Some of the meander loops are in the process of being "cut-off" by migration of the channel. Notice also how the physiography of the canyon walls reflects the stratigraphy—the Quartermaster Formation and Trujillo Sandstone holding up nearly vertical cliffs, while the recessively weathering Tecovas Formation forms a slope between them.

The Cooper Canyon Formation overlies the Trujillo Sandstone, and consists predominantly of red shale. Although the Cooper Canyon Formation is very thin in this area—most of it was eroded away before deposition of the overlying Ogallala Formation—you can still see a bit of slope-forming red shale just above the top of the Trujillo cliffs. Many of the interesting vertebrate fossils found in the Dockum Group were discovered in the Cooper Canyon Formation (see Chapter 4).

Note: As you proceed up the trail, keep your eye out for the base of the Ogallala Group, which is marked by thick beds of conglomerate. Pebbles weather free from the conglomerate and litter the trail and slope.

You may find scattered Cretaceous oyster shells along here. Where in the world do they come from?
STOP 9. VIEW FROM THE TOP OF THE TRAIL, showing the meandering South Prong of the Little Red River, canyon physiography, and the distribution of Holocene alluvium and colluvium.
STOP 10. Ogallala Group

From this vantage point you can see most of the exposed section of the Ogallala Group which is about 50 meters thick in this area. Elsewhere on the High Plains the Ogallala is subdivided into several formations; however, in this area no subdivisions are recognized and the Ogallala is mapped undivided (see Chapter 5). No Jurassic, Cretaceous, or early Tertiary strata are present in this region, and the Ogallala Group (late Miocene to early Pliocene in age) rests unconformably on Triassic strata.

The lower part of the Ogallala in this area is composed of thick beds of tan, coarse-grained, cross-bedded sandstone and conglomerate. These conglomerates are exposed on both sides of the trail on the way up, and are also nicely exposed on the trail on the way back down the other side. The conglomerate is composed of pebbles of sedimentary rocks, chert, quartzite and other metamorphic rocks, and fine-grained volcanic rocks. These were derived from Precambrian rocks exposed in the Sangre de Cristo Mountains of northern New Mexico and from Tertiary volcanic rocks in northeastern New Mexico. Interestingly, you may also find reworked Cretaceous oyster shells in these conglomerates. This part of the Ogallala was deposited by rivers flowing southeastwardly from the rising Rocky Mountains in Miocene time, across the High Plains region. Bones of Miocene elephants, horses, rhinos, and bone-crushing bear dogs are particularly common in these gravels (see Chapter 5).

Overlying the coarse sandstone and conglomerate of the lower Ogallala are massive and thinly bedded, fine-grained, pink or flesh-colored sandstones and siltstones of the upper Ogallala. These fine-grained sediments weather recessively, forming a vegetated slope. Nice exposures can be viewed northeast of the trail. Small carbonate nodules (paleo-caliche) are abundant in these sediments. These deposits are thought to reflect aeolian "sand sheet" accumulation on grass-covered plains where vegetation inhibits dunes from forming. The abundant paleo-caliche nodules indicate that soil development occurred in the sediments as they accumulated. Bones of Miocene grazing herbivores such as horses, camels, and oreodonts are common in these sediments (see Chapter 5). A single, unusual, bioturbated limestone bed holds up a ledge near the top of the section. This appears to be a small lake or pond deposit.
Take a look at the burrows in this bed near the trail.

Note: The thick beds of sandstone and conglomerate in the Ogallala Group are the major freshwater aquifer on the High Plains (see Chapter 5).
STOP 12
modern soil
volcanic ash bed

BLACKWATER DRAW FORMATION
CAPROCK CALICHE
burrowed limestone bed

STOP 11
OGALLALA GROUP

STOP 10
aeolian facies
fluvial facies
top of
DOCKUM GROUP
STOP 11. The Caprock Caliche

At this locality, you can examine the "Caprock" caliche which marks the top of the Ogallala Group. The resistant Caprock is about 3.5 m thick in this area, and holds up the highest cliff along the High Plains escarpment. This is one of the best examples in the world of a mature caliche ("calcrete") profile. The Caprock caliche formed in mid Pliocene time, after deposition of the Ogallala ceased, during a period of landscape stability that resulted in prolonged soil development over the region. It was subsequently modified through the Quaternary.

A caliche is basically a limestone formed by soil processes. In semi-arid regions, dissolved carbonate is transported through the upper part of the soil profile, but it precipitates at depth, initially forming small nodules of calcite. With continued time, the nodules gradually coalesce and "plug up" the soil, impeding the flow of water. At that point calcite begins precipitating in horizontal layers on top of the plugged horizon. In the most mature caliche profiles, dissolution pipes may form and the laminated layers may be broken up to form breccia that is later coated with caliche. The Caprock is a good example of just such a mature caliche. From the trail you can see examples of dissolution pipes penetrating the caliche.

Resting above the Caprock caliche (the top of the Ogallala Group) is the Blackwater Draw Formation. This unit is only about 8 meters thick here, and consists of unconsolidated, red, fine-grained sand and silt with layers of soil-formed carbonate nodules. These sediments were formerly called the Recent "cover sands" of the High Plains (see Chapter 6). The "cover sands" were deposited mostly by wind action, blowing sediment northeastward from the Pecos River Valley throughout Quaternary time. Fertile soils, formed in the upper meter or so of the Blackwater Draw Formation, and watered through irrigation by pumping from the underlying Ogallala aquifer, support the agricultural industry of the High Plains.
STOP 12. Volcanic Ash Bed in the Upper Part of the Blackwater Draw Formation

At this locality, a thick bed of white-weathering volcanic ash is exposed in the upper part of the Blackwater Draw Formation. The ash bed was discovered during preparation for this fieldtrip, and has not been reported previously. Underlying the ash is a layer of dark organic-rich clay. The ash bed itself is up to 1.5 meters thick, with thin laminae of red silt, and small burrows. These observations together suggest that a thin layer of ash, deposited over a broad area, was reworked into a small playa lake and concentrated to greater thickness. Erosion has removed most of the surrounding deposits, leaving this remnant topping the hill.

Ash beds have been found elsewhere on the High Plains in the Blackwater Draw Formation, and in lake deposits that underlie or intertongue with the Blackwater Draw (see Chapter 6). These ashes were erupted at several times during the Pleistocene from the Jemez volcanic center in northern New Mexico and from the Yellowstone area in Wyoming. We do not know which of these ashes is exposed here.

This is the highest point in the section exposed in Caprock Canyons. As you continue along the trail into the valley of the North Prong, you will be going down through the section you have just seen.

Try to locate each of the stratigraphic units we have mentioned. You will find nice examples of many of the features we’ve discussed—and more!

Note: Be sure to stop and see the springs at Fern Cave. From what stratigraphic level does the spring flow?

HAVE FUN AND WE’LL SEE YOU BACK AT THE TRAIL HEAD.
1. Thin layer of ash deposited

2. Ash reworked into playa lake

3. Lake deposits buried

4. Deposits exhumed by erosion

STOP 12
SUPPLEMENTARY ROADSIDE STOPS

Eagles Point Overlook

Eagles Point overlook is on the main park road in Caprock Canyons State Park, between Lake Theo and the Canyon Loop trailhead, where the road crosses a deeply incised tributary of the South Prong of the Little Red River.

The contact between the Permian Whitehorse Group (below) and Permian Quartermaster Formation (above) can be observed in the deep arroyo where the park road crosses over (see Chapter 3). The Whitehorse Sandstone consists of thick beds of massive friable sandstone exposed in the lowermost part of the arroyo. The thick beds of laminated gypsum near the top of the roadcut are identified as the Cloud Chief Gypsum; this marks the top of the Whitehorse Group. The lower part of the overlying Quartermaster Formation caps the hills here, and consists of more thinly bedded and indurated sandstone and siltstone. The gypsum-filled veins in the Quartermaster Formation are clearly secondary (post-depositional) in origin (see Trail Log, Stop 1). However, the Cloud Chief Gypsum beds are primary deposits resulting from evaporite precipitation in saline coastal ponds along the shore of the hot, arid, Permian Seaway. The alternating dark clastic/organic laminae and white gypsum laminae may reflect seasonal sedimentation. Nice examples of soft sediment deformation can be seen in these laminated gypsum beds.

From the scenic overlook, you can see subtle folding or warping in these Permian strata. On close examination, fractures, small-scale normal and reverse faults and zones of chaotically deformed strata are apparent. Detailed structural mapping of the area (Collins, et al., 1986) has delineated a number of elongate to circular synclinal depressions. This deformation is thought to reflect the dissolution of underlying Permian salt beds, and subsequent collapse of the overlying strata into the depressions (Collins, et al., 1986).

Salt dissolution is thought to have been an important geomorphic process shaping the Rolling Plains region throughout the Quaternary, and up to the present day.
EAGLES POINT OVERLOOK
STOP

LAKE THEO ARCHEOLOGICAL SITE
STOP

MAP SHOWING STRUCTURAL FEATURES NEAR EAGLES POINT OVERLOOK (from Collins and others, 1986).
Lake Theo Archaeological Site

Lake Theo is a spring-fed reservoir on Holmes Creek, the southernmost tributary of the Little Red River in Caprock Canyons State Park. The hills on either side of Holmes Creek at Lake Theo are capped by unconsolidated Quaternary alluvial and aeolian sediments of the Lingos Formation (see Chapter 6). These sediments are older than the inset Holocene stream terrace deposits (2,000 years old and younger) observed at lower levels in the stream valleys (see Trail Log, Stop 2). Radiocarbon dates from these deposits indicate that they range in age from about 11,000 to 3,500 years old (Harrison and Killen, 1986).

On the west side of the lake, these older sediments preserve an important archaeological site, which was discovered and excavated during the 1970’s. The lowermost level at the site, a horizon dated as about 9,000 years old, preserves a bone bed of extinct bison (Bison antiquus) with a campsite and butchering station having tools, projectile points, and other artifacts of the Folsom culture (Harrison and Killen, 1986). The Folsom culture was one of the earliest native American (Paleo-indian) cultures. Only the Clovis culture (about 11,000 years before present) is older. Along with other Folsom cultural remains, an unusual (?) ceremonial feature was discovered—a small round hole with bison jaws and leg bones arranged in "standing-on-end" position.

Younger stratigraphic levels at the site yielded artifacts of the Plainview, Archaic, and Historic cultural periods. Apparently, springs discharging from the Ogallala aquifer (see Chapter 5) into Holmes Creek, have provided a permanent source of fresh water in this area for thousands of years. This attracted herds of bison, and made for an ideal hunting/camping area for the early American big game hunters (Harrison and Killen, 1986). These sort of stratified archaeological sites, with materials from successive cultures superimposed in stratigraphic order, are very important in allowing confident dating of the prehistoric human inhabitation of North America.
GEOLOGIC MAP OF LAKE THEO VICINITY
(from Baumgardner, 1986)

CHRONOLOGY OF HUMAN OCCUPATION ON THE HIGH PLAINS
(from Gustavson, 1986)

**Historic**
- A.D. 1540 - A.D. 1810: Early exploration, mainly Spanish

**Prehistoric**
- A.D. 1 - A.D. 1540: Neo-Indian
- 5000 B.C. - A.D. 1: Archaic
- 8000 B.C. - 5000 B.C.: Late Paleo-Indian
- 9000 B.C. - 8000 B.C.: Folsom
- 10,000 B.C. - 9000 B.C.: Clovis
- Pre-10,000 B.C.: Pre-Clovis?
Playa Lakes

There are over 30,000 small, circular playa lake basins scattered over the surface of the Southern High Plains in Texas and New Mexico. Such lakes are an important recharge zone for the underlying Ogallala Aquifer (see Chapter 5). Most of these basins have a diameter on the order of less than a kilometer or so, and are dry part or most of the year. The vast majority of the basins are less than 10 to 20 thousand years old, and have several meters of dark organic-rich lacustrine clay along the floor of the lake, which rests on the Pleistocene Blackwater Draw Formation (the aeolian "cover sands" of the High Plains--see Chapter 6). Larger saline lake basins are fewer in number and may contain more than 20 meters of lacustrine sediment resting on the underlying Ogallala, Cretaceous, or Triassic strata (Reeves, 1990). These larger lakes formed as long ago as the early Pleistocene.

