OF-WTWI-1984-43

STRATIGRAPHIC STUDIES OF THE PALO DURO BASIN: AN UPDATE (1984)

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Prepared for the
U.S. Department of Energy
Office of Nuclear Waste Isolation
under contract no. DE-AC-97-83WM46615

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1984

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INTRODUCTION

Stratigraphic studies of the Palo Duro Basin are now in their 8th year. Although the emphasis in these studies now lies in the San Andres Formation (possible repository host rock) and the Wolfcamp Series (sub-repository aquifer), investigations of other units (fig. 1) continue to be carried out as well. This report includes work that was not available for inclusion in last year's (1983) CSR on stratigraphy (Open File Report No. OF-WTWI-1984-30). The section on pre-Pennsylvanian stratigraphy represents the final stages of study of those units. The report detailing structural control on deposition of the San Andres is part of an ongoing effort to recognize indications of structural or tectonic controls on sedimentation throughout the stratigraphic column in the Palo Duro Basin area. Studies of the Dockum Group are revealing similar controls during the Triassic. Core studies of the Dockum are increasing our resolution of depositional settings in the area.

STRATIGRAPHY AND DEPOSITIONAL SETTING OF PRE-PENNSYLVANIAN ROCKS IN THE PALO DURO BASIN AREA

(S. C. Ruppel)

Introduction

The Palo Duro Basin is one of five sedimentary basins that, along with associated arches and uplifts, make up the Texas Panhandle area (fig. 1). These basins, which along with others to the north and south comprise the Permian Basin, were formed primarily by tectonic forces first active in the Pennsylvanian. The Palo Duro Basin, as it is defined by most, is bounded on the south by the Matador Arch, on the north by the Amarillo Uplift, and on the west by a slight positive feature that separates the Palo Duro from the Tucumcari Basin in central New Mexico (Budnik and Smith, 1982). To the east, separated from the Palo Duro by another slight positive feature, are the Hardeman and Hollis (or Harmon) Basins (Totten, 1956). In the past, some have included the Hardeman and Hollis Basins as part of the Palo Duro. The Dalhart Basin is a minor subbasin separated from the Palo Duro by a positive feature known as the Bravo Dome (fig. 2).

The Palo Duro and associated basins in the Texas Panhandle area contain rocks ranging in age from Precambrian to Recent (fig. 1). The pre-Pennsylvanian sequence, however, comprises only questionable Cambrian, Ordovician (Lower), and Mississippian rocks (fig. 1). Silurian and Devonian deposits are absent. Pre-Pennsylvanian rocks range in thickness up to about 1,000 to 1,200 ft (305 to 366 m) in the Palo Duro and Dalhart Basins and as much as 2,000 to 2,500 ft (610 to 732 m) in the Hardeman and Hollis Basins.

Studies of pre-Pennsylvanian rocks in the Palo Duro Basin area are relatively few. The most comprehensive early study is that of Totten (1956). Other summaries have been published by Roth (1955), Huffman (1959), Nicholson (1960), Best (1963), and Soderstrom (1968). An analysis of the Dalhart and Anadarko Basin area was prepared by Cunningham (1969). More recently, Mapels and others (1979) characterized the Mississippian of the area. Dutton and others (1982) gave a brief summary but no detailed study of pre-Pennsylvanian rocks has been published previously for the Palo Duro area. Such a report is long overdue, considering the continuing interest in the area for possible exploration targets for oil and gas.

Methods

Geophysical well logs from more than 7,500 wells in 57 counties of Texas, Oklahoma, and New Mexico have been examined for this study. About 250 wells penetrate pre-Pennsylvanian units in the Palo Duro Basin. Commercially-prepared sample logs were available for about 175 of these. Rock core was described from one well in the Palo Duro Basin area and eight wells in the Hardeman Basin (fig. 2). Cuttings were examined in 115 wells.

Structural Setting

The general structure of the Texas Panhandle area is indicated by contour maps on the top of the Ellenburger (fig. 3) and the top of the Mississippian (fig. 4). The axis of the Palo Duro

Basin generally trends east-west with the deepest parts occurring in southeastern Floyd/southwestern Motley Counties. Mississippian rocks reach depths of more than 7,500 ft (2,286 m) below sea level in this area (fig. 4). Most of the faulting apparent from contour mapping has a northwest/southeast trend. Seismic data available from the central and western parts of the basin support this interpretation, but also reveal a secondary population of northeast/southwest trending faults (R. T. Budnik, personal communication, 1984). Seismic data also indicate that faulting is much more prevalent than is suggested by structural contour maps based on available well data.

