Geology of the Delaware Basin
Guadalupe, Apache, and Glass Mountains
New Mexico and West Texas

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PART I: SEDIMENTS

STRATIGRAPHY

Over one billion years of rock record is preserved in the Delaware Basin area. Rocks and sediments range in age from 1.3 by for the Precambrian basement to less than 10,000 ybp for Holocene sediments in the Pecos River Valley. A number of these stratigraphic units are source or reservoir rocks for oil and gas and recovery from these units has made the Delaware Basin one of the major hydrocarbon-producing regions in the world. Also among these units is the Capitan Limestone, the classic locality of an ancient reef and visitation site for scores of geologists over the last century.

The overall stratigraphy of the Delaware Basin study area is shown diagrammatically in Fig. 8. Thickness of these stratigraphic units varies considerably throughout the study area and in some localities certain of these formations pinch out, are eroded away, or were never deposited.

The stratigraphy section will be presented in the following order:

(1) From oldest (Precambrian) to youngest (Quaternary).

(2) For each period of geologic time, the Guadalupe Mountains section will be discussed first and the geology of the other regions (e.g., Apache Mountains, Glass Mountains) will be compared to the Guadalupe Mountain section. The reason for this is because the geology of the Guadalupe Mountains has been more thoroughly studied than the geology of other sections.

(3) For each period of geologic time, the discussion of stratigraphy will proceed counterclockwise around the perimeter of the Delaware Basin, from the northwest (Guadalupe Mountains) to the west, south, east, and north; then basinal rocks of the same age will be discussed.

(4) For each period of geologic time where there is a backreef-reef-forereef-basin facies transition, the order of discussion will be from backreef to reef to forereef to basin.

(5) As each period of geologic time is presented, problem topics will be identified and discussed (e.g., for the Ochoan Castile Formation the *Special topic*: Is the Castile Formation a deep-water or shallow-water deposit?).

Names/spelling of all series, groups, formations and members, formal or informal, were taken from the U.S. Geological Survey (1994) database of stratigraphic nomenclature.

Proterozoic

Precambrian

The oldest rocks known from the Delaware Basin are Precambrian in age. The Precambrian basement has been encountered in a number of oil and gas wells in the Delaware Basin area; a contour map of the Precambrian basement as determined from these wells is shown in Fig. 9. Counterclockwise around the basin, the following Precambrian granitic rock occurrences have been encountered in the subsurface: (1) biotite-quartz granite from the Humble Huapache Unit no. 1 well in sec. 35 T23S R22E, Eddy County (Guadalupe Mountains), at a depth of 3847 m, has been dated at about 1330 Ma (Denison and Hetherington, 1969); (2) 25 m of Precambrian granite penetrated by the Humble Antelope Draw no. 1, drilled near the Hovey anticline (Glass Mountain area) (Bloom, 1988); (3) granite and metamorphosed sandstone cut by stringers of pegmatite or aplite in the Shell-Humphreys well, Pecos County, Texas, located along the crest of the Fort Stockton high on the southeastern side of the basin (Jones and Conkling, 1930); and (4) granitic rock in a drill core west of Hobbs, New Mexico (just north of the study area), dated at 1240 Ma (Wasserburg et al., 1962). All of this granitic rock may be part of the "granite-rhyolite terrain" episode of Adams et al. (1993a).

In addition to Precambrian rocks encountered in drill cores, gravity and seismic data indicate that at least part of the Delaware Basin (adjacent to the Central Basin Platform) is underlain by layered mafic rocks, the "Pecos mafic intrusive suite" (Miller et al., 1995). The suite takes the form of a 3-10 km thick sill containing a dike-like structure that may represent the core of a ~100 Ma rift. Outcrops of Precambrian rocks occur just southwest of the study area in the Sierra Diablo-Van Horn, Texas region. Some of the oldest rocks in the Van Horn orogenic belt are part of the Hazel, Allamore, Carrizo Mountain, and Van Horn sequences. The Carrizo Mountain Group rhyolitic ash-flow tuffs have been dated at 1270 Ma (Rudnick, 1983). These rhyolites were extruded over sedimentary rocks of even older Proterozoic age, but the former extent and source of this ancient sedimentary pile is not known. In the Van Horn area a thick mass of ancient Proterozoic sediments was deformed and overridden from the south by a thrust block of Carrizo Mountain schist (King, 1940). Precambrian sediments and volcanic rocks extend considerably to the north of the Van Horn area (Denison and Hetherington, 1969), and it is likely that they also extend into the study area.

Data from both the drill cores and outcrops is scanty, but from this data the story of the earliest beginnings of the Delaware Basin can be tenatively pieced together. Adams et al. (1993) reviewed the geology and tectonics of the middle Proterozoic igneous rocks in the Carlsbad region; bimodality of this rock suggest that continental extension took place in the region between ~1215-1075 my ago and was, in part, coincident with the formation of a Midcontinent rift system. Diabases intruded rhyolitic rocks sometime prior to a regional metamorphic event at about 1200-1000 Ma. This metamorphic event was part of the Grenville Orogeny/

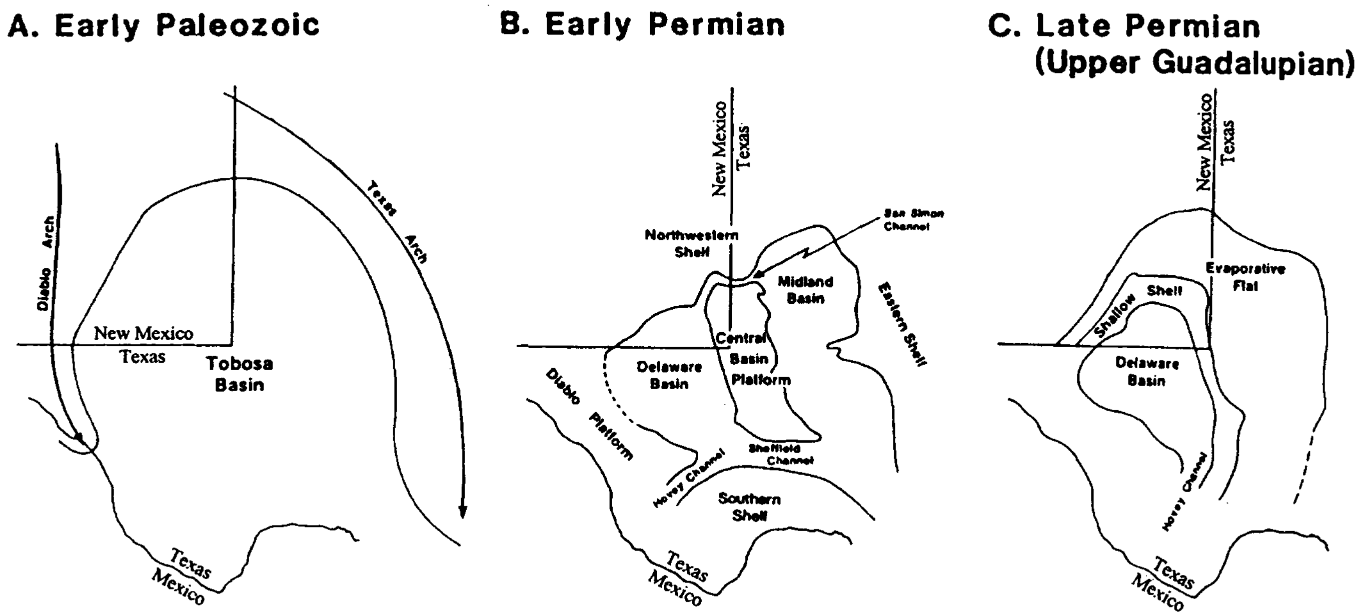

FIGURE 10—Diagram showing the Delaware Basin area in the Paleozoic: (A) the Tobosa Basin in the Ordovician to Pennsylvanian (after Adams, 1965), (B) the Permian Basin in the Early Permian (after Hills, 1984a), and (C) the Delaware Basin in the Late Permian (after Ward et al., 1986, and Garber et al., 1989).

Simpson Group has been encountered in the subsurface in Dodson's Texas American Syndicate no. 1 (515 m thick; Stevens, 1957), and across the Sierra Madera "crypto-explosion structure" (698-721 m thick; Wilshire et al., 1972).

Haigler (1962) and Hayes (1964) described the Simpson as a greenish to gray, brown, and black shale, containing some limestone and coarse- to fine-grained sandstone. Marine ostracods, trilobites, and graptolites are common fossils in this unit (Galley, 1958). Formations (and some sand members) of the Simpson Group are: the Joins, Oil Creek (Connell Sandstone Member), McLish (Waddell Sandstone Member), Tulip Creek, McKee Sandstone, and Bromide (James, 1985; Bureau of Economic Geology, 1989; Fig. 171). All five formations, and two of the sand members (Waddell and Connell), have been recognized from subsurface wells in the area of the Hovey anticline, northwest of the Glass Mountains (Bloom, 1988).

Montoya Group

The Montoya Group of Late Ordovician age unconformably overlies the Simpson Group (Fig. 11A). The Montoya is a thick (140 m or so) sequence of rock composed of light-gray to medium-dark gray, fine- to medium-crystalline, calcareous dolomite, some units of which are interbedded with shale or dark-gray limestone and some units of which contain white to very light-gray porcellaneous chert (Haigler, 1962; Hayes, 1964). The Montoya carbonate limestone-dolomite sequence is dense, impermeable, and non-porous. In the Sierra Madera, Glass Mountains area, the Montoya dolomite varies from 46-155 m in thickness (Wilshire et al., 1972)

Where the Montoya crops out in New Mexico, the formation has been subdivided (from oldest to youngest) into the Cable Canyon Sandstone and into the Cutter Formation, Aleman Formation, and Upham Dolomite, although normally these names are not used in the subsurface (Grant and Foster, 1989). The Montoya is 105 m thick in the Texas Richards well (T20S R32E) and thins to 88 m in the Honolulu Mako well (T14S R28E). Along the Hovey anticline, northwest of the Glass Mountains, the Montoya consists of ~110-140 m of a lower unit of dolomitic limestone containing minor quantities of chert and shale, and an upper unit of dolomite, chert, and lesser amounts of limestone (Bloom, 1988). It also has been encountered in Dodson's Texas American Syndicate no. 1 well in the same area (Stevens, 1957).