The origin of these numerous, small lake basins has been a source of debate and controversy for decades. Recently, Reeves (1990) summarized information on these lakes and presented a general sequence for their development. Basins are thought to originate in low areas on the rolling surface of the aeolian cover sands--perhaps accentuated buffalo "wallows" where animals rubbed dust in their hides and paved depressions filled with rain water (Reeves' Type I Basin). As water and organic material accumulate in the depression, mildly acidic water is generated which, as it percolates down through the underlying sediment, dissolves soil-formed carbonate and the Caprock caliche. This causes the land surface to subside, deepening and widening the original depression (Reeves' Type II Basin). When the lake dries up, the wind may blow sediment out of the basin, deepening it, and creating dunes ("lunettes") on the down-wind side. In some cases, groundwater may infiltrate to greater depth along fractures, leading to dissolution of Permian salt beds and further deepening and widening of the lake basin (Reeves' type III Basin).
KNOWN TYPE III LAKE BASINS ON THE SOUTHERN HIGH PLAINS (from Reeves, 1990). Stippled areas show where the Ogallala Group exceeds 61 m in thickness.

REEVES' SUGGESTED SEQUENTIAL DEVELOPMENT OF LAKE BASINS ON THE HIGH PLAINS (1990).
Pleistocene Lake Beds in the Tule Formation

The Tule Formation is exposed along Tule Creek and its tributary Rock Creek, which cut a deep canyon into the Caprock Escarpment northwest of Caprock Canyons State Park in western Briscoe County. We will examine the Tule Formation in a roadcut on State Highway 207 along the south side of Tule Canyon.

The Tule Formation consists predominantly of drab greenish-gray clay and unconsolidated, pale yellowish sands and silts, with thin indurated beds of bioturbated limestone and dolostone. In the roadcut, you can see large-scale soft-sediment folding in the Tule Formation (? slumping), and the contact with the underlying Triassic Dockum Group. The overlying Blackwater Draw Formation can be seen to the east. The Tule Formation is a good example of the sort of deposits found in one of the large Pleistocene lake basins on the Southern High Plains. These lake deposits are very fossiliferous, and provide a record of the ice age fauna of the High Plains—dominated by horses, camels, elephants, sloths, and glyptodonts (Schultz, 1990). The lower part of the formation has the 1.2-1.3 million year old Cerro Toledo Ash bed (erupted from the Jemez Mountains in New Mexico) and the upper part of the formation has the 600,000 year old Lava Creek B Ash bed (erupted from Yellowstone Park). Hence, the Tule Formation spans a good bit of Pleistocene time.

The origin of these large Pleistocene lake basins is something of a mystery. In the center of the lake basin the base of the Tule Formation rests unconformably on Triassic strata of the Dockum Group. Along the margins of the basin, the Tule Formation rests on the Ogallala Group. This suggests that some erosion must have occurred, to "scour out" a depression in which the lake formed. If a stream eroded the depression, it must have been dammed up later (and for almost a million years!) to allow a lake to form. If the wind hollowed out the depression, some evidence for a dune field should be found around the lake margin. If dissolution of underlying Permian salt beds caused the basin to form, then how do you explain the "missing" Ogallala section beneath the basin?

These are some of the competing theories that seek to explain the origin of these large lake basins (Gustavson, 1986). Suffice it to say that no consensus has yet been reached!
GEOLOGIC MAP OF THE TULE LAKE BASIN AND VICINITY (from Gustavson, 1986).
BIOSTRATIGRAPHY OF THE TULE FORMATION
(from Schultz, 1990)

1. END OF OGALLALA DEPOSITION
   - OGALLALA GROUP
   - DOCKUM GROUP

2. EROSION OF BASIN?
   - aeolian?
   - fluvial?

3. FILLING OF BASIN
   - BLACKWATER DRAW FORMATION (aeolian)
   - TULE FORMATION (lacustrine)

4. EROSION OF TULE CANYON
   - present landscape

DEVELOPMENT OF THE TULE LAKE BASIN

QUATERNARY
Tule Formation

- Lava Creek B (=Peariette Type O) Ash (0.6 m.y.)
- Mayfield Ranch Local Fauna
- Sloth-Camel Quarry (Rock Creek Local Fauna)
- Gidley's Equus scotti (=Horse) Quarry

- Stegomastodon, Mammutus, Glossotherium, Glyptodont (Martin Ranch Local Fauna)
- Cerro Toledo-X Ash (1.2-1.3 m.y.)
  Equus (Dolichohippus) simplicidens, Equus semilicatus, Camelops, Mammutus, Glossotherium (Martin Ranch Local Fauna)

TRIASSIC
- Dockum Group
  Trujillo Formation

Molluscan fauna
Vertebrate fauna
Volcanic ash
The Caprock Canyons area in Briscoe County lies along the eastern side of the Palo Duro Basin. The Palo Duro Basin includes parts of fourteen counties in the Texas Panhandle (Figure 2.1). This is a relatively shallow intracratonic basin bounded by the Amarillo-Wichita uplift on the north, the Matador Arch on the south, the Roosevelt Positive to the west, the Bravo Dome to the northwest, and the Hardeman Basin to the east. These structural features were active primarily in late Paleozoic time, and the main period of subsidence in the Palo Duro Basin was during the Pennsylvanian. There has, however, been recurrent activity on several basin-bounding structures as recently as the Tertiary. The deepest part of the Palo Duro Basin is in the southeastern part of Floyd and the southwestern part of Motley counties, near the Matador Arch.

Precambrian

Precambrian basement rocks in northeastern Roosevelt county, New Mexico, and Castro, northern Parmer, northern Lamb, Swisher, northern Hale, western Floyd, Briscoe and western Donley counties in Texas, are gabbro and diabase. The Swisher Gabbroic Terrane is of Late Proterozoic age. The gabbro appears to be a lopolith that occupies a sag or syncline in the older Precambrian volcanic rocks that cover a more extensive basement section in the Panhandle region (Flawn, 1956).

Cambrian

The Cambrian (?) Hickory Sandstone represents the initial transgression of the Cambrian sea on the Texas craton. In the Palo Duro Basin, this consists of less than 15 m of terrigenous, quartz-rich sandstone, and shales with lenses of dolomite or limestone (Ruppel, 1985). The thickest accumulation of Cambrian sediment is found at the intersection of Hale, Floyd, Briscoe and Swisher counties. The greatest aerial accumulation of sediments occurs in the area of Armstrong, Donley, Briscoe, Hall and Motley counties. The Anadarko Basin to the north was rifted initially during the Cambrian, yet little sediment accumulated in the Palo Duro Basin during that time. (Dutton, 1980).
Figure 2.1 Major structural elements of the Texas Panhandle and surrounding area (from Gustavson and Finley, 1985). Inset shows Texas counties in this region. Arrows indicate field trip area.
Ordovician

The Ordovician Ellenburger Group consists of shallow shelf carbonates. These reach a maximum thickness of 150 m and are present throughout the Palo Duro Basin except where they have been locally removed by erosion. The Ellenburger ranges from fine to coarse-grained sucrosic dolomite, to oolitic limestone. In many places it contains shale and thin quartz sandstones (Ruppel, 1985). The Ellenburger was deposited on a shallow marine shelf that covered a large part of North America.

Mississippian

Silurian and Devonian rocks are missing in the Palo Duro Basin. Mississippian shelf carbonates, unconformably overlie Precambrian, Cambrian (?) and Ordovician rocks, with a maximum thickness of 330 m in Childress County (Dutton, 1980). The Mississippian System of North America is made up of four series: Kinderhookian, Osagean, Meramecian, and the Chesterian. Because the Palo Duro Basin has little biostratigraphic control, Mississippian rocks in the basin are subdivided using informal lithostratigraphic units (Ruppel, 1985).

Kinderhook rocks are quartzose sandstones found only in the northern parts of the basin in Donley and Collingsworth counties. These rocks represent basal transgressive sedimentation formed at the beginning of Mississippian time (Ruppel, 1985).

Osage rocks are the most extensive of all the Mississippian rocks in the area with a maximum thickness of 90 m. Osage rocks are gray to brown, commonly argillaceous, cherty limestones and dolomites (Ruppel, 1985). Osage rocks of the Palo Duro Basin are difficult to correlate, but are roughly equivalent to the Chappel and St. Louis formations in the Hardeman Basin to the east. Environmental conditions varied across the basin at this time. Cores from Osage rocks in Childress County show wackestones with alternating layers of grainstones. Cores from Donley County have layers of argillaceous dolomite and coarse-grained skeletal grainstones. Ruppel (1985) interpreted conditions in Childress County to represent deep water sedimentation, with carbonate skeletal material being moved by turbidity currents. Conditions in Donley County are thought to represent a shallow inner shelf platform environment. Regionally, the Osage rocks show a east-to-west shallowing of water.

The top of the Meramec is recognized by a marked increase in resistivity in geophysical logs. Meramec rocks are 90 m thick, white to buff-colored, fine- to medium-grained limestone, with a fine-grained sandstone common at the top of the unit (Ruppel, 1985). In the Dalhart and western Anadarko Basins, the Meramec is divided into the
upper Ste. Genevieve, middle St. Louis and lower Spergen-Warsaw, but only the Ste. Genevieve is easily recognized in the Palo Duro Basin. Meramec sediments were interpreted by Ruppel (1985) as carbonate shoal deposits, with the basin becoming progressively more shallow because of the increasing amount of sand.

Late Mississippian to early Pennsylvanian erosion left the Chester present only in the central and eastern part of the Palo Duro Basin. The Chester is a fossiliferous, oolitic limestone with interbedded shales and has a maximum thickness of 90 m. Shale content increases in the west central part of the basin. Chester depositional conditions were shallow-water marine with the source of the clastic sediments coming from the northwest (Ruppel, 1985).

Pennsylvanian

The major structural features of the Palo Duro Basin were developed during late Mississippian and early Pennsylvanian time. Both the Matador Arch and the Amarillo-Wichita Uplift were block faulted, and uplifted to provide sources for sediments. This faulting occurred during the deformation of the Southern Oklahoma Aulacogen. Deformation of the Aulacogen coincided with the inferred closing of the Paleozoic Gulf of Mexico (Dutton, 1980). During Pennsylvanian time the Hardeman and Palo Duro Basins were one unit. Sediments thicken toward the Hardeman Basin (Budnik and Smith, 1982).

The Pennsylvanian System of North America is divided into five series: Morrowan, Atokan, Des Moinesian, Missourian and Virgilian. There are no widespread unconformities or regional marker beds in the Pennsylvanian of the Palo Duro Basin. There is, however, a vertical change in facies between the lower units and the upper units (Dutton, 1980).

Lower Pennsylvanian rocks, the Bend and Strawn groups (Morrow, Atoka and Des Moines series), show more terrigenous clastics than limestones. The three depositional systems were: fan deltas, shallow marine deltas and the deep basin environment. Early "granite wash" sedimentation was controlled by the tectonic uplifts that surrounded the basin. The "granite wash" sediments have granite fragments and feldspar grains from the Precambrian rocks of the Amarillo-Wichita uplift and were deposited in fan deltas. A fan delta is a much like a alluvial fan that progrades into a body of water. Dutton (1980) described the fan delta systems of the Palo Duro Basin as deltas with ephemeral discharge and small drainage areas. Because, deposition on the fan deltas was not constant, carbonate bioherms grew on the clastic sediments deposited by the deltas. Later cycles of clastic deposition would be deposited between the bioherms.
Parts of the basin that were far enough away from the sources of sediments were sites for the accumulation of thin shelf carbonates and muds. The thickest area of this shelf sedimentation is found in Cottle, King, Briscoe, Floyd, Randall and Swisher Counties (Dutton, 1980).

Upper Pennsylvanian rocks, the Canyon and Cisco groups (Missourian and Virgilian series), represent a deep basin surrounded by carbonate shelves with small fan deltas. The faulting and uplift stopped during the late Pennsylvanian, and the influx of clastic sediment slowed. Most of the thick shelf limestones were produced by phylloid algae, forming mounds and trapping carbonate mud and debris. Crinoids, bryozoans, fusulinids, echinoids, sponges and brachiopods lived with the algae. Diagenesis has transformed the shelf margin limestones to brown, medium to coarsely crystalline dolomite with vuggy porosity and some chert (Dutton, 1980).