The Matador Arch, which forms the southern boundary of the Palo Duro Basin is a complex feature composed of isolated high areas commonly bounded by faults (fig. 4). Faulting along this structure is ubiquitous and complex apparently resulting in numerous small fault blocks. The NRM field, the only pre-Pennsylvanian (Mississippian) hydrocarbon discovery in the Palo Duro Basin area (now abandoned) appears to be located on a small fault sliver in one of these extremely complex areas.

The nature of the margins of pre-Pennsylvanian units in much of the area is problematic. Most depictions tend to suggest that it is erosional (Huffman, 1959; Nicholson, 1960). This is most certainly true of the Ellenburger along the Texas Arch (fig. 3). Available seismic data, however, indicate that in some areas these margins are fault-controlled (for example, in western Deaf Smith County) as indicated by the somewhat linear nature of many segments of these contacts.

The Hardeman Basin area is separated from the Palo Duro Basin by a positive area of low relief that extends generally north/south through Cottle and Childress Counties (fig. 4). The Hardeman Basin, in turn, is apparently separated from the Hollis by an east/west line of high structures and a similarly trending fault zone. Displacement along this fault zone exceeds 1,000 ft (305 m). Depths in the Hardeman Basin are generally similar to those in the Palo Duro; the Hollis Basin is somewhat shallower (fig. 4).

The Dalhart Basin occupies the northwestern corner of the Texas Panhandle (fig. 2). It is separated from the Anadarko by northwest and northeast trending faults and a north/south trending positive feature (Cimarron Arch/Keyes Dome). Pre-Pennsylvanian units in the Dalhart Basin are quite shallow (fig. 4).

Stratigraphy and Depositional Environments

The pre-Pennsylvanian sequence of rocks in the Palo Duro, Dalhart, and Hardeman Basins comprises only three parts: (1) a basal, thin unit of terrigenous (Cambrian?) clastics, (2) an overlying interval of Lower Ordovician (Ellenburger Group) dolomites, and (3) an uppermost sequence of Mississippian carbonates that is predominantly composed of limestones. Although these deposits are variably developed throughout the area, figure 5 illustrates typical sections for each of the basins.

Basal Clastics (Cambrian?)

Thin beds of terrigenous clastics overlie the basement in several parts of the area (fig. 6). Although thick sequences of these deposits have been reported from the Hardeman Basin (Montgomery, 1984), throughout most of the area thicknesses greater than 50 ft (15 m) are rare. In most cases, these deposits comprise rounded, quartz sandstones. Gray and green shales are observed at some localities; clasts of dolomite and/or limestone are also locally present. These basal sandstones grade downward into the underlying weathered basement rocks in many places making precise distinction of the two units locally difficult.

The distribution of these basal sandstones generally corresponds to that of the overlying Ellenburger Group (fig. 3). This indicates that these deposits once covered the entire Panhandle in a thin veneer. Middle Paleozoic erosion along the Texas Arch (fig. 3; Adams, 1954) removed most of these deposits in the central part of the Palo Duro Basin along with the Ellenburger. There are at least two areas in the Palo Duro Basin, however, where substantial thicknesses of basal, rounded quartz sandstones are present without any apparent association to the distribution of the Ellenburger. In southeastern Swisher and southeastern Floyd Counties (fig. 6), more than 200 ft (61 m) of such deposits have been reported. These rocks are overlain by Mississippian carbonates.

The exact age of the basal clastics in the Panhandle area is unknown. In some instances, they have been correlated with the Hickory Sandstone (member of the Cambrian Riley Formation), which crops out in central Texas (Barnes and others, 1959). Other deposits of similar lithology are known, however, from younger Cambrian and Ordovician units (Wilberns Formation) that outcrop in central Texas. Barnes and others (1959), in fact, indicate that the Wilberns overlies basement in northern Texas. This suggests that the basal clastics of the Panhandle area should be regarded as Wilberns, not Riley (Hickory). In the absence of good core data, however, the precise correlation of the Panhandle basal sandstones cannot be established. Because of their stratigraphic position below the Ellenburger Group, it is likely that most of these deposits are Cambrian in age. They were probably formed during the general marine transgression of the area at the beginning of the Paleozoic.

The origin and age of the thick deposits of sandstone in Swisher and Floyd Counties is more enigmatic. Because they are overlain by Mississippian rocks, several possibilities exist: (1) they may be basal deposits formed during the transgression of the area in the late Devonian/early Mississippian, (2) they may be Precambrian sandstones that have been exposed by erosion along the Texas Arch, or (3) they may be equivalent to the other Cambrian basal clastics in the area, preserved because they were deposited in structural depressions or fault blocks. Meager evidence favors the latter interpretation. The thick accumulation in Floyd County, for example, is known to coincide with a structural low between uplifted blocks along the Matador Arch (R. T. Budnik, personal communication, 1984).