Silurian

Fusselman Dolomite

During the Silurian the axes of the Tobosa Basin were generally the sites of relatively deep water where dense limestones and shales were deposited (Fig. 11B). In this ancient basin the carbonate sequence of the Fusselman Dolomite was laid down in response to a highstand in sea level (Canter et al., 1992). The Fusselman was defined by Richardson (1908) who described a 300 m-thick section of massive white limestone in Fusselman Canyon, Franklin Mountains, West Texas. In the Delaware Basin the Fusselman of Lower Silurian age unconformably overlies the Montoya Group of Late Ordovician age with a fairly sharp lithologic break. The formation consists almost entirely of

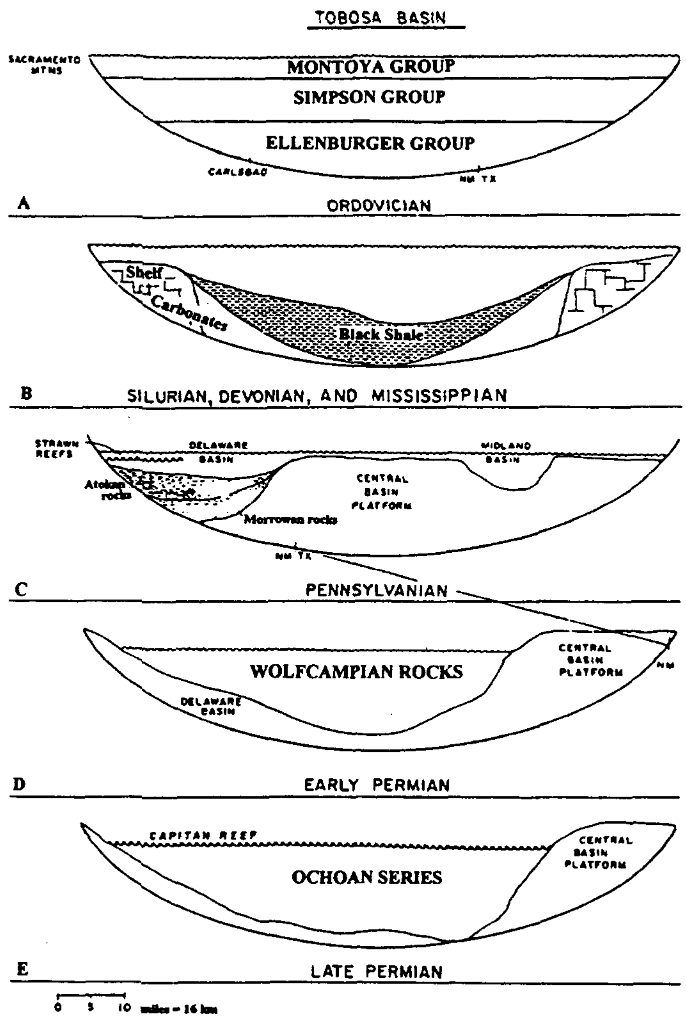

FIGURE 11—Conceptual diagram showing the development of the Tobosa Basin into the Midland Basin, Central Basin Platform, and Delaware Basin from the mid-Ordovician to Late Permian. Cross sections across Lea and Eddy Counties, New Mexico. After Cheeseman (1978).

white to light-gray, coarse- to medium-crystalline dolomite which contrasts rather sharply with the darker-colored and finer-grained Montoya (Hayes, 1964). Fossils found in the Fusselman include *Pentamerus*, *Amplexus*, and *Favosites*. In the southeastern, Sierra Madera-Glass Mountains part of the basin, the Fusselman has been reported to vary between 15-50 m in thickness (Wilshire et al., 1972; Bloom, 1988). In the Hovey area, the Fusselman is 20 m thick in Dodson's Texas American Syndicate no. 1 (Stevens, 1957).

Regional emergence in the Late Silurian through Middle Devonian (~40 my duration) was accompanied by widespread karstification (Mazzullo, 1992). Several distinct karst events that affected Silurian/Devonian strata are: (1) top of the Fusselman, (2) top of the Wristen, (3) top of the Thirtyone, and (4) pre-Woodford. The contact between the Fusselman and overlying Wristen Formation represents a major unconformity and is typically described as an irregular or eroded surface with many karst features such as collapse breccias, cave sequences, and associated cavernous and dissolution porosity (Canter et al., 1992; Fig. 12). The widespread occurrence of this exposure surface suggests that it was related to a platform-wide drop in sea level. Paleokarst is found in the Fusselman throughout the Permian Basin subsurface and contributes substantially to the reservoir porosity of producing fields (Mazzullo and Mazzullo, 1992).

Wristen Formation

A subsurface interval previously referred to as the "Silurian shale," "Upper Silurian," or "Green Silurian" was named the Wristen Formation by Hills and Hoenig (1979), the type section being described from a well in the Wickett field, Ward County, Texas. Canter et al. (1992) reported that the Wristen Formation is thickest (450 m) in the northeastern part of the Delaware Basin, but that it thins northward on the Northwest Shelf and eastward on the Central Basin Platform. The Wristen Formation is primarily a mud- and shale-rich rock dominated by dolomite.

Following a post-Fusselman lowstand, a major transgression occurred during early Wristen time. First there was a rise in sea level followed by a significant sea level fall later in the Wristen, then there was another highstand in which the latest, uppermost intervals of the Wristen were laid down (Canter et al., 1992). The sea level lowstand resulted in exposure and some karstification on a local scale. The contact between the Silurian Wristen and overlying Lower Devonian Thirtyone Formation is unconformable (Fig. 12). This unconformity is indicated by erosional truncation and further significant regional karst development.

Devonian

Thirtyone Formation

At the end of the Silurian and into the early Devonian, shelf carbonates were deposited along the shallower margins of the Tobosa Basin and dense limestones, cherts, and black shales accumulated in deeper waters (Fig. 11B). Carbonate deposits reached over 300 m in thickness during this time (Netherland et al., 1974). The Thirtyone Formation was one of the carbonate deposits laid down in the Tobosa Basin during the Early and Middle Devonian (Fig. 8).The name "Thirtyone Formation" was proposed by Hills and Hoenig (1979) for a unit previously known as the "Lower Devonian cherty limestone" or the "Devonian limestone-siliceous." The type section was described from a well in Blk. 31, Crane County, Texas, where the formation consists of about 300 m of light-colored siliceous and cherty limestone, with an upper interval composed mostly of crinoid-rich limestone and minor sandy limestone. The "Devonian" in Reeves, southwestern Pecos, and northern Brewster Counties has been described as being almost entirely chert and having a thickness of 32-43 m (Bloom, 1988). The "Devonian" is 80 m

The top and base of the Mississippian limestone is usually easily recognizable in well samples and mechanical logs (Grant and Foster, 1989).

The Mississippian in Kalinga Corporation's Margaret no. 1 well, along the Hovey anticline in the Glass Mountains area, is predominantly a black shale with minor amounts of limestone. The shale is gray to black, hard, platy, pyritic, organic and very siliceous. The limestone is brown to dark-brown, microcrystalline to very finely crystalline, generally clean, commonly sandy, and dolomitic (Bloom, 1988). In Dodson Texas American Syndicate no. 1, the unit is 52 m thick (Stevens, 1957).

The Mississippian limestone in Gaines and Andres Counties reaches 335 m thick. In Humble's Antelope Draw no. 1 it consists of over 76 m of crinoidal limestone, but in Kalinga Corporation's Margaret no. 1, 9.6 km southeast of the Antelope Draw well, it is only 15 m thick. Thus, overall, Mississippian limestone thins southward over the study area.

Barnett Shale

Overlying the Mississippian limestone is an upper unit, called the Barnett Shale, consisting of partly silty, brown shale and containing very fine-grained sandstone and siltstone (Hayes, 1964; Wilshire et al., 1972; Grant and Foster, 1989). The Barnett Shale is on the order of 60-140 m, thickening somewhat to the west of the Pecos River. In some places the total thickness of Mississippian strata can reach 250-335 m (Netherland et al., 1974). In Kalinga Corporation's Margaret no. 1 well, the Barnett consists of 60 m of shale with minor units of sandstone and limestone. The shale is gray to dark-gray towards the top, becoming dark-gray or black with depth; the shale is hard, siliceous, subfissile, organic, and very pyritic (Bloom, 1988).

In the area of Carlsbad, the top of the Mississippian below the "last" or lowest Pennsylvanian sand is called the Chester member, an informal unit which is composed of fine-grained sandstone, siltstone, argillaceous limestone, and shale (Andersen, 1976). The Chester interval is usually considered to be of Mississippian age, but may actually be the lowest part of the Pennsylvanian sequence. The Chester represents the first progradation of clastics from the Pedernal landmass and entrance into pre-delta deposition, and it signals the beginning of Pennsylvanian sedimentation in the basin.

The Barnett Shale was the result of tectonic movement at the onset of the Ouachita orogeny (Gardiner, 1990). Sedimentation continued unabated in the Tobosa Basin until the Late Mississippian, but beginning at that time there was a tectonic upheaval that would, in the Pennsylvanian, divide the Tobosa Basin up into the Midland Basin, Central Basin Platform, and Delaware Basin (Fig. 11C). This ended the existence of the Tobasa Basin, but during the 200 my or so from the Ordovician to Late Mississippian, this ancient basin accumulated as much as 2000 m of dolomite, limestone, shale, and sandstone — a continuous sedimentary record interrupted only by relatively brief periods of exposure at the end of Early, Middle, and Late Ordovician time, and again at the end of the Silurian, Early Devonian, and Mississippian periods.