The main sources of clastic sediment at this time were in the Wichita Mountains to the east. Sediments were transported into the Palo Duro Basin through the Hardeman Basin. Dutton (1980) compares the late Pennsylvanian clastic influx to the modern Indonesian Mahakam Delta with the combination of shelf carbonates and high constructive elongate deltas. Deposition in the Palo Duro Basin was continuous from late Pennsylvanian time to early Permian time.

Hydrocarbon Potential

The necessary controls for oil and gas accumulation are: a source for the generation of hydrocarbons, porous and permeable rocks acting as a reservoir, and a seal to prevent further migration of the hydrocarbons.

No commercial oil production has been established in the Palo Duro Basin. There are oil fields along the flanks of the basin: on the Matador Arch and Amarillo uplift, and in the Dalhart and Hardeman basins. Some oil and gas shows have been reported in wells drilled in the Ellenburger, Mississippian and the Pennsylvanian rocks of the basin. The basin has abundant traps, yet there is debate whether the Palo Duro Basin had a geothermal gradient high enough for the generation of hydrocarbons. The geothermal gradient in the Palo Duro Basin is 1.1°F/100 ft. This is lower than the worldwide mean of 2.0°F/100 ft. Given such a low gradient, the conclusion is that potential source rocks needed to reach 10,400 ft to attain burial temperatures necessary to generate oil (Birsa, 1977). Dutton (1980) reports that the sediments of the Palo Duro Basin may not have been buried deeply enough, but time as well as temperature is a factor in the formation of hydrocarbons. Ruppel (1985) disputes the findings of both Dutton and Birsa, and concludes that formerly the geothermal gradient was higher.
The Bureau of Economic Geology studied the Palo Duro Basin as a U.S. Department of Energy repository site for the disposal of high-level nuclear waste. The organic geochemistry of the basin was also studied, and the basal Pennsylvanian shales indicate that they are fair to good source rocks. The kerogen showed indications of temperatures high enough to generate hydrocarbons, but the basin was not a source of major hydrocarbon generation (Dutton, 1980).

There may be some hydrocarbon reserves to be found in the Palo Duro Basin for the following reasons: the Osage rocks locally meet minimum requirements for source rocks, kerogen rich shales are abundant in the central parts of the basin, and Mississippian rocks have sufficient porosity and permeability to form reservoirs. Potential structural traps appear to exist along the Matador Arch, and carbonate buildups similar to the prolific carbonate build mounds found in the Hardeman Basin may also occur in the eastern part of the Palo Duro Basin (Ruppel, 1985).

Overall the oil prospects of the Palo Duro Basin are small. Ruppel (1985) concludes that finding this oil will require a synthesis of seismic data, geology, and geochemical analyses.
Exposed in the Rolling Plains at the foot of the Caprock Escarpment are red beds of the Upper Permian Whitehorse Sandstone and the Cloud Chief Gypsum of the Whitehorse Group. These units consist of interbedded red shale, siltstone, sandstone, and gypsum. Overlying the Whitehorse Group are the thinly-bedded sandstones and shales of the Permian Quartermaster Formation (Figure 3.1).

**Whitehorse Group**

The Whitehorse Group consists of reddish-brown, fine-grained sandstone and siltstone, with some thin dolomite and gypsum beds. The Whitehorse rests conformably on the underlying Permian Blaine Formation and is unconformably overlain by the Permian Quartermaster Formation. The entire group ranges from about 200 feet to over 600 feet in thickness, but only the upper part of the section is exposed in the Caprock Canyon area. The type area is at Whitehorse Springs in Woods County, Oklahoma; however, the Whitehorse extends northward into Kansas and southward into Texas. Lloyd and Thompson (1929) recognized that the Whitehorse and overlying Permian strata could be traced into north Texas from Oklahoma where they were first described. As mapped in Texas, however, the Whitehorse includes some additional strata above the upper boundary at the type locality in Oklahoma. Similarly, some strata in Texas are mapped with the underlying Blaine Formation that probably are equivalent to part of the Whitehorse in Oklahoma.

In various parts of Kansas, Oklahoma, and Texas, different formations and key marker beds of dolomite or gypsum have been formally named units within the Whitehorse (Fay, 1964). In Texas, the group is mapped undivided in many areas. In the Texas Panhandle region, a prominent bed of dolomite (named the Childress Dolomite) marks the base of the group, and a resistant scarp-forming sandstone (the Dozier Sandstone) occurs near the middle of the section. Several beds of gypsum (up to 20 feet thick) mark the top of the Whitehorse. These are thought to be equivalent to the Cloud Chief Gypsum of Oklahoma. The Cloud Chief is mapped as a separate formation in Oklahoma but it is generally included within the Whitehorse for mapping purposes in Texas.

Dolomitized algal limestone, particularly near the base of the Whitehorse Group, contains fossils of brackish-water marine bivalves and brachiopods. These indicate that the Whitehorse red beds are time equivalent to the Late Guadalupian limestones of the Permian Basin (Capitan reef deposits; Figure 3.2, Figure 3.3).
Figure 3.1 Permian stratigraphy of north central Texas showing the position of the Whitehorse and Quartermaster formations (from Ross & Ross, 1988). On right is the general physiographic setting of Permian exposures in Caprock Canyons State Park, and the types of deformation observed (from Collins et al., 1990).
Quartermaster Formation

The Quartermaster Formation consists of about 300 feet of interbedded reddish-orange siltstone, very-fine grained sandstone, and shale. Cross-cutting veins of satinspar gypsum are common in the base of the formation. There are also several thin, discontinuous dolomite beds. One of these, the Alibates Dolomite, near the top of the formation, is extensive over much of the Panhandle region and is a readily traceable subsurface marker bed. In places near Amarillo, this bed has been silicified and the chert nodules were quarried by the native Americans for manufacturing stone tools. This high-grade "Alibates agate" was traded over much of the Southwest. From about 1100 A.D. to 1450 A.D., a thriving culture was established in the Panhandle region, quarrying and trading this agate.

The base of the Quartermaster appears to be conformable with the underlying Cloud Chief Gypsum and Whitehorse Group. However, the uppermost parts of the Whitehorse may actually be missing from much of north central Texas. Hence, this contact may be unconformable (Figure 3.1; see Ross and Ross, 1988). The top of the Quartermaster is unconformable with the overlying Triassic Tecovas Formation. In places, this unconformity is mildly angular. However, in the Caprock Canyons area an unusual aeolian sandstone intervenes between the top of typical Quartermaster sediments and the base of the Tecovas (see trail log). The age and affinity of this aeolian unit are uncertain. The bulk of the Quartermaster Formation is thought to be latest Permian (Ochoan) in age, and time equivalent to the extensive gypsum and salt beds of the Salado, Rustler, and Castille formations in the Permian Basin (Ward, et al., 1986). However, the rarity of fossils within these non-marine red-beds makes correlation difficult.

Depositional Environments

The red beds of the Whitehorse, Cloud Chief, and Quartermaster formations represent continental and coastal sand flat and brine pan deposits. These sediments accumulated along the northeastern shore of the Permian Sea as it retreated to the southwest during Late Permian time (Figure 3.2, Figure 3.3). This part of North America was hot and extremely arid during late Permian time, and the coastal environments represented by the Whitehorse and Quartermaster must have been very sparsely vegetated, and largely uninhabited by land animals. There are very few fossils in these deposits. This contrasts with conditions in early Permian time, when this region was vegetated and inhabited by a wide array of reptiles and amphibians, fossils of which are common.

The Whitehorse and Quartermaster red beds accumulated along the Permian shoreline in shallow lagoonal water, and
Figure 3.2 Depositional environments and sedimentary facies interpretation for typical Permian red-bed deposits along the northern margin of the Permian Basin (above). Below is a generalized transect across the Permian shelf, shelf-margin, and basin in Guadalupian time (from Ward et al., 1986). Permian strata observed in the field trip area represent coastal flat and playa facies.
Figure 3.3 General facies map for Upper Guadalupian strata in the Permian Basin (from Ward et al., 1986). The field trip area is just off the northern border of the map, and are an extension of the "red sandstone" facies.
inland on windblown tidal flats and in tidal channels. Evaporation of seawater in shallow coastal ponds and playas precipitated beds of laminated gypsum. Periodically, during periods of rising sea level, lagoon waters inundated broad areas of these coastal flats and thin beds of limestone were deposited over the region. The limestone beds were subsequently dolomitized and locally silicified as sea level fell and red bed deposition resumed (e.g. Ward et al., 1986).

Deformation Caused by Salt Dissolution

Deformation of the Permian strata in the Caprock area has been described in detail by Goldstein and Collins (1984). Dissolution of Permian salt deposits is thought to have caused the collapse of strata or subsidence which resulted in the folds, faults, and gypsum-filled veins (see trail log). The Permian strata can be divided into two zones based on the types of deformation observed (Figure 3.1). The upper zone is the massive, relatively undeformed beds in the top of the Quartermaster Formation. It is characterized primarily by joints. The lower zone, composed of the Whitehorse Group and the lower Quartermaster Formation, is characterized by thinly bedded siltstone with gypsum beds. The deformation in this zone consists of normal faults, reverse faults, and gypsum veins which occur along fault zones, bedding planes, and joint surfaces. The position of the boundary between the deformed and the overlying undeformed zone is a function of bed thickness and vertical distance from the salt dissolution. Beds in the lower zone are from 0.1 to 1.0 meter thick. The Upper Quartermaster Formation beds are generally greater than 2.0 meters thick and commonly as much as 10.0 meters thick. The thicker beds are more rigid, therefore less readily deformed.

Undeformed Zone

This zone, which consists of the upper part of the Quartermaster Formation, is characterized by systematic and nonsystematic joints. The systematic joints predominantly strike north, northeast, east, or northwest, with northwest-striking joints being the least common. One or more of the systematic joint sets predominates in different locations in the region (see trail log, Stop 4). Nonsystematic joints are curved and irregularly spaced, and they truncate against systematic joints. The deformation which caused these joints is thought to predate formation of the lower gypsum-filled veins.
Deformed Zone

One of the most striking features of the Permian strata are the gypsum-filled veins that lace the canyon walls. Three varieties of gypsum are found in the canyon: (1) satinspar, fibrous with a silky sheen; (2) selenite, colorless, transparent, and commonly occurring in sheet-like masses; and (3) alabaster, fine-grained and massive. Satin spar is the most common variety and occurs in thin bands interbedded with mudstones, shales, and sandstones.

When a landlocked body of sea water in an arid climate becomes separated from the ocean, one of the most common salts to precipitate is hydrous calcium sulfate, or gypsum. It is thought that the alabaster gypsum beds (type 3 above) of the Whitehorse Group and the lower part of the Quartermaster Formation were deposited in this manner during the late Permian Period. As the water evaporated, gypsum was precipitated. Periodic influxes of silt and mud would account for the interbedded shales and mudstones found with the gypsum.

Some believe that much of the satinspar and selenite gypsum (types 1 and 2 above) was originally anhydrite and became gypsum in the presence of water. Microscopic examination of gypsum samples reveals residual anhydrite crystals embedded in the gypsum. Also, some attribute the localized small scale folds within the gypsum beds were due to the hydration of the anhydrite. As hydration occurred and anhydrite was converted to gypsum, the gypsum exerted both a lateral and vertical pressure, thus responsible for the small anticlines and synclines. This produced the crumpled wave-like features that are characteristic of the gypsum beds of this area. Others have suggested that the deformation is due to gypsum dissolution (Collins, et al., 1986). As the gypsum was dissolved and carried away, the removal of the supporting layers allowed slumping of the overlying shales and mudstones (see trail log, Stop 1; Matthews, 1969).

The satinspar gypsum filling the veins and faults gives the best clue to the directions of the forces involved in the deformation. The veins are either vertical, parallel to bedding, or cutting the bedding at thirty to sixty degrees. The fibrous gypsum crystals grew as the joints opened. Hence, the fiber direction will indicate the direction of principle extension exerted during mineralization. The gypsum fibers in the vertical veins are horizontal, indicating that they were formed during horizontal extension. The horizontal and inclined veins contain vertical fibers indicating mineralization occurred during vertical extension.
STRATIGRAPHIC CROSS SECTION OF FIELD TRIP AREA, showing Upper Permian, Triassic, Tertiary, and Quaternary strata (from Gustavson, 1986).
The Dockum Group is a sequence of Triassic non-marine "red bed" sediments which outcrop around the Southern High Plains and through the Canadian River Valley in the Texas Panhandle (Figure 4.1). Outcrops of the Dockum Group are held up by their proximity to the resistant "Caprock" caliche which marks the top of the overlying Ogallala Group.

History of Lithostratiqraphic Nomenclature

Efforts to describe mappable lithologic units within the presumed Triassic section began during a series of four annual geological surveys of Texas. On the first such survey, Cummins (1890) recognized a peculiar section of red shales and micaceous, conglomeratic sandstones near the now-abandoned town of Dockum in Dickens County, Texas. The Dockum Beds, as Cummins called them, unconformably overlay what he thought were Permian rocks and were lithologically distinct from them. The age of the Dockum Beds was not determined until vertebrate remains of unequivocal Triassic age were found by Cope (1893) as described elsewhere in this guidebook.

Drake (1892) divided the Dockum Beds, which he instead called the "Triassic Formation", into three units. The thickness of each of these varied around the belt of Dockum outcrops. The Lower and Upper Beds are characterized by sandy shales, while the intervening Central Beds are composed of sandstones, conglomerates and some sandy shales.

Based on work conducted in the Canadian River valley, Cummins' Dockum Beds were recognized by Gould (1907) as the Dockum Group and divided into two formations. The lower formation was named the Tecovas Formation and subdivided into two informal members. The lower member consisted of white, lavender, maroon, and yellow shales. The upper member of the Tecovas is more consistently magenta in color. Neither of the members is present in every outcrop of the Tecovas. Locally sandy facies of both of the members also complicates their lateral correlation. The upper formation of the Dockum Group was named the Trujillo Formation. It is composed of sandstones and conglomerates which are interbedded with shales.

Although Gould (1907) did not directly relate his classification scheme to that of Drake (1892), May (1988) concluded that the Tecovas Formation was equivalent to the Lower Beds of Drake (1892). May (1988) also showed that the Central and Upper beds of Drake (1892) were comparable to the basal sandstone and upper shaley parts, respectively, of the Trujillo Formation.

In an attempt to extend the stratigraphic classification farther south, Hoots (1925) divided the Triassic section in Borden, Scurry, Howard and Mitchell
Figure 4.1 Map of the Triassic outcrop belt in western Texas and northeastern New Mexico (stippled). Dashed line denotes the lower boundary of the Triassic section in the subsurface. Arrow indicates the field trip area.
<table>
<thead>
<tr>
<th>AMERICAN TRIAS</th>
<th>MARCOU, 1856, 1858</th>
<th>DRAKE, 1892</th>
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<th>HOOTS, 1925</th>
<th>CHATTERJEE, 1986</th>
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<th>THIS REPORT</th>
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PERMIAN SYSTEM

Figure 4.2. Comparison of lithostratigraphic nomenclature for Triassic rocks in Texas (modified after May, 1988).
counties of Texas into two unnamed units. The lower unit included red shales and numerous sandstones. It was equated with Drake’s Lower and Central Beds. Hoots’ upper unit was composed almost entirely of red shales and regarded as the equivalent of the Upper Beds of Drake (1892).

Chatterjee (1986) suggested that the status of the Dockum Group be reduced to that of a formation and that the units within it be reduced to members. The Tecovas Member was defined as the equivalent of Gould’s (1907) Tecovas Formation. The Trujillo Member was equated to the Central Beds of Drake (1892). He also designated a stratotype section of what he named the Cooper Member. The Cooper Member was equated to the Upper Beds of Drake (1892), or the shaley portion of Gould’s (1907) Trujillo Formation above the thick, basal sandstone. This unit was identified as the Chinle Formation of eastern New Mexico by Adams (1929). Frelier (1987) noted that the stratotype section of the Cooper Member did not include the uppermost 100 meters of the Dockum Group. The Trujillo Formation was limited to the thick, basal sandstone of Gould (1907) and the Cooper Member of the Trujillo Formation was raised to formation status by Hunt and Lucas (1990). Recently, Lehman et al. (1992) have redescribed the Cooper Member as the Cooper Canyon Formation. Gould’s (1907) definition of the Tecovas Formation and the definitions of the Trujillo and Cooper Canyon formations used by Hunt and Lucas (1990) will be used hereafter in this guidebook when referring to Triassic outcrops in Texas. Changes in lithostratigraphic nomenclature of the Triassic in Texas are summarized in Figure 4.2.

Darton (1922) first used the name "Santa Rosa Sandstone" to refer to the sequence of sandstones which directly overlie Permian strata in the Pecos River Valley of New Mexico. Darton (1928) later included the Santa Rosa in the Dockum Group, the name of which was adopted from Gould (1907). Gorman and Robeck (1946) first subdivided the Santa Rosa Sandstone and mapped unnamed lower and middle sandstone members, a shale member and an uppermost sandstone member in northern Guadalupe County, New Mexico. The type section of the Santa Rosa Sandstone was designated by Kelley (1972a) who identified only two units within the Santa Rosa. The lower unit was equivalent to that of Gorman and Robeck (1946) while the upper member lay unconformably upon the lower one and was comparable to the upper three of Gorman and Robeck (1946). Lucas, et al. (1985) suggested that the Santa Rosa be referred to as a formation, instead of using "Sandstone" as a lithologic name, because the Santa Rosa includes a substantial amount of shale. In his study of the subsurface geology of northeastern New Mexico, Broadhead (1984) could not distinguish between the lower and middle sandstone members of the Santa Rosa Formation; otherwise, he used the subdivisions of Gorman and Robeck (1946).

The stratigraphic nomenclature of the Santa Rosa Formation was again revised by Lucas and Hunt (1987). Its
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Figure 4.3. Comparison of lithostratigraphic nomenclature for the oldest Triassic sediments of northeastern New Mexico (Modified after Lucas and Hunt, 1987).
lower sandstone member was named the Anton Chico Formation, and later designated as a member of the Moenkopi Formation of Arizona (Lucas and Hunt, 1989). The other informal parts of the Santa Rosa were named, in ascending order, the Tecolotito, Los Esteros, and Tres Lagunas members of the Santa Rosa Formation (Lucas and Hunt, 1987). Their scheme of stratigraphic nomenclature is shown in Figure 4.3.

The balance of the Triassic section which overlies the Santa Rosa Formation in New Mexico was first identified as the Chinle Formation (Adams, 1929) as shown in Figure 4.4. It was divided into three unnamed parts (Northrop et al., 1946; Griggs and Hendrickson, 1951). The lowermost and uppermost members are shale and are separated by a medial sandstone interval, which was mapped later by Wanek (1962). The three members were formally designated and mapped as the lower, Cuervo Sandstone, and the upper members of the Chinle Formation by Kelley (1972a, 1972b). Broadhead (1984) also recognized the tripartite division of the Chinle Formation proposed by Kelley (1972b).

The uppermost part of the Chinle Formation in parts of Guadalupe, San Miguel, Quay and Harding counties, is an interval of regularly bedded siltstones and sandstones which was designated as the Redonda Member (Dobrovolny et al., 1946). It was raised to formational status by Griggs and Read (1959). The Redonda Formation was mapped as being distinct from the Chinle Formation by Wanek (1962) and Kelley (1972b). Others, such as Bachman and Dane (1962), Dane and Bachman (1965), Kelley (1972a), Dinwiddie and Clebsch (1973), and Barnes (1983), mapped it as part of the Chinle. The status of the Redonda was lowered to that of a member of the Chinle Formation (Lucas et al., 1985), and then raised back to that of a formation (Lucas and Hunt, 1989).

Lucas and Hunt (1989) revised the stratigraphic nomenclature of the Chinle Formation. Its Lower Shale Member was named the Garita Creek Formation. The Cuervo Sandstone Member was identified as the Trujillo Formation, while the Upper Shale Member was called the Bull Canyon Formation. Use of the name "Chinle Formation" in east-central New Mexico was abandoned. In the rest of this study, the nomenclature of Lucas and Hunt (1989) will be used to refer to Triassic rocks which lie above the Santa Rosa Formation in New Mexico.

Correlation of Triassic Rock Units from Texas into New Mexico

Hoots (1925) described sandstones and conglomerates exposed in Eddy County, New Mexico and Loving and Ward counties, Texas. He concluded that they were equivalents of Darton's (1922) Santa Rosa Sandstone. The Dockum "series" was divided by Adams (1929) into two units in the subsurface of Lea County, New Mexico and Midland, Mitchell and Crockett
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Figure 4.4. Comparison of lithostratigraphic nomenclature for Triassic rocks lying above the Santa Rosa Formation in northeastern New Mexico (Modified after Lucas, et al., 1985).
counties, Texas. The lower sandstone was equated to the Santa Rosa Sandstone while the upper shale interval was compared with the Chinle Formation of Arizona. Roth (1932), without substantiation, equated Drake's sand-rich Central Beds to the Santa Rosa sandstone of Darton (1922). Similarly, Sidwell (1945) likened the Santa Rosa Sandstone (of Darton, 1922) in the Pecos River valley to the Trujillo Formation of Gould (1907) in the Canadian River valley and in Palo Duro Canyon on the bases of similar heavy mineral fractions and grain roundness distribution.

Lithostratigraphy of the Dockum Group in Texas

Tecovas Formation

The Tecovas Formation was named by Gould (1907) for Tecovas Creek in western Potter County, Texas. At most localities, the Tecovas Formation unconformably overlies sediments which are known to be Permian in age (Figure 4.5). However, in the field trip area, the Tecovas Formation is separated from the underlying Permian Quartermaster Formation by an aeolian sandstone of equivocal age (Figure 4.6).

Aeolian unit. In outcrops from southern Armstrong County to Northeastern Floyd County, the Permian-Triassic contact appears to be gradational. It is overlain by an interval whose thickness varies from 3 to 18 meters and is composed of brown to orange, very fine to medium-grained sandstone. These deposits have been interpreted as aeolian in origin (May, 1988). The unit is largely characterized by large-scale trough cross-bedding with lesser horizontal bedding and parallel lamination.

The age of this unit is unknown. It is petrographically dissimilar to both the underlying Permian Quartermaster Formation and the balance of the overlying Triassic Tecovas Formation. No fossils have been found in this unit, and it is separated from predominant lithologies of the Tecovas by an intervening paleosol (May, 1988). Rampant speculation suggests that this unit may be equivalent to the Moenkopi Formation of Middle Triassic age in the Colorado Plateau.

Paleosol. South of Tule Canyon (including the Caprock Canyons area) a paleosol rests on top of the aeolian unit. North of Tule Canyon the paleosol immediately overlies known Permian rocks. Where the paleosol is present, its thickness varies between 2 and 11 meters. At most locations the paleosol is purple to purple-gray in color. In some outcrops, it includes bright yellow or mottled red intervals. At three locations north of Caprock Canyons, the paleosol includes a white or maroon, strongly lithified silcrete horizon.

Channel sandstones. At Caprock Canyons and several other locations along the Eastern South Plains Escarpment
Figure 4.5. Measured stratigraphic section of Triassic rocks in Caprock Canyons State Park (Modified after May, 1988).
Figure 4.6. Schematic cross-section of the Triassic Dockum Group along the Eastern Escarpment of the Southern High Plains, Texas (after Lehman, et al., in prep).
the Tecovas is characterized by discontinuous sandbodies. Their thicknesses vary from .5 to 48 meters while widths range from a few meters to several tens of meters. Such sandbodies also occur at higher stratigraphic levels within the Tecovas Formation. They may be represented by single lenticular bodies or composites of several depositional units. These sandbodies in the Tecovas Formation may be either flat-bottomed or concave-upward.

Multi-colored mudstones. Although this unit is not present in southern Briscoe County (including Caprock Canyons) or in Floyd County because of depositional thinning over the Matador Arch, the multicolored mudstones are an easily recognizable feature of the Tecovas Formation in Palo Duro Canyon and Crosby and Dickens counties. Located above the paleosol and the basal sandstones this interval of mudstones, where present, varies in thickness from 5 to 32 meters. Called the "variegated shales" by Gould (1907), they are characterized by a variety of colors—maroon, red, red-brown, and yellow. There may be up to eight colored zones found at one location. As in other mudstone units of the Dockum Group, carbonate nodules are very common. Thin, highly quartzose sandstone intervals are also scattered throughout this unit. They are often underlain by conglomerates primarily composed of carbonate nodule fragments.

Upper mudstone. This rock unit is typically red or maroon in color and is present at almost every outcrop of the Tecovas Formation on the eastern High Plains Escarpment. Immediately below the overlying Trujillo Formation, there are locally well developed paleosols in this unit.