Pennsylvanian

The history of the Delaware Basin as a separate entity began with the rise of an Early Pennsylvanian medial ridge in the Tobosa Basin, brought about by the Marathon-Ouachita orogeny which was the dominant structural event that determined the nature of Pennsylvanian sedimentation (Galley, 1984). Movement along zones of weakness inherited from Precambrian faulting (Fig. 9) induced the emergence of a complex series of fault blocks near the center of the Tobosa Basin (collectively known as the Central Basin Platform), which divided the Tobosa Basin into the Midland Basin on the east and the Delaware Basin on the west (Figs. 10B and 11C). In the southwestern part of the Central Basin Platform (near Fort Stockton, Texas), coarse, arkosic sandstone, varicolored shale, and subordinate carbonates of Pennsylvanian age cover a large area of the shelf and directly overlie the Precambrian basement, the intervening Paleozoic rocks having been eroded away from the uplifted fault block (Mazzullo, 1986). Also during this time tectonic activity increased on the Diablo Platform to the west of the Delaware Basin (Fig. 13), and these rocks were folded, faulted, and deeply eroded prior to the transgression of the Permian sea. The highlands which occupied the Apache Mountain area subsequent to uplift were stripped down to Ordovician rock, as can be seen in Humble's Reynolds Cattle Company "B" no. 1 well in the Apache Mountains where Permian Wolfcampian beds rest unconformably on the Ordovician Bliss Sandstone. In the Sierra Diablo west of the Apache Mountains, Wolfcampian beds rest on Precambrian rocks.

The Delaware Basin subsided rapidly during Pennsylvanian time. Material was eroded from the Pedernal uplift ("Ancestral Rockies"), Diablo Platform, Marathon-Ouchita fold belt (Fig. 13), and other highland areas, and was deposited as thin sequences of sand and shale with interbedded carbonates on and along the edges of the shelves. A deep, starved, shale basin occupied the central and southern parts of the broad Delaware depression throughout the Pennsylvanian, and during this period of geologic time approximately 600 m of sediment accumulated in the basin. Within the area of the Northwest Shelf, intrashelf basins (such as the Tatum Basin; Fig. 18) formed in the Late Pennsylvanian as a result of regional tectonism, and these persisted until the earliest Permian when the basins became filled (Grover, 1993).

Provincial series names for Pennsylvanian rock in the study area are: the Morrowan, Atokan, Des Moinesian, Missourian, and Virgilian. Equivalent formations (groups) are the Morrow, Atoka, Strawn, Canyon, and Cisco, respectively. From a petroleum standpoint these names are somewhat arbitrary, being based on a number of log tops and facies changes which are difficult to correlate from well to well (Grant and Foster, 1989). Meyer (1966) indicated a

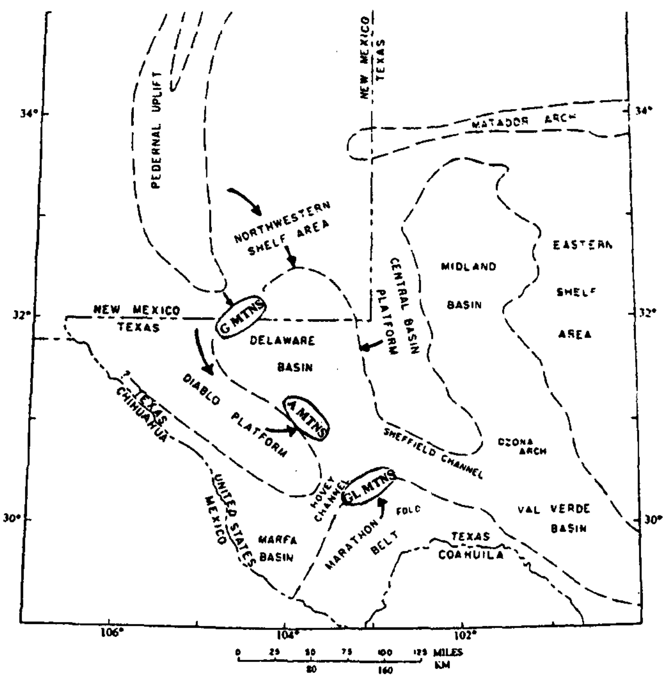

FIGURE 13—Paleogeography of Pennsylvanian time showing approximate location (dashed lines) of land masses and submerged areas. The Pedernal uplift, Diablo Platform, Marathon-Ouachita fold belt, and Central Basin Platform supplied sediment to the Delaware Basin during the Pennsylvanian and also later, in the Permian. After Hayes (1964).

maximum of 900 m of Pennsylvanian rock in the Delaware Basin area, but in an Eddy County reference well, it is only 810 m thick. Pennsylvanian rocks thin to the east and are absent as a result of erosion on fault blocks associated with the uplift of the Central Basin Platform, or as the result of onlap.

The record of Pennsylvanian rock has been recovered mostly from oil and gas wells in subsurface units (Morrow, Atoka, Strawn, Cisco and Canyon), but in the Glass Mountains Pennsylvanian rocks do crop out (the Gaptank Formation). These outcrops make up less than 5% of Pennsylvanian rock strata in the Delaware Basin and cover only the Late Pennsylvanian, but they constitute important field evidence for sedimentary conditions during this time.

Morrowan Series

Morrow Formation

The Morrow Formation of Lower Pennsylvanian age lies unconformably over Mississippian rocks (Fig. 8). A rapid subsidence of the Delaware Basin portion of the old Tobosa Basin began in Early Pennsylvanian time and clastics and carbonates were deposited in this new basin (Fig. 11C). The Morrow varies considerably over its range, consisting of limestones, sandstones, shales, and siltstones (James, 1985). Over much of the area it is a fine- to coarse-grained and conglomeratic sandstone and gray shale with some interbedded limestone and, in the west, it is a red shale and limestone pebble conglomerate. In the Delaware Basin reference well located in Eddy County, the bulk of the Pennsylvanian has been assigned to the Morrow; here, the Morrow interval is 508 m thick and contains brownish limestone interbedded with gray shale in the upper part and gray to brown, coarse-grained to conglomeratic sandstone interbedded with dark gray shale and some brownish, oolitic limestone in the lower part (Grant and Foster, 1989).

James (1985) divided the Morrow Formation into the lower, middle, and upper members (also designated the "A", "B", "C" zones, respectively; Fig. 14). The lower Morrow rests unconformably on the underlying Mississippian Chester member or Barnett Shale. The middle Morrow is marked by a prominent basal shale, called the "Morrow shale" by James (1985, p. 1044) which consists of shales and sandstones. The upper Morrow consists of limestone with interbedded shales and sands, and the top of the Morrow is commonly referred to as the "top of the Morrow clastics."

From the Chester member (last of the Mississippian beds) to the Morrow "C" zone (youngest of the Morrow beds), four pulses of prograding clastics, separated by transgressive (and radioactive) marine shales, record Morrow fluctuations from a marine to primarily deltaic environment (Andersen, 1976; Fig. 15). Sand bodies within the Morrow are primarily of fluvial-deltaic origin; sands represent channels and point bars (James, 1985). The sand bodies are multiple, stacked, and overlap each other. The composite thickness of the lower and middle Morrow intervals can reach as much as 510 m in the deeper parts of the basin and reflects the maximum progradational episode of Pennsylvanian clastic input into the basin from the surrounding Pedernal highlands to the north and west and Central Basin highlands to the east (Fig. 16). Subsequent upper Morrow sediments record a major marine inundation of this clastic wedge prior to Atokan time wherein areas on, or immediately adjacent to, the source areas received little to no Morrow sedimentation.

Atokan Series

Atoka Formation

The Delaware Basin continued to receive an uninterrupted input of sediment in Atoka time. The basal contact of the Atoka Formation appears to be conformable with the below-lying Morrow. The Atoka is a brown, fossiliferous limestone with interbedded shale in its lower portion and primarily shale with occasional limestone in its upper portion (Netherland et al., 1974). The unit also contains light- to dark-gray chert and intervals of poorly-sorted sandstone (commonly conglomeratic) which continue into the eastern-most part of the study area. In the Delaware Basin reference

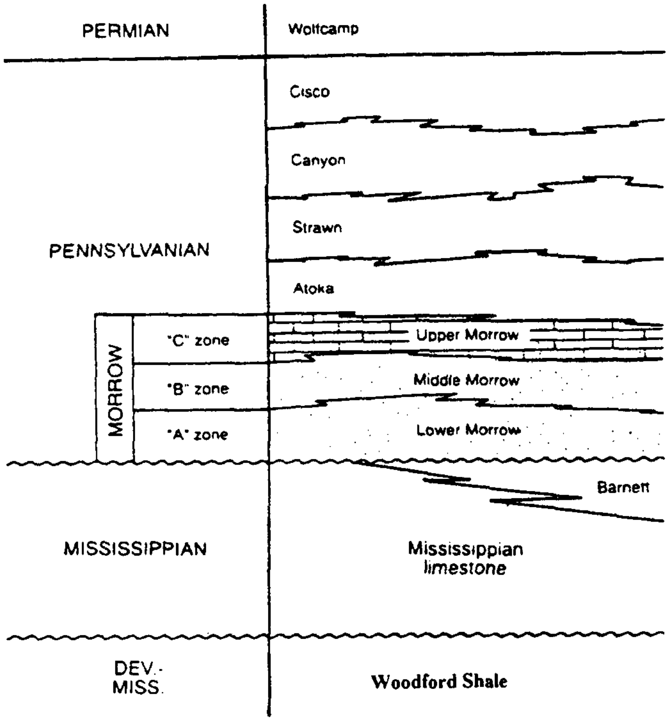

FIGURE 14—Generalized stratigraphic column of Mississippian-Pennsylvanian interval, Delaware Basin. From Mazzullo and Mazzullo (1985) and Speer (1993e).

well in Eddy County, the Atoka has been found to consist of light-gray to light yellowish-brown and dark-gray shale with some gray, medium- to coarse-grained sandstone (Grant and Foster, 1989). About 140 m of Atoka are present in this well. The Atoka thickens from about 120 m in the north to about 200 m in the southwest.

Renewed uplift during Atoka time caused an influx of clastics from the Pedernal highlands that were deposited in a shallow marine environment (Fig. 17). These sandstones were interpreted by James (1985) to be a prograding system of beaches and bars deposited along northeast trends parallel to ancient shorelines. The sand trends migrated with the shorelines into the basin during Atoka time. The first carbonate bank and patch reef structures in the Delaware Basin date from the Atoka (Hills, 1984a; Fig. 17). This pattern of reef building continued throughout the rest of the Pennsylvanian and reached its zenith in the Permian with the Capitan Reef Complex.