Trujillo Formation

This formation also was first described by Gould (1907) along Trujillo Creek in western Oldham County, Texas. It unconformably overlies the Tecovas Formation. The contact of the Trujillo with the overlying Cooper Canyon Formation is a gradational one. From Palo Duro Canyon to Caprock Canyons the Tecovas-Trujillo contact has substantial erosional relief in some locations and is overlain by a 10 to 90 meter-thick, massive, composite sandstone interval. Northwest of Caprock Canyons in Tule Canyon, the entire thickness of the Tecovas Formation was eroded away, leaving a channel-fill sequence of Trujillo sandstones resting on Permian rocks. South of Caprock Canyons the sandstone unit is less continuous.

The Trujillo Formation is composed of greenish gray sandstones and conglomerates with some intervening red, maroon, or gray mudstones. The typical Trujillo sandstone lithology is characterized by very fine to fine-grained quartz and substantial percentages of rock fragments and mica. Conglomerates are primarily composed of intrabasinal clasts such as carbonate rock fragments, mudclasts, and reworked sandstone and siltstone fragments. In a few
locations, conglomerates are dominated by extrabasinal clasts such as quartzite, chert, and vein quartz fragments.

Cooper Canyon Formation

The Cooper Canyon Formation overlies the Trujillo Formation and has a gradational contact with it. It is dominated by red mudstones, although it includes discontinuous, lenticular bodies of sandstone up to 60 meters thick which are petrographically identical to those of the underlying Trujillo Formation. Although the Cooper Canyon Formation is as much as 200 meters thick in Mitchell and Borden counties, it thins northward to less than 50 meters. In the Caprock Canyons area, less than 20 meters remains following post-Triassic erosion.

Sandstone Petrography

Most of the sandstones in the three formations of the Dockum Group are primarily composed of quartz. Types of quartz found in Dockum sandstones include both monocrystalline and polycrystalline texture. Grains with both straight and undulose extinction are found.

Feldspars comprise up to 10 percent of detrital grains. Feldspar content in sandstones of the Trujillo Formation increases toward the south. Orthoclase is by far the most common of the feldspars.

Metamorphic rock fragments dominate the rock fragment fraction in most samples. They are most often composed of quartz associated with strongly foliated muscovite or biotite. Sedimentary rock fragments consist of carbonate nodule fragments, mudstone, chert, and reworked sandstone clasts. Non-essential constituents in sandstones of the Dockum Group include detrital muscovite (up to 6 percent), biotite, chlorite, tourmaline, zircon, and hematite.

The compositions of sandstones taken from Dockum outcrops between Palo Duro Canyon in Randall County and Crosby County to the southeast are shown in Figure 4.7. Sandstones from the Tecovas Formation contain more quartz than sandstones from the Trujillo and Cooper Canyon formations. Chert is most common in the sedimentary rock fragment fraction of Tecovas sandstones; whereas, the same fraction of most Trujillo sandstones is dominated by metamorphic rock fragments. A third domain of compositions in the primary ternary diagram in Figure 4.7 are largely composed of carbonate rock fragments. This group includes granular sandstones and pebble conglomerates from all three formations of the Dockum Group. Such rocks are generally found at the bases of sandy depositional units and represent the coarsest part of the stream load. The clasts in these samples are much coarser yet more rounded than the silliciclastic fragments with which they are associated. The carbonate grains are thought to represent pedogenic nodules
Figure 4.7. Ternary diagram showing compositions of sandstones collected from the Tecovas and Trujillo formations along the eastern escarpment of the Southern High Plains, Texas (Modified after May, 1988).

- Tecovas Formation (Group A)
- Trujillo Formation (Group B)
- Calclithic conglomerate (Group C)

1. Quartz arenite
2. Subarkose
3. Sublitharenite
4. Arkose
5. Lithic arkose
6. Feldspathic litharenite
7. Litharenite
eroded from overbank facies and deposited in associated channel facies.

Depositional Systems

Before the development of modern sedimentology, in which the characteristics of ancient rock units are related to modern depositional processes; several workers speculated on the nature of environments in which sediments of the Dockum Group were deposited. Cummins (1891) concluded that the Dockum Beds had been deposited in a "shallow, freshwater sea" as did Drake (1892) because of the preponderance of vertebrate remains and the presence of freshwater clam shells. The northwesterly dip of cross-bedding in the Dockum led Drake (1892) to conclude that the agents which transported Dockum sediments flowed from the southeast. He also believed that the finer-grained facies had been deposited in deeper water. Adams (1929) suggested that Dockum sediments had come from Paleozoic sources which outcropped to the south and east, and that the sediments were deposited on alluvial fans and flood plains. Green (1954) indicated only a "continental" origin of Dockum sediments, but also noticed the northwesterly trend of cross bedding in Garza, Crosby, and Briscoe counties. Nevertheless, he concluded that the Dockum Group in that area was derived from igneous and metamorphic rocks and Permian sedimentary rocks exposed to the northeast, east and southeast. Without detailed descriptions of facies, Kiatta (1960) suggested that the Tecovas Formation was composed of flood-plain deposits, while the Trujillo Formation consisted of the channel deposits of streams which flowed from the Central Mineral Region of Texas. As a result of the study of grain size trends and paleocurrent data, Cramer (1973) concluded that conglomerate and sandstones of the Dockum Group were deposited by braided streams flowing from a source area which lay to the southeast.

In a regional study of gamma-ray logs recorded in wells from the Southern High Plains, McGowen et al. (1979) concluded that the Triassic rocks of Texas and northeastern New Mexico were deposited in a large lacustrine basin by streams transporting sediment centripetally. Sediment sources included the Amarillo Uplift, Wichita Mountains, Ouachita Tectonic Belt, Glass Mountains, Diablo Platform, Sacramento Uplift, Sierra Grande Arch, and Sangre de Cristo Uplift (McGowen et al., 1979). Alternation of wet and dry climatic conditions were responsible for deposition of two different depositional facies groups. Relatively wet periods were characterized by deposition of widespread lacustrine mudstones which were overlain by progradational, elongate deltaic and perennial fluvial facies. Dry conditions were reflected by deposition in smaller lakes, small fan deltas, interdeltaic mudflats, and headward eroding, ephemeral streams (McGowen et al., 1983). In far
eastern Oldham County, mudstones were interpreted as lacustrine and delta-front deposits, while sandstones were construed as fan-delta plain deposits (McGowen et al., 1979). Johns and Granata (1987) regarded the sandstones in the same area as fluvial or deltaic sediments which emanated from sources to the east-northeast.

Alternatively, May (1988) interpreted the Dockum Group as found from Randall County to Dickens County to have had an entirely fluvial origin. Channel, point bar, valley-fill, abandonment, levee, crevasse splay, sheetflow, and floodplain facies were found. The Tecovas Formation was shown to be composed of deposits of low sinuosity streams and various overbank environments.

In contrast to those of the Tecovas, the Trujillo Formation and sandy intervals of the Cooper Formation were described as deposits of highly sinuous streams. Fluvial systems which deposited the Tecovas Formation flowed to the southeast, while those of the Trujillo and Cooper Canyon formations flowed in a northwesterly direction (May, 1988). Frelier (1987) recognized that sediments of the Dockum Group exposed from Crosby County to Scurry County were also composed solely of fluvial sediments. Channel, point bar, levee, sheetflood, crevasse splay, and distal floodplain deposits were present. The largest sand bodies indicated that the paleoflow direction was toward the north (Frelier, 1987).

Granata (1981) noted close correspondence of lithostratigraphic boundaries to boundaries of depositional units found in northeastern New Mexico. The Anton Chico Formation was interpreted as a upward-fining series of channel-filling and interchannel sheet-flow sandstones deposited on the distal portion of an alluvial fan. The Tecolotito Member of the Santa Rosa Formation was viewed as a fluvial sequence including deposits of channel lag, middle point bar, chute, and upper point bar facies. Near its type section, the Los Esteros Member was regarded as lacustrine and prodeltaic accumulations. The overlying Tres Lagunas Member was interpreted as distributary channel deposits overlain by a transgressive lacustrine delta front-shoreface sequence (Granata, 1981, Figure 8).

The Garita Creek Formation was described as mudstone deposited in widespread lacustrine and prodelta environments (Granata, 1981). The Trujillo Formation (sensu Lucas and Hunt, 1989) was regarded as a series of stacked sequences of lacustrine mudstones, delta front deposits, and distributary channel deposits of which the progradational direction was east-southeast (Granata, 1981, Figure 45). The series is capped by an interval interpreted as fluvial deposits (Granata, 1981, Figure 13). In Guadalupe and San Miguel counties, the same interval has been described, alternatively, as deposits of a northward flowing braided river system (DeLuca and Eriksson, 1989). The Bull Canyon Formation was also construed as having been deposited in a lacustrine environment (Granata, 1981). The Redonda
Formation was viewed as a sequence of cyclic, transgressive and progradational couplets of lacustrine and shoreface facies (Granata, 1981, Figure 15).

**Sedimentary Facies**

**Aeolian Facies**

The lowermost sandy unit of the Tecovas Formation in the Caprock Canyons area may have had an aeolian origin. This interval is dominated by large-scale trough cross-bedding in sets between 50 cm and 110 cm thick (Figure 4.8). Less common are horizontal parallel lamination and inclined parallel lamination. This structural assemblage has been interpreted as a dune sub-facies. The large trough cross-beds represent the preserved foresets of migrating crescentic dunes. Parallel-laminated sands represent wind ripple deposition in sand sheets which surrounded the dunes and were transitional between dune and interdune sub-facies. The interdune sub-facies is represented in this unit by thin siltstones and mudstones which resulted from grainfall between dunes and layers of clay clasts ballistically eroded to granule and pebble sizes.

**Fluvial Channel Facies**

The lowermost parts (up to 1 meter) of many sandbodies in the Dockum Group represent the load carried in the thalweg of streams and are characterized by conglomerates and relatively coarse sandstones (Figure 4.9). Clasts comprising the channel lag commonly include intrabasinal rock fragments (pedogenic carbonate nodules, clay clasts, and reworked sedimentary rock fragments), unionid clam shells, fossil wood, and vertebrate bone.

The major portion of the channel sandstone bodies are characterized by one or more sequences of sedimentary structures and fining-upward grain size patterns which reflect generally upwardly decreasing stream power. The most common sedimentary structures found in the channel sandstones include closely associated upper flow regime plane beds, trough cross-beds, and rare tabular cross-bedding. Composite sandbodies are characterized by scour surfaces which reflect laterally shifting streams and erosion of parts of underlying depositional units. Thin clay drapes which lie at the top of some depositional units reflect deposition from suspension after a dramatic decrease in stream power. A feature present in some channel sandbodies is epsilon cross-bedding (Figure 4.10). The lateral accretion surfaces which delimit such intervals dip at low angles and in a direction approximately normal to the direction of stream flow. An epsilon cross-bed unit represents an event of rapid lateral accretion which occurs on a point bar during a flood event. The greatest known
Figure 4.8. Example of the dune (D) and interdune (I) facies in aeolian deposits of the Dockum Group in Caprock Canyons State Park. Note the large-scale trough cross-beds and parallel lamination of the dune facies. The interdune facies contains layers of laminated siltstone and clay clasts (after May, 1988).

Figure 4.9. Example of the channel facies showing, at the base, the coarsest part of the channel load which is moved in the thalweg (Th). Note the fining-upward grain-size sequence in the channel (Ch) facies. Deposits of the floodplain facies (Fp) are shown above those of the channel (After May, 1988).
Figure 4.10. Cross-sectional outcrop drawing of a channel deposit in the Dockum Group which exhibits epsilon cross-beds (lateral accretion surfaces). Note the dip of the surfaces and the thickness of the sequence (After May, 1988).

Figure 4.11. Example of deposits in an abandoned channel (Ab). Note that grain size is fining-upward, and that the abandonment sequence is both underlain and overlain by sediment deposited in active channels (Ch) (After May, 1988).
vertical thickness of such epsilon cross-bed sets is 12 meters which represents the minimum bank-full depth of the stream which deposited it.

Within the channel sandbodies or at their tops may be occasional intervals which reflect deposition in abandoned channels (Figure 4.11). Abandonment may result from cutoff of chutes or meander necks, or channel avulsion (Collinson, 1986). In the Dockum Group, the abandoned channel facies are characterized by sequences of laminated siltstones and mudstones a few meters in thickness which may be interbedded with thin sandstone. Common sedimentary structures in this facies are parallel lamination, ripple cross-lamination, and climbing ripple cross-lamination.

Overbank Facies

Levee deposits result by deposition of suspended load when, during flood events, a stream flows out of its banks. Levee deposits typically slope away from the channel onto the floodplain and display an upward and outward fining textural progression (Figure 4.12). Sedimentary structures often found in levee deposits are ripple cross-lamination, climbing ripple cross-lamination, and parallel lamination.

The natural levee of a stream may be breached during flood events. Sediment which is deposited in or on the floodplain comprises a crevasse splay (Figure 4.13). Preserved crevasse splays in the Dockum Group are usually thin (thickness between 1 and 2 meters) and found within floodplain sequences. They are characterized by small-scale trough cross-beds, massive bedding, ripple cross-lamination, and parallel lamination. Plant debris is common in crevasse splay deposits.

Sheet flow facies characterizes wide areas flanking the stream channel when flood discharge cannot be confined within the channel. In the Dockum Group the sheet flow facies is represented by sandy intervals up to 25 centimeters thick which are composed of fine to coarse sand and a significant proportion of calcareous nodule fragments. Sedimentary structures found in sheet flow deposits include massive bedding, upper flow regime plane beds, and small-scale trough cross-beds. Lower surfaces of these deposits may reflect paleotopographic variations at the time of deposition. Upper surfaces may dip several degrees.

The floodplain facies occurs in the distal portion of overbank environment and is represented in the Dockum by vertical accretion of clay, silt and sand from suspension. Such deposits are marked by parallel lamination, climbing ripple cross-lamination, and small-scale trough cross-beds (Figure 4.13).
Figure 4.12. Example of natural levee deposits (L). Note the inclined lamination, fining upward grain sizes, alternating sand and mud layers, and that the whole levee sequence is both overlain and underlain by floodplain deposits (Fp) (after May, 1988).

Figure 4.13. Example of crevasse splay (CSp) and floodplain (Fp) deposits. Note upward-fining grain size, diversity of sedimentary structures and abundance of plant material. Also note that the crevasse splay sequence is both underlain and overlain by floodplain deposits (Fp) (After May, 1988).
Paleocurrent Analysis

It has been shown that the dip directions of the lee sides of subaqueous dunes are reliable in indicating current flow direction (Miall, 1974; Potter and Pettijohn, 1977). There have been several analyses of paleocurrent indicators in the Dockum Group along the Eastern Southern High Plains Escarpment (Kiatta, 1960; Cazeau, 1960; Cramer, 1973; Frelier, 1987; May 1988) (Figures 4.14 and 4.15). Most of these studies primarily relied upon dip directions of preserved large-scale trough cross-beds. Although the vector means resulting from these studies are different, they indicate a mean paleoflow direction toward the northwest. The vast majority of data points in these studies were taken from sandstones in the Trujillo Formation, indicating that the source area for sediment in the Trujillo Formation was to the southeast. A small number of paleocurrent data points indicate a northerly source area for sediment in the Tecovas Formation (May, 1988) (Figure 4.15). A particular northerly source terrain for sediment in the Tecovas Formation has not been found. Investigation of the paleocurrent indicators in the stratigraphic equivalents of the Tecovas and Trujillo formations in the Canadian and Pecos river valleys confirm the findings of the investigations noted above (Fritz, 1991).

Depositional History

During Middle Triassic time the lower sandstone of the Santa Rosa Formation (exposed in the Pecos River valley was deposited by westward-flowing ephemeral streams (Fritz, 1991). It lies unconformably above Permian strata. A period of erosion and/or non-deposition followed, during which a widespread paleosol formed. This paleosol is exposed in both the valley of the Pecos River and along the Eastern Escarpment of the Southern High Plains (including Caprock Canyons). In Late Triassic time, streams flowing from the northeast deposited the Tecovas Formation and the equivalent middle and upper members of the Santa Rosa formation in the Pecos River valley (May, 1988; Fritz, 1991). Uplift along the margins of the rift zone forming the incipient Gulf of Mexico resulted in the development of a system of northwesterly-flowing streams and the deposition of the Trujillo and Cooper Canyon formations.

Vertebrate Fauna of the Triassic Dockum Group

The Dockum Group has been considered an important source of vertebrate fossils since 1893 when Edward Drinker Cope first recognized the remains of amphibians and reptiles in these Triassic red beds. Exposures of the sediments are good and numerous outcrops are found in the tributary
Figure 4.14. Radial histograms showing results of studies of paleocurrent indicators in the Dockum Group. X=vector mean, n=sample population) (Modified after May, 1988).

Cazeau (1960)
\( \bar{x} = N 40 \ W \)
\( n = 200+ \)

Cramer (1973)
\( \bar{x} = N 29 \ W \)
\( n = 235 \)

Trujillo sandstones
\( \bar{x} = N 15 \ W \)
\( R = 43 \)
\( L = 54\% \)
\( n = 80 \)

Tecovas sandstones
\( \bar{x} = S 33 \ W \)
\( R = 4 \)
\( L = 68\% \)
\( n = 6 \)

Frelier (1987)
\( \bar{x} = N 2 \ E \)
\( n = 158 \)

Total Dockum
\( \bar{x} = N 19 \ W \)
\( R = 41 \)
\( L = 47\% \)
\( n = 86 \)

Figure 4.15. Radial histograms showing distribution of paleocurrent indicators in the Tecovas and Trujillo formations (after May, 1988).
valleys of the Brazos, Colorado, and Canadian rivers. Small "badland" exposures of red mudstone units within the Dockum host numerous vertebrate fossils (Chatterjee, 1985). Today, exposures of the Dockum Group are world-famous collecting sites for a wide variety of Triassic vertebrate fossils.

During Late Triassic time a subtropical hot and arid climate existed in Europe and North America (Laurasia) that was interrupted by a wet monsoonal phase (Simms and Ruffel, 1990). The Late Triassic saw a marked increase in the abundance and diversity of vertebrate fauna (Hunt, 1990). This sudden population explosion is synchronous with, and perhaps related to, the change in climate from dry to humid (Simms and Ruffel, 1990). The Triassic was a time when land animals experienced an evolutionary explosion as well. The earliest representatives of the dinosaurs, mammals, pterosaurs (flying reptiles), and birds are found in Triassic sediments.

The fossil assemblage of animals from the Dockum Group suggests that they were divided into three habitats: 1) aquatic habitats, consisting of rivers, lakes and ponds; 2) lowland habitats, consisting of the margins of rivers, lakes and ponds; and 3) upland habitats, which would have been the divides between streams. The majority of fossils found are of aquatic animals. Because of the rarity of terrestrial fossils found, it is thought that they inhabited a higher terrain than the streams and ponds where most Dockum fossils have been found (Chatterjee, 1985).

The three environments were inhabited by the following:

**Aquatic:** fish (freshwater sharks, lungfish, and coelocanths), metoposaurs (large flat headed, flat bodied amphibians), and phytosaurs (large crocodile-like armored carnivores);

**Lowland:** labyrinthodonts (primitive amphibians), rynchosauras (heavily built herbivorous quadrupeds about six feet in length), protosaurus (small, lizard-like animals that became extinct at the end of the Triassic), trilophosaurs (lightly built herbivores) and lizards;

**Upland:** aetosaurs (squat herbivores heavily endowed with horny armor plates), coelurosaurus (very small, lightly built and agile carnivores), poposaurs (large advanced, carnivorous bipedal reptiles), and ictidosaurus (small reptiles thought to bridge the gap between reptiles and mammals). A sampling of the fauna is shown in Figure 4.16 (Chatterjee, 1985).

The food chain started with the plant eating animals and progressed through larger carnivorous animals. At the top of the food chain was Postosuchus, a large, swift moving, erect, and bipedal carnivore with a voracious appetite. Postosuchus was a thecodontian reptile (a poposaur), a group that was nearing extinction in the Late Triassic. A chart showing the food chain is provided in Figure 4.16 (Chatterjee, 1985).

Recent discoveries of vertebrate fauna in the Dockum Group have been both promising and controversial. The jaw of
Figure 4.16 Reconstructions of some of the Triassic animals found in the Dockum Group (from various sources including Romer, 1966; Colbert, 1969; Chatterjee, 1985, 1991).
an ictidosaur (Pachygenelus milleri), a small carnivorous reptile, was found in the Dockum Group of Texas in 1982 (Chatterjee, 1983). Ictidosauria, though reptilian in most respects, are considered to be precursors of the mammalian condition. The ictidosauria had such mammalian features as a large temporal opening confluent with the orbit, emphasis on certain skull bones and the suppression of others, teeth in an advanced stage of differentiation, and a doubly articulated jaw.

Pachygenelus milleri appears to be similar to other ictidosauria found in South Africa and Argentina, both of which were part of Gondwanaland in Triassic time. This may support the theory that the supercontinent, Pangea, was intact instead of rifted into two unconnected supercontinents, Laurasia and Gondwanaland, at the beginning of Mesozoic time (Chatterjee, 1983).

Even more controversial is the recent discovery by Sankar Chatterjee of a fossil bird skull and fragmentary skeleton from the Dockum Group in Texas that predates the earliest known bird, Archaeopteryx, by 75 million years (Chatterjee, 1991). Chatterjee identifies Protoavis (Figure 4.16) as a bird by its elongated forelimbs, temporal fossae confluent with the eye socket, presence of a furcula (wishbone), a typically avian quadrate, and a plate-like sternum with a small keel (Beardsley, 1986). Protoavis texensis also has reptilian features like a tail, clawed fingers, and four teeth in the forward part of its jaw (Beardsley, 1986). Some paleontologists, however, do not believe that Protoavis was a bird.

Protoavis is controversial for two reasons: first, because it appears at the time of the earliest dinosaurs, it seems to argue against the prevailing theory that dinosaurs were the ancestors of birds; and, second, it suggests that Archaeopteryx is not the direct ancestor of modern birds (Beardsley, 1986).

Also controversial is the recent discovery of a small skull (about the size of a dime!) in the Dockum Group that may be from the earliest known mammal. Lucas and Hunt (1990) named this animal Adelobasileus (meaning "obscure king") cromptoni. The oldest mammal fossils were previously known from western Europe. Adelobasileus is about 10 million years older than these, and suggests that the mammals may have originated in the New World, not the Old.

These and other controversial discoveries ensure that the Dockum Group will be a focus of paleontological research for years to come.
TERTIARY

The Ogallala Group consists of Upper Tertiary (late Miocene-early Pliocene) non-marine strata which outcrop around the High Plains and rest unconformably on Permian, Triassic, and Cretaceous rocks. Locally, the Blanco Formation (late Pliocene) rests in small basins erosionally inset within the top of the Ogallala Group. Together, the Ogallala and Blanco comprise the Tertiary deposits of the High Plains. No early Tertiary (Paleocene-Oligocene) strata are known.

During early Tertiary time the High Plains region experienced severe erosion, and by middle Miocene time a deeply scarred erosional surface with several deep paleo-valleys was present. Deposition of the Ogallala Group began during the late Miocene when eastward-flowing high energy braided streams began to fill the paleo-valleys with fluvial sediments (Figure 5.2). After fluvial deposition had all but ceased, eolian sediments were deposited over most of the fluvial deposits and paleo-uplands (Figure 5.3). A period of landscape stability then ensued, resulting in the formation of the distinctive "Caprock" caliche, which today holds up the escarpment of the High Plains. In places, lake basins formed on the caprock, in which deposits of the Blanco Formation accumulated.

Ogallala Group

Tertiary deposits on the High Plains are represented by the middle Miocene to early Pliocene Ogallala Group (Elias, 1948; Bretz and Horberg, 1949; Brown, 1956; Frye, 1964; Reeves, 1970; Seni, 1980; Gustavson and Winkler, 1988). The Ogallala is one of the most widespread stratigraphic units in North America, extending beneath the High Plains from South Dakota to Texas. The Ogallala or High Plains Aquifer is an economically indispensable source of water for agriculture, ranching, and municipalities throughout this region. The Ogallala Formation was originally named by Darton in 1899 for exposures near Ogallala, Nebraska. However, it was not until the 1930's that similar strata in Texas were identified as part of the Ogallala Formation (Stirton, 1936).