A major unconformity exists in the latest Atokan that is associated with renewed uplift of structurally high blocks, where chert conglomerates were deposited in structural lows. This unconformity is followed by transgression and initial deposition of Strawn shales and limestones.

Des Moinesian Series

Strawn Formation

The Strawn Formation is separated from the Atoka by a minor unconformity, but no regional tectonism intervened during this time to mark the boundary between these two units (Adams, 1962). Carbonate bank-reef mound development increased in Middle Pennsylvanian-Strawn time when a northeast-southwest trending line of algal (*Ivanovia*) reefs formed along the northwest margin of the basin (Fig. 18). The Strawn mounds were deposited as shelf-edge detrital carbonates around the Delaware Basin and Tatum Basin to the north (Harris, 1990).

The Strawn consists mostly of brown limestone and gray shale. The limestone is massive and contains a varied sequence of both dark- and light-colored limestone, fine- to medium-grained arkosic sandstone, dark- to light-gray shale, and occasional reddish-brown, greenish-gray, or bituminous shale (Netherland et al., 1974). In the west it contains abundant fine- to coarse-grained sandstone and some red shale and limestone pebble conglomerate. In the Eddy County Delaware Basin reference well, the Strawn interval is 54 m thick and is limited to a yellowish-brown to dark-gray, argillaceous, cherty limestone with a minor amount of dark-gray shale (Grant and Foster, 1989). The Strawn ranges in thickness from 90-120 m over most of the Delaware Basin, but thicknesses of more than 200 m occur in the vicinity of the Strawn reefs in the northwestern part of the basin (Netherland et al., 1974). A wide range of fossils can be found in the unit: brachiopods, forams, bryozoans, corals, and crinoids (Andersen, 1976).

Missourian Series

Canyon Formation

In the Late Pennsylvanian dark shales continued to characterize the negative deeps of the Delaware Basin, and water depths in several of the depressions may have exceeded 600 m. During the Late Pennsylvanian the carbonate and siliclastics of the Canyon and Cisco formations were deposited. The Canyon interval of Late Pennsylvanian age is mostly a brown limestone/dolomite and gray shale with some white sandstone and conglomerate near the base (Grant and Foster, 1989). Sandstone decreases to the east and the carbonate section is mostly dolomite. Chert is more abundant in the east. The Cisco/Canyon together are often designated as "Upper Pennsylvanian" rock. Major reservoirs (e.g., the Indian Basin gas fields) occur in the dolomite banks of the Cisco/Canyon (Fig. 178).

Virgilian Series

Cisco Formation

The Cisco Formation consists of limestone/dolomite with some medium- to dark-gray and red shale and minor light-

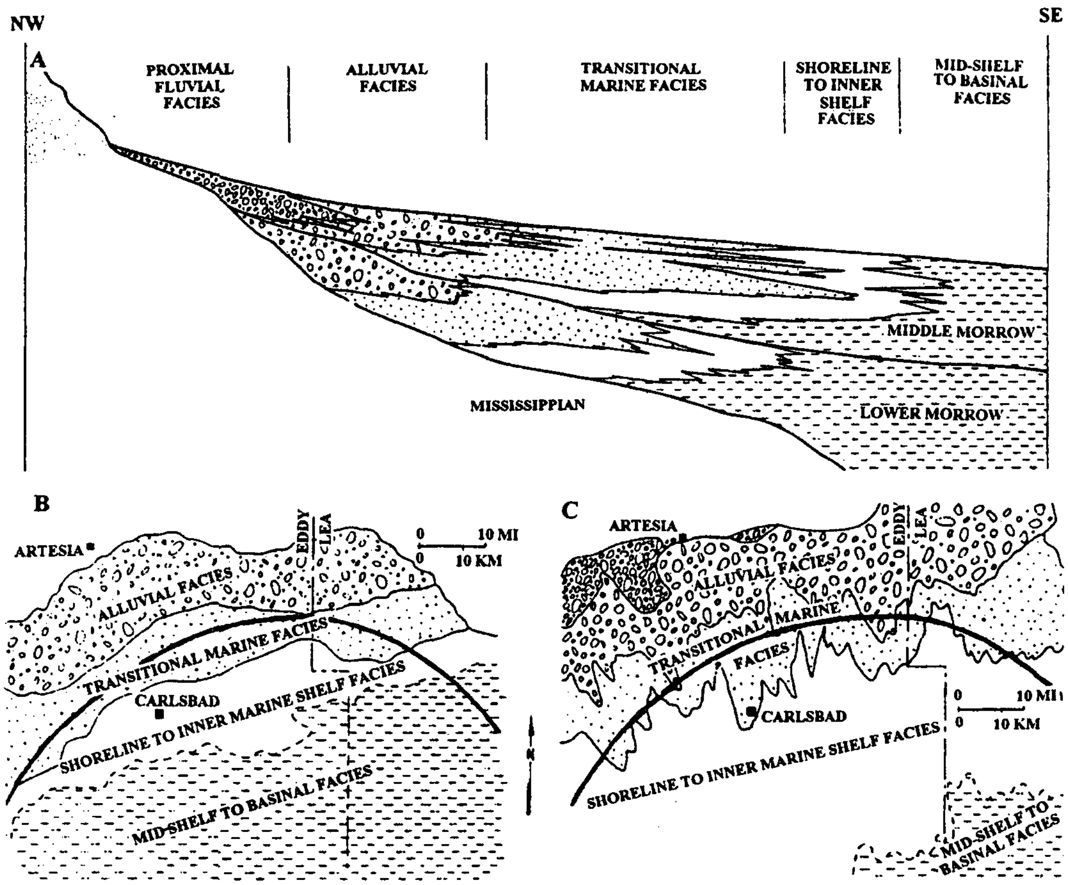

FIGURE 15—Schematic regional cross section of Eddy and west Lea Counties showing cyclic progradational and transgressive clastic facies tracts and patterns in both the lower and middle Morrow intervals. The facies tracts in (A) are mapped in (B) for the lower Morrow and in (C) for the middle Morrow. Outline of the Delaware Basin study area is shown in (B) and (C). From Mazzullo and Mazzullo (1985) and Speer (1993e).

gray, generally fine-grained sandstone (Grant and Foster, 1989). The amount of red shale decreases from west to east, but some is still present in the easternmost part of the area. King (1948, p. 12) reported rocks of Pennsylvanian age containing *Triticites* in the Updike Williams no. 1 well, located 4.8 km south of El Capitan in the northern Delaware Mountains and thought this rock to be stratigraphically "a little higher than the top of the Canyon" (probably the Cisco). David (1976) described the Cisco in Gulf no. 1 Springs Unit well as a white to tan to brown, medium- to coarse-grained dolomite with interstitual to cavernous porosity.

Glass Mountains

Gaptank Formation

The only place in the Delaware Basin where Pennsylvanian rocks crop out is in the Glass Mountains. The Gaptank Formation of Upper Pennsylvanian (Missourian-Virgilian) age was named by Udden et al. (1916) for Gap Tank, Pecos County, Texas, where an earthern water tank was once situated and where a nearly complete, 550 m thick, section is exposed. Böse (1917) described the Gaptank as a gray limestone containing brachiopods, pelecypods, and gastropods, with rare ammonoids. Udden (1917a), Baker and Bowman (1917), Keyte et al. (1927), Schuchert (1927), King and King (1928), Keyte (1929), Smith (1929), Plummer and Scott (1937), King (1937), and Cooper and King (1957) also described Gaptank fossils. Middle and upper parts of the Gaptank contain *Triticites*, *Schistoceras*, and numerous other Upper Pennsylvanian fossils. Ross (1965) identified 11 new species of *Triticites* and a new species of *Waeringella* from the Gaptank. The lower part of the Gaptank contains *Fusulinella*, *Chaetetes*, and *Chonetes mesolobus* among others.

The Gaptank is seen only in the basal cliffs and foothills of the Glass Mountains that face the Marathon Basin to the south. The rest of the earlier Carboniferous (Mississippian-Pennsylvanian) is found in the Marathon Basin as a great clastic group, strongly folded and nearly 2400 m thick (Ross, 1986). This group includes the Tesnus Formation, Dimple Limestone, and Haymond Formation, rocks that do not occur in the study area and which will not be discussed further. However, it should be remembered that these formations do occur in the subsurface of the Glass Mountain area and are correlative with other, much-less deformed, Mississippian and Pennsylvanian rocks within the Delaware Basin. The Tesnus Formation in the Marathon Basin is approximately equivalent to the Mississippian limestone and Chester shale in the Delaware Basin, the Dimple to the Morrow, and the Haymond to the Atoka (Wilde, 1990a).

The Gaptank Formation is exposed from Wolf Camp northeast to Gap Tank at the southern edge of the Glass Mountains and study area. Its lower contact with the Haymond Formation is unconformable (Fig. 19), and its

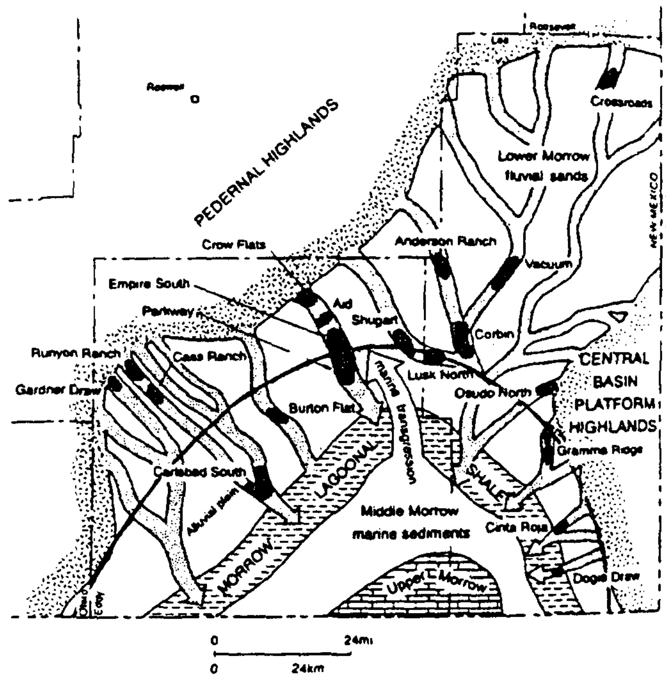

FIGURE 16—Schematic paleogeographic map of southeastern New Mexico with clastic depositional axes and related facies patterns during early Morrowan time. Several Morrow reservoirs have been superimposed. The Delaware Basin study area is shown in outline. From James (1985) and Speer (1993e).

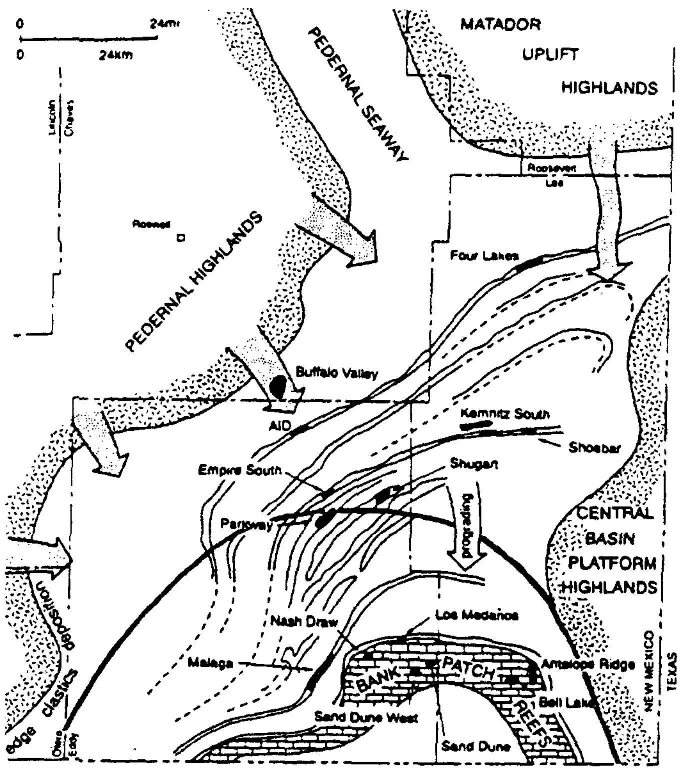

FIGURE 17—Schematic paleogeographic map of southeastern New Mexico during Atokan time with various progradational strandlines and their resultant sand trends. Several significant gas reservoirs are superimposed. The Delaware Basin study area is shown in outline. From James (1985) and Speer (1993d).

upper contact with the Wolfcamp Formation is also unconformable (King, 1937). In places the Gaptank Formation is separated from the Wolfcamp by a strong angular unconformity, a relationship that is clearly shown in many places in the southwestern part of the Glass Mountains.

King (1930) chose the *Chaetetes*-bearing limestone bed as the base of the Gaptank; however, Ross (1967) later restricted the type Gaptank to shallow-water shelfal clastics and carbonates deposited on the surface of the Marathon allochthon after a time of tectonic movement, and as such, transferred the *Chaetetes*-bearing limestone member to the underlying Haymond Formation and placed the Gaptank over faulted and folded strata of Des Moinesian and older rock (Fig. 19). The Gaptank is unconformably overlain by the Neal Ranch Formation of Permian Wolfcampian age. King (1942) defined the break between the top of the Pennsylvanian Gaptank and base of the Permian Wolfcamp as occurring at the bottom of a *Pseudoschwagerina*-bearing limestone (Wolfcamp) which lies directly on the *Uddenites*-bearing shale member. However, Ross (1959) was of the opinion that King's "*Uddenites* zone" did not represent the boundary between the Pennsylvanian and Permian; instead, Ross thought that the overlying "gray limestone member" of King was Pennsylvanian in age based on Pennsylvanian (Cisco) faunas. Therefore, according to Ross, the Pennsylvanian Gaptank includes both the *Uddenites* zone and "gray limestone member" thought by King to be part of the Permian Wolfcamp Series (Ross, 1963a, 1987a). Cys (1975) and Davis (1984) thought that the base of the gray limestone member seemed like the most logical place for the Pennsylvanian-Permian boundary.

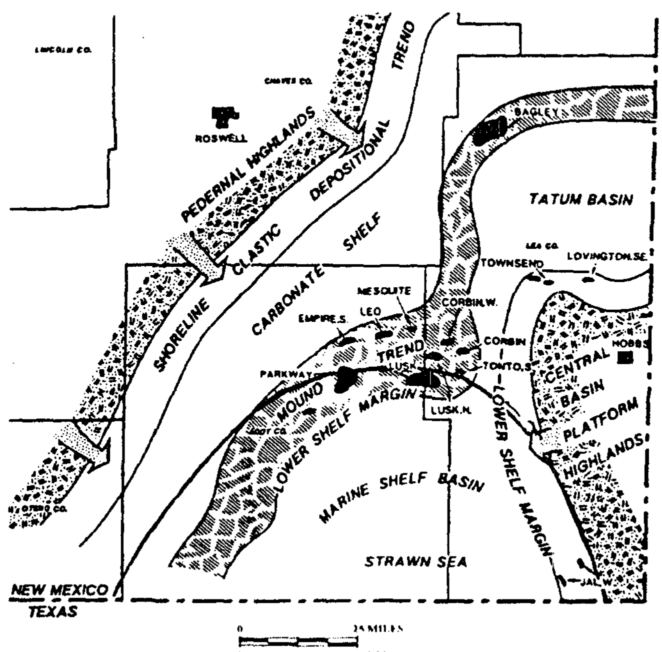

FIGURE 18—Depositional environments during Strawn time and oil reservoirs hosted by the Strawn Formation. The Delaware Basin study area is shown in outline. From James (1985).

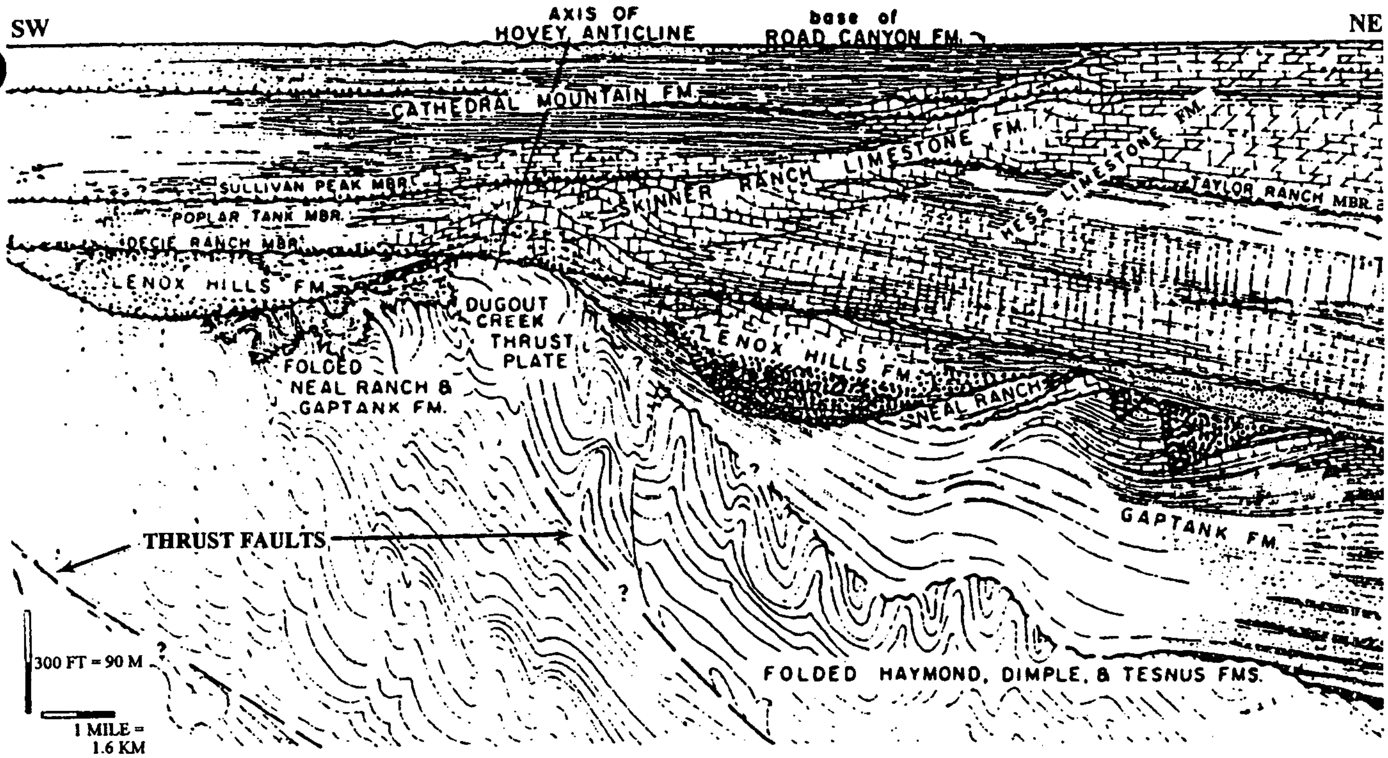

FIGURE 19—Generalized stratigraphic and structural relationships of late Paleozoic rock in the northern part of the Marathon Basin and Glass Mountains. Datum is the base of the Road Canyon Formation (i.e., King's, 1930, "1st Word limestone"). After Ross (1962b, 1968a).

Most of the fossils described from the Gaptank in the Glass Mountains are long-ranging and of little correlation significance, but the fusulinids *Fusulinella meeki* and *haworthi* indicate that the lowest part of the lower Gaptank is probably of Canyon (Missourian) age, and ammonoids and *Triticites* fauna indicate that the upper Gaptank is probably of Cisco (Virgilian) age (Keyte et al., 1927; Keyte, 1929; King, 1930). However, Ewing (1991) placed the Gaptank Formation in the mid-Des Moinesian, and Wilde (1990a) placed it equivalent to the Strawn (Des Moinesian), Canyon (Missourian), and Cisco (Virgilian). Schuchert (1927) noted that 150-180 m of upper Gaptank was absent in the Glass Mountains. King (1930) attributed this period of non-deposition to deformation, uplift, exposure, and erosion during the Late Pennsylvanian/Early Permian during the Marathon-Ouachita disturbance.