In the first geological surveys of the High Plains in Texas, Cummins (1890, 1891, 1892, 1893) included Tertiary strata in what he called the "Blanco Canyon Beds," named for Blanco (= White River) Canyon just east of Lubbock. Later, the name Blanco Beds or Blanco Formation was restricted to only the uppermost white-weathering lake sediments in Blanco Canyon (Gidley, 1903). The Blanco Formation is now known to rest unconformably on the underlying Ogallala. The strata we now recognize as the Ogallala Group were in various places around the Texas High Plains called the "Clarendon
Figure 5.1 Geologic map of part of the Southern High Plains in eastern New Mexico and western Texas showing distribution of Tertiary and Quaternary strata mentioned in text (from Gustavson, 1990). The locations of several places mentioned in text are shown, as is the field trip area (black arrow).
beds," the "Hemphill beds," or the Panhandle Formation until the 1930's (Figure 5.4; Sellards et al., 1932).

In Crosby, Dickens, and Lubbock counties, along the eastern escarpment of the Llano Estacado, Evans (1949) described the Ogallala Group as consisting of two formations separated by an unconformity. The basal Couch Formation is a pinkish-gray, well sorted sand with local gravels, and the overlying Bridwell Formation is reddish-brown, unconsolidated sand and clay with local thick channel gravels capped by the resistant cliff-forming "Caprock" caliche. In the Canadian River Valley, north of Amarillo, the Ogallala Group is subdivided into a basal series of gravels called the Potter Formation, and a sequence of unconsolidated sand and thin limestone beds called the Coetas Formation (Patton, 1923). Neither of these subdivisions of the Ogallala Group can be applied in the Caprock Canyons area. Exposures of the Ogallala here as well as in most other areas in Texas are mapped undivided (Figure 5.5). Where water wells are drilled on the Southern High Plains, separate formations are not recognized and the Ogallala is described as a coarse to fine brownish sand, with numerous clays, gravels and caliche layers (Reeves, 1970). Bretz and Horberg (1949) described the Ogallala's thickness as ranging from a few tens of feet to over 550 feet in paleo-valleys.

The top of the Ogallala Group is marked by a massive, extremely hard, siliceous caliche of mid-Pliocene age, commonly called the "Caprock" of the Southern High Plains (Brown, 1956; Frye and Leonard, 1957; Frye, 1964; Reeves, 1970). Gile (1981) estimated that the Caprock required several hundred-thousand years of stable climatic conditions to form. Reeves (1970) noted that its thickness ranges from inches in the north to several tens of feet in the southern part of Llano Estacado, and Brown (1956, p. 4) indicated a range in thickness of 3 feet to 150 feet in paleo­depressions.

The source of the Ogallala's sediments on the Southern High Plains is thought to be the Southern Rocky Mountains in New Mexico. The Rocky Mountains in central New Mexico were uplifted in Miocene time during the opening of the Rio Grande rift zone (Chapin and Seager, 1975). This provided a source area for rivers flowing eastward depositing sediments of the Ogallala Group (Figure 5.2). Sellards et al. (1932), and later Seni (1980), believed that Ogallala rivers coalesced to form a broad alluvial apron or series of alluvial fans. Reeves (1972) pointed out, however, that a significant part of the Ogallala Group, particularly south of Lubbock, consists of wind-blown sand. He believed that these aeolian sediments were blown eastward from the ancestral Pecos River Valley. More recently, Gustavson and Winkler (1980) have shown that aeolian sediments are extensive in the Ogallala Group, particularly in areas between the major paleo-river valleys (Figure 5.3).
Figure 5.2 Topography of the top of the Ogallala Group, showing areas of thick sand deposits in the Ogallala and inferred paleo-flow directions of Ogallala streams (from Gustavson and Finley, 1985). The field trip area (black arrow) represents an inter-stream area.
Figure 5.3 Topography of the pre-Ogallala erosional surface, showing the location of paleovalleys filled primarily with fluvial sediments, and paleo-uplands covered primarily with aeolian sediments (from Gustavson and Winkler, 1990). Ogallala sediments in the field trip area (near location 4 in Briscoe County) are believed to represent aeolian upland deposits.
**Figure 5.4** History of stratigraphic nomenclature applied to strata of the Ogallala Group and overlying Quaternary deposits in western Texas (from Winkler, 1985).
Figure 5.5 Typical stratigraphic sections of the Ogallala Group in eastern New Mexico and western Texas (from Gustavson and Winkler, 1990). Sections 1 and 2 represent primarily fluvial valley-fill deposits, and sections 3 and 4 represent primarily aeolian upland deposits. Section 4 was measured just north of the field trip area.
Blanco Formation

The Blanco "Beds" or Blanco Formation consists of white-weathering deposits that locally overlie the Ogallala Group. The Blanco Formation is not present in the Caprock Canyons area, nor are any equivalent deposits. The Blanco Formation was first recognized by Cummins (1890; 1891; 1892; 1893) who suggested the unit was deposited in a large inland sea. Later, after a detailed study, Gidley (1903) concluded that the unit was fluvial in origin and was deposited in a broad valley. Gidley also inferred a Late Pliocene age for the unit based on available vertebrate fossil evidence. Matthews (1924) agreed with Gidley (1903) and inferred an intermittent (ephemeral) stream environment. Evans and Meade (1945) provide the first qualitative description of the unit at its type locality on Mount Blanco at the juncture of Crawfish Draw and Blanco Canyon in Crosby County, Texas. They describe the unit as consisting of well-bedded, light gray, calcareous sands and clays with some freshwater limestones, tufa, diatomite, and coarser gravels (Figure 5.6). The Blanco Formation reaches a maximum thickness of approximately 18 meters (60 feet). The sediments grade upward and laterally to coarser gravels that consist of locally reworked caliche. Evans and Meade (1945) postulated that the Blanco was lacustrine in origin. This interpretation is accepted today. Based on vertebrate fossil evidence they inferred a Pleistocene (Nebraskan) age for the unit. The Rita Blanca beds in Hartley County and the Cita Canyon beds in Randall County are considered to be equivalent to the Blanco.

Frye and Leonard (1957) inferred an alluvial origin, involving sedimentation in a semi-arid terrain; however, they agreed with Evans and Meade (1945) on the unit’s age. Izett and others (1972) studied two ash beds that occur in the type area of the Blanco Formation. They correlated the upper ash (which overlies the Blanco Formation) with the 1.4 Ma-old Guaje Ash. This ash was erupted from the Jemez Cauldera in New Mexico. They were unable, however, to date or correlate the Blanco ash, which is 9 to 10 meters below the top of the Blanco. Pierce (1973), after detailed mineralogical studies of the Blanco and the overlying 1.4 Ma-old Guaje Ash, concluded that the unit was deposited during Late Pliocene and perhaps early Pleistocene (pre-Nebraskan) time under semi-arid to arid climatic conditions. Later Boellstorff (1976) was able to obtain a date of 2.8 +/-0.3 Ma for the lower Blanco Ash, indicating that at least the lower half of the unit, and perhaps all of the Blanco is indeed of Pliocene age.
Figure 5.6 Outcrop drawing showing the top of the Blanco Formation in its type area, and the overlying Blackwater Draw Formation (from Holliday, 1990). Note buried soil horizons in the Blackwater Draw, and position of the Guaje ash bed.

Figure 5.7 Biostratigraphy of the Ogallala Group in Crosby County area - south of the field trip area, showing correlation with the North American Land Mammal Ages (from Winkler, 1990).
Vertebrate Fauna of the Ogallala Group and Blanco Formation

The Ogallala Group of the Texas High Plains is famous for vertebrate fossils of Late Miocene–Early Pliocene time. W. F. Cummins and E. D. Cope described the first fossils found in this formation in 1893 while working for the State Geological Survey of Texas. Among the fossils collected by Cummins and Cope were remains of horses, camels, carnivores, proboscideans (elephants), and large tortoises, all from the Late Tertiary and Pleistocene. Since that time, extensive collections of fossils have been made for, among other places, the American Museum of Natural History, the University of California, West Texas State University, Texas Tech University, the Frick Laboratory, Midwestern State University, Yale University, and Harvard University. Sediments of the Ogallala Group and overlying Quaternary on the High Plains provide one of the best places in the world to study the evolution of land mammals during the later part of the Cenozoic.

Vertebrate fossils, in particular mammals from the Ogallala Group are so diverse and abundant that these faunas were chosen to represent the "type" examples for several of the North American Provincial Ages of the Tertiary as defined by Wood and others (1941) (Figure 5.7). The type sections for the Clarendonian (named for Clarendon, Texas) and Hemphillian (named for Hemphill County, Texas) Land Mammal Ages are in the Ogallala Group north of Caprock Canyons. The type fauna for the Blancan Land Mammal Age (named for Blanco Canyon, Texas) is from the Blanco Formation south of Caprock Canyon. The Provincial Ages were later redesignated as North American Land Mammal Ages (Evernden et al., 1964, and Savage, 1962). These faunal ages allow for correlation of Tertiary non-marine sediments throughout North America.

Climate during the Clarendonian Age (Late Miocene) was mild and subhumid. The High Plains region was dominated by savanna conditions with quiet ponds, lakes, and marshy areas on grass-covered floodplains, and heavily wooded areas that bordering streams. Channel sands and silts were deposited in floodplains, small lakes and ponds, and today these are locally fossiliferous. Sinkholes formed by dissolution of underlying evaporite minerals in Permian red beds often trapped unsuspecting animals and led to their untimely, but intact, demise. Their completely articulated skeletons can sometimes be found today, but removal of the skeletons from the concretionary sandstone matrix has proven difficult. The yellow to brown Ogallala sandstone deposits in the sinkholes were more resistant than the surrounding Permian red beds, which have since eroded away, and in some locations the sinkholes can be seen today as high, flat-topped hills capped by fresh water limestone.

The Clarendonian fauna is dominated by medium to large grazing mammals, most notably (Figure 5.8): perissodactyls, such as pliohippine and hipparion (3-toed) horses,
rhinoceroses (Teleoceras, short, hippo-like amphibious rhinos); artiodactyls (the small camels Procamelus and giraffe-camels Aepycamelus); small, horned ruminants (dromomerycids and moschids, browsers that are possibly related to deer and giraffe, but are, in fact, difficult to assign to either group); and protoceratids, (which include Synthetocerus tricornatus, a three horned so-called "slingshot" deer), proboscideans (mastodons), and dogs (Ischyrocyon, bone-crushing bear-dogs; Cynarctus, racoon-like dogs; Aelurodon, hyena-like dogs; and Epicyon, wolf-like predators). Lower vertebrates included land tortoises and aquatic turtles, alligators and gar fish.

Much of the Clarendonian fauna became extinct about 6 Ma ago, about the time that marks the division between early and late Hemphillian faunas (Late Miocene-Early Pliocene). The climate at that time was beginning a progressive trend toward aridity and the savanna was slowly being replaced by grassland prairie and steppe. The Hemphillian contained many of the advanced species of the Clarendonian Age, but had fewer browsing herbivores and a lower diversity of grazers, probably due to the changing climate. It also marked the appearance of some immigrant fauna, primarily carnivores and sloths. The Hemphillian faunas were dominated by rhinoceroses (Aphelops), gomphotheres, also known as trilophodons Amebelodon, a long-jawed, two-tusked mastodont), and large carnivores (Epicyon, dogs; Nimravides, long-legged scavenging cats; and Barbourofelis, saber-toothed cat-like carnivores). At some localities, the fauna is dominated by the gomphothere elephants and carnivores. Kitts (1957) suggested that the carnivores were either preying upon or scavenging the carcasses of the mastodons.

The Blancan fauna is also dominated by grazing herbivores such as camels, horses, and gomphotheres. There are also browsers such as deer, ground sloths, and peccary, as well as giant armadillos (glyptodonts). The Blanco Formation and other deposits with Blancan fauna formed in shallow lakes surrounded by grassy plains, under a semi-arid climate comparable to that found today on the High Plains.

Most of the above information on Tertiary mammal faunas is summarized from chapters written by Gerald E. Schultz in two excellent guidebooks previously published on the Ogallala Group (Schultz, 1990, and Schultz, 1990).