Ross (1967, 1986) described the Gaptank Formation at its type section as having three informal members: a lower conglomerate member 165-195 m thick, a middle sandstone and shale member 160-180 m thick, and an upper limestone member 270 m thick. In the lower member are five well-defined limestone-cobble conglomerate layers, each lenticular and disconformable. Large limestone and chert cobbles occur near the base of the conglomerate member. The unit varies in thickness from place to place and suggests that the conglomerates are either local lenses filling early Gaptank river channels or river-mouth bars and spits (Ross, 1967). Probably the conglomerates and sandstones filled valleys cut into the deformed Haymond, Dimple, and Tesnus formations (Ross, 1978a; Fig. 19). The middle sandstone and shale member is the least exposed of the three members. This member is, in part, a lithologic facies of both the underlying conglomerate member and the overlying limestone member. The fine-grained sandstones are well-sorted and evenly bedded suggesting deposition in a zone of wave action. Ross (1963a) was of the opinion that the middle member was deposited in a shallow-water environment such as might occur in a strongly wave-agitated backreef area. The upper limestone member consists of a lower, 120-135 m thick, resistant, massive limestone and an upper 120-150 m thick, less-resistant limestone. Higher limestones form a low discontinuous cuesta westward along the base of the Glass Mountains to the Wolf Camp Hills.

Ross (1965, 1967) also recognized three major facies in the upper limestone member where changes in faunal content can be seen over short distances: a shallow-shelf limestone facies, a shelf-edge limestone facies, and a deeper-water limestone facies. The shallow-shelf limestones are very light-gray and have abundant algal fossils. This facies was apparently deposited on a shallow carbonate shelf and subjected to strong wave and current action (Ross, 1967). The shelf-edge facies limestones are medium-gray, poorly- sorted, and commonly thick-bedded. In some beds biostromes of fossil fragments may reach 6 m in thickness and a hundred meters in diameter. This facies seems to represent rapid accumulation of poorly-sorted shell debris as carbonate banks at the outer edge of the shallow-water carbonate shelf (Ross, 1967). The banks were probably not wave-resistant (reef) structures. The deeper water limestone facies grades laterally into the shelf-edge limestone facies in one direction and tongues out into sandstone and shale in the other direction. This facies was believed by Ross (1967) to represent the deeper water, basinward thinning of shelf-edge carbonate banks. These limestones are dark-gray, pyritic, and poorly-sorted calcilutites and calcarenites.

Gaptank time began quietly but sometime after the commencement of early Gaptank time, tectonic folding began in the Marathon Basin to the south. The end of Gaptank deposition was marked by the culmination of the Marathon-Ouachita disturbance and rocks formed before and during this time period were highly compressed and thrusted northward for many kilometers (Fig. 19). The uplifted and folded strata of the Gaptank were then subjected to exposure and erosion. A great hiatus in time occurred before the sea again traversed over the region in the Early Permian.

Permian

More is known about Permian-age rock in the Delaware Basin than all of the pre-Permian rock combined. This is because an estimated 95% of all the outcrops in the Delaware Basin date from this period. Permian strata in the Delaware Basin reach thicknesses of over 2000 m.

Adams et al. (1939) were the first to divide the Permian in the Delaware Basin into four series: the Wolfcampian, Leonardian, Guadalupian, and Ochoan. These divisions were based primarily on the fossil content of the rock (Table 1), but also on its lithology. The Wolfcampian Series, named for the town of Wolf Camp in the Glass Mountains, is characterized by species of the fusulinid genera *Pseudoschwagerina*, *Schwagerina*, *Paraschwagerina*, and *Triticites*; the ammonoid genus *Properrinites*; and the brachiopod genus *Parakeyserlingina* (King, 1931; Adams et al., 1939; Skinner and Wilde, 1955). It is also characterized by thick sequences of shale and limestone. The Leonardian Series was named for Leonard Mountain in the Glass Mountains. The unit contains the brachiopods *Prorichthofenia*, *Scacchinella*, and *Dictyoclostus*; the ammonoids *Perrinites;* the conodonts *Neo-* and *Meso-gondolella idahoenis*; and the fusulinid *Schwagerina* in the lower Leonardian and *Parafusulina* in the middle and upper Leonardian (Table 1). It is composed primarily of shale and thin-bedded limestone. The Guadalupian Series, named for the Guadalupe Mountains, is characterized by massive reef limestones and dolomites which rim practically the entire basin (Fig. 2). The lower Guadalupian constitutes the zone of the conodonts *Neo-* and *Meso-gondolella serrata*, the ammonoid *Waagenoceras*, and the fusulinid *Parafusulina*, and the upper Guadalupian is dominated by *asserta* condonts and *Polydiexodina* fusulinids (Skinner and Wilde, 1955; Wardlaw et al., 1990). The Ochoan Series, named for the old Ocho post office in T24S, R34E, Lea County, New Mexico, is characterized by some of the thickest evaporite deposits in the world. It is unfossiliferous except for a few pelecypods found in non-evaporitic units. Lucas and Anderson (1994) rejected the use of the term "Ochoan" as a Late Permian stage (series) because the unit generally lacks fossils and absolute dates necessary for an ideal stratotype, and viewed these strata as a lithostratigraphic unit, the Ochoa Group. However, since the Ochoan Series is firmly entrenched in the literature of the Delaware Basin, it is retained in this publication.

In Wolfcampian time the seas spread over the whole of West Texas and southeastern New Mexico. In Leonardian time they became progressively restricted so that belts of red beds and evaporites encroached farther toward the Delaware Basin (Fig. 24), and by the end of Guadalupian time the seas had became entirely restricted to the Delaware Basin (King, 1942; Fig. 34). During Ochoan time the basin became desiccated and filled with evaporites; these became topped by red beds during the final demise of the Permian sea. Superimposed on the overall retreat of the Permian sea are major cycles of transgression and regression which were dependent both on tectonic and sea level (eustasy) changes (Ross and Ross, 1987; Sarg, 1991; Fig. 23). Upon these major cycles are superimposed many smaller, minor cycles which have left records of their passing as individual beds or laminae ranging from meters to millimeters in thickness.

The Permian in the Glass Mountains is considered the standard section for the region because a complete and continuous 1500-2000 m rock sequence exists there, from the base of the Wolfcampian up to the top of the Guadalupian. The fossil record in the Glass Mountains was used to set up stratigraphic divisions for the rest of the Delaware Basin and also for other Permian rock in North America (Adams et al., 1939; Stropoli, 1991b). Permian stratigraphic units in the Glass Mountains are time-correlative with units in the Guadalupe and Apache Mountains, but the names of formations are different (Fig. 20). The reason for this is two-fold: (1) stratigraphic units in the Glass Mountains were studied (and named) independently from units in the Guadalupe and Apache Mountains, and it was not until the late 1920's that the various stratigraphic sections around the basin were correlated, and (2) there is a dissimilarity of lithology in the Glass Mountains as compared to the Guadalupe and Apache Mountains because the source of Permian rock in the Glass Mountains was the Marathon-Ouachita fold belt, while that for the Guadalupe and Apache Mountains was the Pedernal landmass/Diablo Platform (Fig. 13).

The Permian stratigraphy of the Glass Mountains was worked out early in the century by Udden (1917a) and King (1930), but since then, work in these mountains has been limited (compared to the Guadalupe Mountains), with the notable exceptions of the works of Cooper and Grant (1964, 1966, 1972, 1977); Ross (1959, 1960, 1962a,b, 1963a,b, 1964, 1965, 1967, 1978a,b, 1979, 1981, 1984, 1986, 1987a,b); Ross and Oana (1961); Ross and Ross (1985, 1987); and a couple of dozen master theses and other studies

TABLE 1—Permian correlation fossils, Delaware Basin

Series	Location	Formation	Zone Fossils	Other fossils	References
Guadalupian	Guadalupe Mtns.	Tansill	*Paraboultonia*	Sponges, algae, brachiopods,	Richardson (1904)
upper		Yates	*Polydiexodina*	pelepods, *Yabeina*, gastropd.	Girty (1908)
		Seven Rivers		Gastropods (most abundant),	Dunbar & Skinner (1937)
		Queen		crinoids, brachiopods, algae,	Adams et al. (1939)
		Grayburg		nautiloids, pelecypods	King (1942; 1948)
		Capitan	*Polydiexodina*	Algae (*Mizzia*), sponges,	Newell et al. (1953)
				ammonoids (*Timorites*)	Hayes (1964)
	Apache Mtns.	Capitan	*Polydiexodina*	Algae (*Mizzia*), sponges	Behnken (1975)
	Glass Mtns.	Capitan	*Polydiexodina*	Algae, sponges, bryozoans,	Clark & Behnken (1979)
				brachiopods, corals	Wardlaw et al. (1990)
	Basin	Bell Canyon	*Polydiexodina*	Ammonoid (*Timorites*)	Kozur (1992)
		Altuda	*N. postserrata*	brachiopods, bryozoans	
middle	Guadalupe Mtns.	Goat Seep	*Parafusulina*	Sponges, brachiopods,	
	Apache Mtns.	Goat Seep/Munn	*Parafusulina*	pelecypods, bryozoans	
		C. Can. ss tongue	*Parafusulina*		
	Glass Mtns.	Vidrio	*Parafusulina*	Ammonoid (*Waagenoceras*)	
		Word	*N. serrata*		
	Basin	Cherry Canyon	*Parafusulina*	Pelecypods, brachiopods,	
			N. aserrata to	algae, coral, bryozoans	
			serrata change		
lower	Basin	Brushy Canyon	*Parafusulina*	Brachiopods	
			N. serrata		
Transition:	Guadalupe Mtns.	San Andres-lower	*Parafusulina*	Brachiopods and corals	Boyd (1958)
Leonardian-		Cutoff		common; nautiloids,	Behnken (1975)
Guadalupian			*N. serrata*	echinoids, crinoids	Clark & Behnken (1979)
	Apache Mtns.	Cutoff/Vict. Peak	*Parafusulina*	*Skinnerina*	Wilde (1986a,b)
	Glass Mtns.	Road Canyon	*N. idahoenis* to	Ammonoid (*Perrinites hilli*)	
			serrata change		
Leonardian	Guadalupe Mtns.	Yeso	*Parafusulina*	Brachiopods, bryozoans,	Böse (1917)
				gastropods, crinoids	King & King (1929)
		Victorio Peak	*Parafusulina*	Brachiopods, spirifers,	Dunbar & Skinner (1937)
				productids, bryozoans,	Newell (1937)
				crinoids	Plummer & Scott (1937)
	Apache Mtns.	Victorio Peak	*Parafusulina*	Brachiopods	Miller & Furnish (1940)
	Glass Mtns.	Cathedral Mtn.	*Parafusulina*	Brachiopods (*Institella*)	Newell et al. (1953)
			N. idahoenis	*Tubiphytes*	West Texas Geological
		Skinner Ranch	*Schwagerina*	Crinoids, ammonoids,	Society (1960)
		Hess		conodonts	Headley (1968)
	Delaware Mtns./	Bone Spring	*Parafusulina*	Brachiopods, pelecypods,	Adams et al. (1939)
	basin		*N. idahoenis*	gastropods, trilobites,	Wardlaw et al. (1990)
				crinoids, sponge	
Wolfcampian	Guadalupe Mtns.	Hueco	*Pseudo-*	Brachiopods, gastopods,	Böse (1917)
		(subsurface)	*schwagerina*	fusulinids	Keyte et al. (1927)
	Delaware Mtns.	Hueco	*Pseudo-*	Brachiopods, gastopods,	Schuchert (1927)
		(subsurface)	*schwagerina*	fusulinids	King (1930; 1942)
	Glass Mtns.	Neal Ranch	*Schwagerina*	*Triticites, Uddenites*,	Dunbar & Skinner (1937)
		Lenox Hills	*Pseudo-, Para-*	Ammonoids (*Properrinites*),	Neeham (1937)
			schwagerina	brachiopods	Adams et al. (1939)

from the 1960's to the present. The Permian rocks of the Glass Mountains have been progressively subdivided since the earliest stratigraphic work of Udden (1917a), in accordance with an increased need for detail and convenience of discussion and in response to updated fossil correlations (Cooper and Grant, 1966). Udden (1917a) established the basic stratigraphic framework and introduced the Wolfcamp, Hess, Leonard, and Word formations. The Wolfcamp and Leonard formations were elevated to the rank of series by Adams et al. (1939), after which Ross (1959) subdivided and introduced new formation names for the former Wolfcamp Formation, and Cooper and Grant (1964) introduced new formation and member names for rocks of the former Leonard Formation. Permian strata in the Glass Mountains are generally divided into an eastern carbonate facies and a western clastic facies. The western facies has traditionally

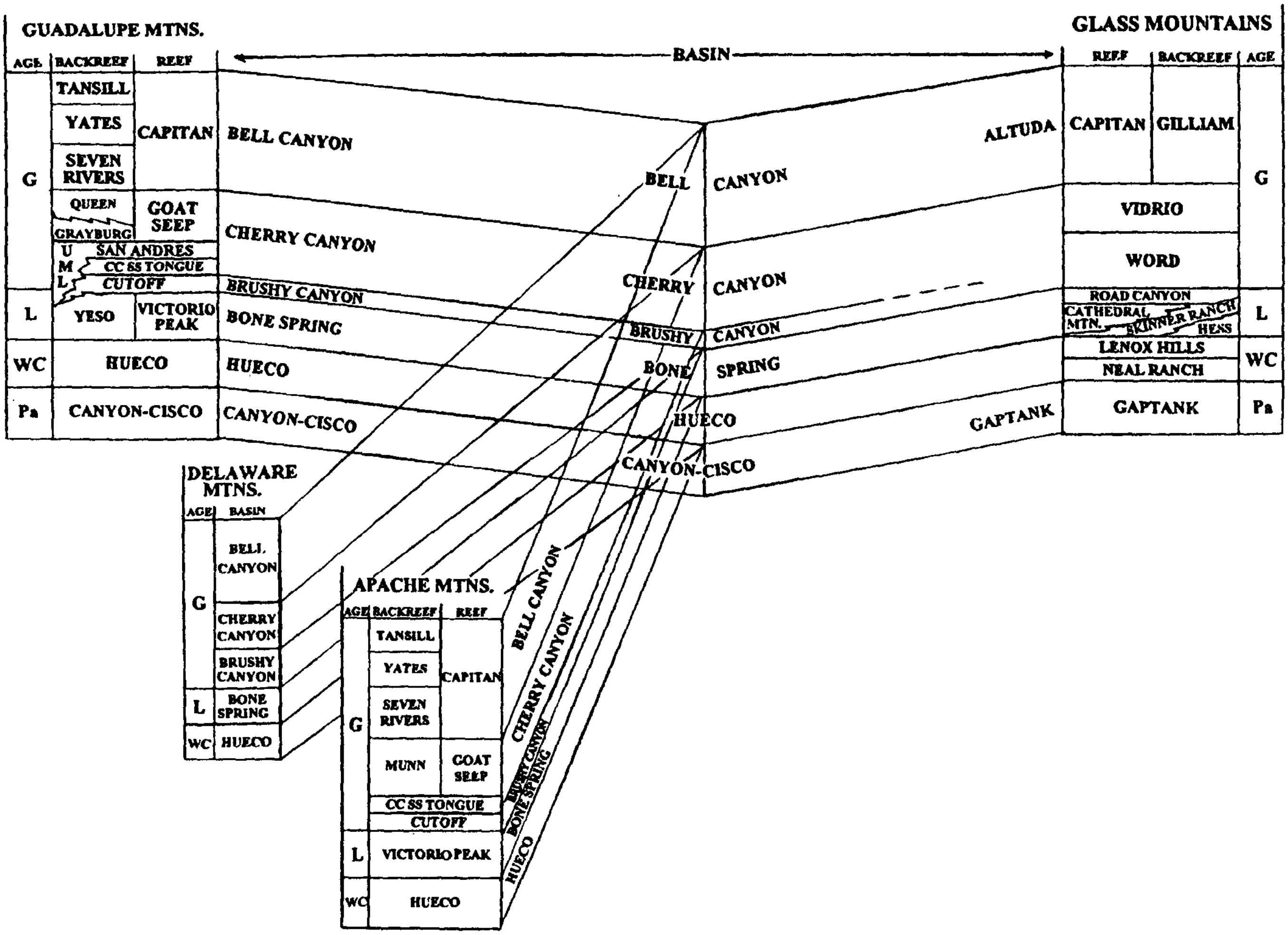

FIGURE 20—Correlation of Pennsylvanian and Permian stratigraphic units in the Guadalupe, Delaware, Apache and Glass Mountains, and basin.

been considered to represent deeper-water clastic sedimentation at the margin of the Hovey Channel, but this clastic sedimentation at the margin of the Hovey Channel, but this interpretation has recently been challenged by a number of masters theses which have determined that upper Leonardian and Guadalupian strata in this western area is of shallow-lagoon to deltaic origin (see the *Special topic*: Where is the Hovey Channel?)

Permian stratigraphy in the Apache Mountains has not been nearly as well documented as in the Guadalupe Mountains or Glass Mountains. Richardson (1914), Johnson (1942, 1951), and Johnson and Dorr (1942) described some of the fossils of the Apache Mountain area, and Blanchard and Davis (1929), McNutt (1948), and King (1949) reported on the general stratigraphy of the region, but it was not until the 1950's that this remote area was mapped in some detail, mostly by master theses students. These studies, a few field-trip guidebooks (DeFord, 1951; West Texas Geological Society, 1960; Wilde and Todd, 1968), and the comprehensive works of Wood (1965, 1968) form the basis of information on the Permian in the Apache Mountains.

Wolfcampian Series

The Wolfcampian Series crops out in the Glass Mountains and occurs in the subsurface of the Guadalupe Mountains, Delaware Mountains, Apache Mountains, and basin. The Wolfcamp in the Glass Mountains is equivalent to the Hueco Limestone in the subsurface of the Guadalupe, Delaware, and Apache Mountains (Fig. 20), and both of these units are correlatable and equivalent to the Hueco Limestone exposed in the Sierra Diablo and Hueco Mountains west of the study area. In general, the amount of limestone in the Wolfcamp increases to the north while some sandstone is present in the eastern part of the area in proximity to the Central Basin Platform. Wolfcamp sediments in the subsurface have been encountered in a number of wells. In the Delaware Basin, Eddy County, reference well, the Wolfcamp is ~480 m thick (Grant and Foster, 1989). In the Lusk field, located on the Northwest Shelf at the boundary between Eddy and Lea Counties, the lower 135 m of Wolfcamp strata is a dark-gray to black, carbonaceous shale with very thin limestone stringers. The bottom of this zone is generally referred to as the "Permo-Penn" by oil drillers (Thorton and Gaston, 1967, p. 16); in the Delaware Basin the "Permo-Penn" is predominantly a gray shale with limestone and minor dolomite. The upper 60 m of Wolfcamp strata in the Lusk field consists of interbedded limestone and dolomite with thin, dark-gray, shale stringers.

At the beginning of Wolfcampian time, the Central Basin Platform, Diablo Platform, Pedernal Massif, and Marathon-Ouachita belt were active and uplifted areas which supplied sediment to the sinking Delaware Basin (Fig. 13). The majority of sediment was eroded from the Marathon-Ouachita Mountains and thick deposits accumulated primarily in the southern part of the basin. The Wolfcampian Series progressively thins from about 2400 m of shale and limestone in the southern Glass Mountains region to 200 m or so of limestone and lesser shale in the north and west (the Hueco Limestone in the Apache Mountains, Guadalupe Mountains, and Northwest Shelf; Hayes, 1964). Dark-colored limestone, siltstone, and shale, with minor coarse sandstone and conglomerate, formed in the center of a deep-marine basin, while carbonate reefs and banks grew around the shelf margins in the more stable platform areas. These "reefs" occurred as isolated mounds ("patch" or "knoll" reefs), and algae were the main reef-building organisms (Adams and Frenzel, 1950; Newell et al., 1953). Wolfcampian bank-reefs sometimes grew on the crests of older, Pennsylvanian bank-reefs. By the end of Wolfcampian time, only the Central Basin Platform and Marathon regions were emergent above the late Wolfcampian sea.