Hydrogeology

The Ogallala aquifer is an extensive unconfined aquifer underlying most the Southern High Plains. About 30 percent of the groundwater used for irrigation in the United States is pumped from the Ogallala, and about 500 billion cubic meters of drainable water is stored there (Weeks and Gutentag, 1984). The saturated sands and gravels of the Ogallala are also in hydraulic continuity with underlying
Figure 5.8 Reconstructions of the Tertiary land mammals found in the Ogallala Group and Blanco Formation (from Colbert, 1969).
Teleoceras
hippo-like rhinoceras

Protolabis
small camel

Ischyrocyon
bone-eating bear dog

Neohipparion
3-toed horse

Machairodus
saber-toothed cat

Aepycamelus
large giraffe camel
sandstone aquifers in the Triassic Dockum Group and Cretaceous limestone aquifers.

Recharge to the Ogallala aquifer is primarily from surface water which collects in thousands of small playa lake basins covering the surface of the High Plains. About 90 percent of the local runoff drains into these playa basins. Surface water also recharges the aquifer at lesser rates by diffuse percolation through the unconsolidated aeolian cover sands and through the floors of ephemeral drainages that cross the High Plains. There remains a good deal of debate, however, regarding the true proportion of annual precipitation that is currently being recharged into the aquifer, and regarding the rates of recharge. Estimates vary by several orders of magnitude (from essentially 0 to over 25 mm/year; Nativ, 1992). Mean annual precipitation on the High Plains ranges from 350 to 500 mm, and annual pan evaporation from 1500 to 2300 mm.

Regional groundwater flow in the aquifer is to the east-southeast. Water discharges from the Ogallala aquifer (and underlying Dockum aquifer) at numerous springs along the Caprock Escarpment. Pumpage of groundwater from wells (today numbering over 70,000) has resulted in declining water levels over the past 40 years, and concern about the future of irrigated farming in this area. However, unusually wet years over the past decade have actually resulted in dramatically rising water tables in many areas. This suggests that the recharge rate may be much higher than generally supposed. The High Plains Underground Water Conservation District is charged with managing this resource, and there will no doubt be a great deal of interest and research here in the future as the effects of leaking underground storage tanks and sanitary landfills become apparent.
Quaternary sediments on the Southern High Plains and on the Rolling Plains east of the Caprock Escarpment consist of aeolian, lacustrine, and fluvial deposits. The complex interfingering of these sediments resulted from oscillating climatic conditions initiated by Pleistocene glacial activity in North America. Aeolian erosion and groundwater dissolution of Permian salt beds during the Quaternary, caused subsidence of the land surface and the formation of numerous lake basins both on top of the High Plains and on the Rolling Plains to the east. Erosional retreat of the Caprock Escarpment, and isolation of the High Plains plateau began in Pleistocene time and continues to the present day (Figure 6.1).

Blackwater Draw Formation

The Blackwater Draw Formation is an extensive blanket of unconsolidated windblown sand and silt, modified by soil formation, that covers much of the upper surface of the Southern High Plains (Figure 6.2). Frye and Leonard (1957) were the first to study these deposits in detail, and proposed the informal name Illinoisan "cover sands" for this unit. The unit was considered Illinoisan in age because it locally overlies the Tule Formation (then thought to be Kansan in age) and is overlain by the Tahoka Formation (thought to be Wisconsin age). Later Frye and Leonard (1964;1965) recognized a strongly developed soil in the upper part of the unit which they considered to be a Sangamon soil. Reeves (1976) proposed the name Blackwater Draw Formation to replace the informal name Illinoisan "cover sands" for several reasons. First, the unit is very extensive (greater than 100,000 square kilometers). Second, the unit's thickness ranges from a "feather's edge" in the southwest to 27 meters in the northeast. Third, the term Sangamon soil should not be applied to the well developed surface soil, because the unit contains multiple buried soil horizons at the type locality which lies three miles north of Lubbock, Texas. Hence, Reeves believed that the unit accumulated episodically during Illinoisan time. Reeves (1976) also believed that the source area for the windblown sediments was the Pecos River Valley to the southwest. This was based on the observed fining of the sediment from a medium sand in the southwest to a loamy silt in the northeast (Figure 6.2). Holliday (1985; 1989) agreed with Reeves (1976) that the unit accumulated episodically, after noting that the Blackwater Draw Formation contains six recognizable well-developed buried soil horizons at the type locality (Figure 6.3). Holliday (1990) described each horizon as a well developed $B_{tca}$ horizon with Stage II to III carbonate
Figure 6.1 Physiographic evolution of the Southern High Plains region from Late Tertiary time (end of Ogallala deposition) through the Quaternary to the present time (from Gustavson and Finley, 1985).
Figure 6.2 Distribution of the Blackwater Draw Formation over the surface of the Southern High Plains, showing the general fining in grainsize to the northeast (from Holliday, 1990). Several localities mentioned in the text are shown: Bw = Blackwater Draw type locality, MB = Mount Blanco type section for the Blanco Formation, Tb = Tule Lake basin. The field trip area is indicated by the large arrow.
accumulation, red to reddish-brown (2.5YR to 5YR hue) in color, and consisting of silty sand. Outcrops of the Blackwater Draw Formation at different localities on the Southern High Plains reveal that the unit also contains the .62 Ma-old Lava Creek B ash (erupted from the Yellowstone area) and the 1.4 Ma-old Guaje ash (erupted from the Jemez Cauldera in New Mexico), and that not all six soils are present at each locality (Holliday, 1985;1989). Based on this new evidence Holliday inferred that the unit was deposited episodically during most of the Quaternary Period. Holliday also suggested that erosion, deposition, stabilization of the landscape, and soil formation (similar to the present climate) occurred prior to deposition of each succeeding aeolian sand sheet. He based this deduction on the varied number of buried soil horizons at different outcrops, and on the absence of any buried A horizons, which are less capable of withstandling aeolian deflation than are Btca horizons.

Holliday (1990) and Vincent (1991) have suggested that the Blackwater Draw Formation be extended to include all sediments on the Southern High Plains that have been deposited since Ogallala time. This would require the reduction of local lacustrine units such as the Blanco, Tule, Double Lakes, and Tahoka to member status in the Blackwater Draw Formation (Figure 6.4).

Tule Formation

The Tule Formation is a large lake deposit that rests unconformably on Triassic strata in the headwaters of Tule Creek, just west of Caprock Canyons (Figure 6.5; see trail log). The Tule Formation is younger than the Blanco Formation, but nowhere is it observed to overly the Blanco on the Southern High Plains. Lake deposits of the Tule Formation are in part time equivalent to aeolian deposits of the Blackwater Draw Formation. Cummins (1891;1993) suggested that the Tule Formation had a lacustrine origin and Pleistocene (Kansan) age. Later, Evans and Meade (1945) agreed with Cummins on the origin and age. They described the unit as consisting of "well bedded, gray, unconsolidated, sands and greenish-bentonitic clays with thin continuous freshwater limestones" and occasional gravel lenses. The unit reaches a maximum thickness of 45 meters (150 feet) at the type locality (9 miles northwest of Silverton, Briscoe County, Texas; see trail log). Frye and Leonard (1965) interpreted the unit as having been deposited in a large lake basin during the cool, wet climatic conditions of Kansan time. This conclusion was based on faunal evidence, and the presence of the underlying Pearlette volcanic ash bed (.62 Ma-old Lava Creek B ash; Izett and Wilcox, 1982). Reeves (1966;1968;1970) stated that the unit was definitely lacustrine in origin based on the presence of dolomite, and the clay minerals sepiolite.
Figure 6.3 Typical sections of the Blackwater Draw Formation, showing the grainsize/texture (SCL=sandy clay loam, L=loam, SIL=silt loam, SiC=silty clay, CL=clay loam, C=clay) and Munsell color values (from Holliday, 1990). The zonation of buried soils is also shown. The Tule Basin section was measured a few miles west of the field trip area.

Figure 6.4 Generalized stratigraphic cross-section of Late Tertiary and Quaternary strata on the Southern High Plains, showing the intertonguing of soil-modified aeolian deposits and lake sediments (from Holliday, 1990).
Figure 6.5 Location of most of the larger Quaternary lake basins on the Southern High Plains. The stippled area indicates the general region of the High Plains underlain by Cretaceous strata (from Reeves, 1970). The field trip area is indicated by the arrow.
and attapulgite, which are indicative of hypersaline (playa) environments.

Double Lakes Formation

The Double Lakes Formation, was first identified by Reeves (1974) in the Double Lakes Basin, Lynn County, Texas (Figure 6.5). Reeves (1976) later described the Double Lakes Formation as early-Wisconsin age lacustrine sediments consisting of "dark olive-gray (5GY 4/1 hue) dense clay with intermittent zones of gypsum" which reach a maximum thickness of 8.5 meters. The Double Lakes Formation has since been tentatively identified in five additional lake basins on the High Plains (Rich, Mound, Tahoka, Cedar, and Brownfield; see Figure 6.5 for localities) in which it unconformably overlies Triassic, Cretaceous, Ogallala, and Blackwater Draw sediments. Because of its distribution in topographic lows on the unconformity overlying Cretaceous rocks and its dark lithology, Reeves (1976) suggested that the sediment was derived from erosion of the underlying black Cretaceous-age Kiamichi shales.

Tahoka Formation

The youngest Pleistocene lacustrine sediments on the Southern High Plains belong to the late-Wisconsin age Tahoka Formation, first identified by Evans and Meade (1945). This unit is described as consisting of black to blueish-gray calcareous, gypsiferous lacustrine clay (predominantly sepiolite) with gray alluvial sand and gravel, dune sand, and minor thin to nodular carbonate lenses (Evans and Meade, 1945; Frye and Leonard, 1957; Reeves, 1976). Reeves (1970) divided the Tahoka Formation into a lower Rich Lake Member and an upper Brownfield Lake Member. The two units are separated by the Vigo Park Dolomite bed. The Tahoka Formation reaches a maximum thickness of 10.5 meters and unconformably overlies Triassic, Cretaceous, Ogallala, Blackwater Draw, and Double Lakes sediments (Figure 6.5).

Lingos Formation

Quaternary sediments on the Texas Rolling Plains adjacent to the Caprock Escarpment are represented by the Lingos Formation (Caran and Baumgardner, 1990). The Lingos Formation is a complex sequence of lacustrine, eolian, and fluvial deposits that covers 9,250 square kilometers east of the Caprock Escarpment (Figure 6.6). Deposited in the last 300,000 years, the Lingos Formation consists of a lower unit of middle Pleistocene alluvial fan and slope wash sands and gravels and locally saprolite. These deposits were laid down as the Caprock Escarpment retreated to the west by
Figure 6.6 Distribution of the Quaternary Lingos Formation and related strata on the Rolling Plains (from Caran and Baumgardner, 1990).
Figure 6.7 Evolution of the Caprock Escarpment in Quaternary time, showing deposition of the lower, middle, and upper parts of the Lingos Formation (Q1) and entrenchment of the modern drainage (from Caran and Baumgardner, 1990).
headward erosion of tributaries of the Red River and Brazos River (Figure 6.7). The middle Lingos Formation is represented by lenses of calcareous, fossiliferous, lacustrine clays deposited 35,000 to 8,000 years ago. The lakes formed in areas subject to groundwater dissolution of Permian salt beds. The upper part of the unit is represented by Recent (8,000 years ago to present) eolian sheet sands and silts, and fluvial terrace deposits of sands and gravels. The Lingos Formation has a maximum thickness of 82 meters and is the result of a complex interaction between the erosional retreat of the Caprock Escarpment and dissolution of underlying Permian "salt" Formations.

Holocene sediments on the High Plains and Rolling Plains

Recent (post-Wisconsin) sediments on the Southern High Plains consist of aeolian dune sands, playa lake clays, fluvial and marsh deposits in ephemeral streams, and basin-associated aeolian lunettes (Evans and Meade, 1945; Frye and Leonard, 1957:1965; Reeves, 1970, 1976; Holliday, 1989, 1990; and Williams, 1992). Holocene alluvial deposits in the ephemeral streams that cross the high Plains, have yielded important stratified archaeological sites that document the progression of human culture in the region beginning about 11,000 years ago (see trail log).

Tributary streams of the Red River and Brazos River continue to incise canyons into the Caprock Escarpment in Holocene time. The floors of these canyons contain as much as 15 meters (50 feet) of Holocene stream terrace deposits and colluvium (see trail log). These sediments accumulated in the stream valleys up until about 1000 to 800 years ago. At that time, perhaps owing to change to an arid climate, the streams began incising down into the older Holocene deposits (Hall, 1990).


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