Hueco Limestone

The Hueco Limestone of upper Wolfcamp age occurs in the subsurface in the western and northern part of the Delaware Basin, and it outcrops in the Sierra Diablo just west of the study area. The Hueco Limestone has been penetrated by a number of subsurface wells in the Guadalupe and Delaware Mountains. Upson (1951, 1960) described the Hueco in Gulf's Grisham no. 1 well, sec. 18 Blk. 99 PSL, Culberson County, Delaware Mountains, as a dark, shaly limestone, containing black chert. King (1942, 1948) described the Hueco from another well in the Delaware Mountains (the Anderson and Prichard well) as consisting of gray, fossiliferous limestone, with a basal unit of limestone-pebble conglomerate and black shale and dark limestone. In the Apache Mountains the Hueco has been penetrated by Humble's Reynold Cattle Company "B" no. 1 well where it was described as 321 m of tan and gray, crystalline dolomite with interbedded chert, and medium- to coarse-grained, gray sandstone with some shale in the basal section (Maley, 1945; Huffington, 1960). In the basin the Hueco consists of darker shale and limestone; this dark rock grades into lighter-colored dolomite and greenish-gray shale towards the Northwest Shelf (Hayes, 1964).

Brachiopods, gastopods, and fusulinids are abundant in the Hueco Limestone; pelecypods are less abundant and ammonoids are generally absent (King, 1942). Fossils in the Hueco are significantly different from those in the overlying Bone Spring Limestone of Leonardian age, and the two series do not have a single significant brachiopod, gastropod, or pelecypod species in common. As seen in the Sierra Diablo, the contact between the Hueco and overlying Bone Spring is unconformable, sharp, well-marked, and persistent. The Hueco sits unconformably on lower Wolfcamp and older sediments. The unit at its base is the Pow Wow conglomerate, which unit crops out in the Sierro Diablo and in the Hueco and Sacramento Mountains. The Pow Wow is the upper (last) unit to have been tectonically affected by the Marathon-Ouachita disturbance, prior to Leonardian-Guadalupian post-orogenic fill. More than any other event, the structure recorded by the Pow Wow defines and controls the ensuing sedimentary architecture of basin fill (R. Sarg, pers. comm., 1996).

Glass Mountains

In the Glass Mountains the earliest Permian rocks of the Wolfcampian Series rests unconformably upon strongly folded Gaptank rocks of Pennsylvanian age (Fig. 19). Wolfcampian beds are present along almost the whole of the Glass Mountains escarpment which flank the Marathon Basin to the south. At the close of Wolfcamp time there was a slight reoccurrence of crustal movement (the final pulse of the Marathon disturbance) in the Glass Mountains, and Wolfcampian beds were tilted and partially eroded before rocks of Leonardian age began to be deposited (King, 1930; Hills, 1942; Fig. 19). The entire Wolfcampian Series in the Glass Mountains was originally represented by the Wolfcamp Formation (King, 1930), but this formation was subsequently subdivided by Ross (1959) into the Neal Ranch and Lenox Hills formations. According to Wilde (1990a), the Neal Ranch is early to middle Wolfcampian in age, separated from the late Wolfcampian Lenox Hills by an unconformity which represents the close of the Marathon-Ouachita orogeny in the middle Wolfcampian.

Neal Ranch Formation

The lowest Wolfcampian beds in the Glass Mountains are part of the Neal Ranch Formation (Ross, 1959, 1963a). This formation is 90-140 m thick and consists of biohermal limestone, shale, sandstone, and conglomerate. It was named for Neal Ranch in the vicinity of the Wolf Camp Hills. The Neal Ranch Formation rests unconformably on the shelf-edge limestones of the Gaptank Formation in the Wolf Camp Hills and at Gap Tank (Ross, 1959, 1986). It also occurs at the foot of the Lenox Hills and at the base of the Hess Ranch horst.

The Neal Ranch Formation consists of a succession of 17 or more cyclical shales and calcarenites containing numerous, small, circular biostromes about 1 m high and 3-10 m in diameter. Both the shales and calcarenites have weathered to orange-brown or light-brown colors and are fetid on fresh surfaces. Often, the calcarenite beds are cemented by sparry calcite (Ross and Oana, 1961). The limestone units of the Neal Ranch crop out as ledges; these thicken and thin laterally and locally pass into biostromes 1-1.5 m thick (Ross, 1963a). Biostromal fossils include sponges, bryozoans, corals, brachiopods, and the fusulinids *Pseudoschwagerina, Paraschwagerina, Schwagerina,* and advanced species of *Triticites* (Ross, 1986). Fusulinids which characterize, and are restricted to, the Neal Ranch Formation include *Triticites uddeni, Schwagerina emaciate,* and *Pseudoschwagerina uddeni*, among others (Ross, 1959). The contact between the Pennsylvanian and Permian in the Glass Mountains is based on the first appearance of *Pseudoschwagerina*. This first appearance in the Neal Ranch does not include King's (1930) "Bed 2" which is actually part of the Pennsylvanian Gaptank Formation (Cooper, 1957; Ross, 1978a).

The Neal Ranch Formation unconformably overlies the Gaptank Formation of Pennsylvanian age and is overlain unconformably (and angularly-truncated eastward) by the basal conglomerate of the Lenox Hills Formation (Ross, 1963; Cys, 1977b), which conglomerate is equivalent to the Pow Wow conglomerate of the Hueco elsewhere in the basin. Ross and Oana (1961) found a significant difference in carbon isotope ratios across the Gaptank-Neal Ranch unconformity (Appendix 1). These values indicate that the depositional environment had changed from a shallow, high-wave energy, well-aerated environment in the Late Pennsylvanian to a low-wave energy, poorly-aerated environment in the Early Permian. Also, brachiopod (and other fossil) assemblages seem to indicate that the Neal Ranch was deposited on a shallow shelf to shelf edge, with the deepest black-shale facies being deposited in water not more than 30 m deep (Stropoli, 1991b).

The Neal Ranch Formation was deposited in the Glass Mountains prior to the last major tectonic pulse of the Marathon orogenic belt as evidenced by faulted and folded strata (Fig. 19). In the southwestern part of the Lenox Hills the Neal Ranch was faulted prior to the deposition of the overlying Lenox Hills Formation.

Lenox Hills Formation

The upper part of the Wolfcampian Series in the Glass Mountains belongs to the Lenox Hills Formation (Ross, 1959, 1963a; Fig. 20). This formation was named for the Lenox Hills, located about 5 km north of Dugout Mountain and just within the limits of the study area. The Lenox Hills exhibits a high degree of variability and displays strong facies changes across the Glass-Del Norte Mountain front (Cooper and Grant, 1972, 1977). Lenox Hill beds are somewhat discontinuously distributed along the southeast cuesta of the Glass Mountains for a lateral distance of approximately 40 km, with good exposures in the Gaptank, Wolfcamp Hills, Hess Ranch horst, Leonard Mountain, Lenox Hills, and Dugout Mountain areas (Stropoli, 1991b). The formation ranges up to 200 m thick and consists of reef and biohermal limestone, backreef limestone, shale and sandstone, forereef clastics, and conglomerate. The clastic facies represent channel-fill and deltaic deposits. Bioherms formed adjacent to the delta eventually built a massive reef-like feature that now forms the southeastern point of Leonard Mountain (Ross, 1978a, 1986). Just north of the Hess ranch house, at the entrance to Hess Canyon, the edge of the biohermal facies is exposed. Further east, reddish and yellowish shale, siltstone, and sandstone form a distinctive facies beneath the cliff-forming limestone of the Hess Formation.

At Lenox Hills the base of the Lenox Hills Formation is unconformable and marked nearly everwhere by the accumulation of conglomerates and coarse clastic deposits (Ross, 1963a). The type section consists of 40 m of conglomerate succeeded by 50 m of sandstone, clastic limestone, and shale (Ross, 1959). In the southwestern part of Lenox Hills the entire formation changes to conglomerate. Throughout the Glass Mountains the Lenox Hills Formation is marked by a persistent basal conglomerate (informally named the "Stockton Gap formation" by Wilde, 1990b), but the strata above can change facies within short distances. At the top of the Lenox Hills is a marked unconformity which separates Wolfcampian and Leonardian strata (Fig. 19). Fusulinids which characterize, and are restricted to, the Lenox Hills Formation are *Schwagerina extumida* and *Pseudoschwagerina tumidosus* (Ross, 1959). Based on faunal correlations Ross (1963a) thought that the Lenox Hills Formation was equivalent in age to most of the Hueco Limestone (Fig. 20).

Flores et al. (1977) and Flores and McMillian (1981) described the Lenox Hills Formation in the Leonard Mountain area of the Glass Mountains as a calcirudite, calcarenite, biolithite, calcilutite and dolomite. Vertical and lateral variations of these lithofacies suggested to these authors a subtidal to intertidal environment of deposition for the Lenox Hills, and they concluded that progradational deltaic facies had interfingered laterally with areas of non-turbid carbonate bank deposition which had allowed for enhanced biohermal growth. Ross (1986) stated that the Lenox Hills Formation represented a diverse set of marine, marginal marine, and non-marine facies traceable laterally along the entire lower face of the Glass Mountains.

Leonardian Series

During Leonard time the Delaware Basin continued to subside, although not as rapidly as during Wolfcamp time. In the late Wolfcampian-early Leonardian, sections of the basin in the north, northwest, and northeast were shallow enough and free enough of clastic material for carbonate banks to grow along the seaward edge of these shallows (Adams, 1965). These banks, called the "Abo reefs," were the first of the reef formations in the Leonardian Series and are correlative with early Bone Spring beds (Fig. 21). (These reefs, known as the Abo Formation, are not located within the boundary of the study area and will not be discussed further).