Lower Ordovician, Ellenburger Group

The Ellenburger Group comprises all Lower Ordovician deposits in the subsurface areas of north and west Texas and southeast New Mexico (Barnes and others, 1959). Partially equivalent rocks in Oklahoma and the northern part of the Texas Panhandle are included in the Arbuckle

Group. By definition, the Arbuckle differs from the Ellenburger by containing Upper Cambrian as well as Lower Ordovician rocks. For purposes of this report, however, all of these deposits in the Texas Panhandle area will be referred to as Ellenburger, since this term is more common in Texas usage and because the exact age of these deposits in not known.

The Ellenburger is present throughout the Texas Panhandle except where it has been removed by erosion along the Amarillo Uplift and Texas Arch (fig. 7). Where present in the Palo Duro and Dalhart Basins, the Ellenburger reaches maximum thicknesses of only about 500 ft (152 m). In the extreme eastern part of the Palo Duro Basin, thicknesses of greater than 1,000 ft (305 m) are known (fig. 7). Although thicknesses as great as 2,000 ft (610 m) have been indicated in the Hardeman and Hollis Basin areas (Barnes and others, 1959; Huffman, 1959), such values are not supported by available well data. It is clear, however, that the Ellenburger does thicken markedly into the area immediately south of the Amarillo-Wichita Uplift (Collinsworth, Childress, and Hardeman Counties, Texas, and Harmon County, Oklahoma).

Sample logs indicate that the Ellenburger Group comprises fine- to coarse-grained, sucrosic to rhombic dolomite throughout the area. Shale and medium to coarse-grained, rounded, quartz sandstone are locally common. Limestone is extremely rare. Chert is common throughout most of the Ellenburger; in many areas it is oolitic. Glauconite and pyrite are present as minor accessory minerals. Glauconite is especially common at the base and the top of the Group. Color in the Ellenburger is quite variable. Dolomite is most commonly gray to brown, but white, cream, pink, and yellow colors are also reported. These rocks do not, however, show a progressive southwest to northeast darkening of color as suggested by Barnes and others (1959). Most shales in the Ellenburger are reported as waxy and gray-green in color; red-brown shales are less common. Chert is most commonly white to pink, although shades of blue are also reported.

Examination of available core in Hardeman County shows the Ellenburger to be composed nearly entirely of crystalline dolomite. Allochems are rarely preserved. Brecciated zones are present at irregular intervals throughout the Group; they are particularly common at the top of the sequence. In general, the Ellenburger of the current study area is quite similar to that described by Folk (1959). The Arbuckle Group of Oklahoma is also apparently similar (Cardwell, 1977).

The Ellenburger was deposited in a quite shallow marine setting that covered large areas of the North American continent during the Lower Ordovician (Cloud and Barnes, 1948). There is little indication that significant environmental diversity existed anywhere in the Panhandle area. Even in the Anadarko Basin, where thicknesses exceed 2,000 ft (610 m), the Ellenburger (Arbuckle) appears to represent shallow-water, subtidal to supratidal deposition. The unit does grade from being predominantly dolomite, in West Texas and the Panhandle area, into limestone in central Texas and southern Oklahoma (Barnes and others, 1959). This may indicate that a slight west to east freshening (decrease in salinity) of water existed during deposition. Folk (1959), however, pointed out that there is no evidence to suggest that this is related to any major change in bathymetry.

The Mississippian System

Deposits of apparent Mississippian age occur throughout large areas of the Texas Panhandle region (fig. 4). These rocks overlie the Ellenburger throughout most of the Dalhart and Hardeman Basins. In the Palo Duro, they overlie the Ellenburger or rest directly on Precambrian basement (fig. 8). Middle and Late Ordovician, Silurian, and Devonian rocks are, for the most part, only present in the Anadarko Basin (fig. 8). These Middle Paleozoic deposits were apparently removed from much of the Panhandle area by erosion during the Middle Devonian (Huffman, 1959; Amsden and others, 1967). Middle and Late Ordovician are present in the northern fringes of the Dalhart Basin and in the extreme eastern part of the Hardeman Basin (fig. 8).

Mississippian rocks reach thickesses of as much as 4,000 ft (1,220 m) in the Anadarko Basin, north of the Amarillo Uplift (fig. 9). South of the uplift area, greatest thicknesses are found in the Hollis and Hardeman Basins; as much as 1,400 ft (427 m) of Mississippian has been reported in this area (fig. 9). The Palo Duro and Dalhart Basins contain maximum thicknesses of about 900 ft (274 m).

The Mississippian System of North America comprises four series: Kinderhook, Osage, Meramec, and Chester (Dott, 1941; Cheney and others, 1945). It has been common practice in the Panhandle area to use these series to subdivide the Mississippian section (fig. 1). Because of the scarcity of biostratigraphic control, however, the recognition of these units in the subsurface of the Panhandle is primarily based on lithologic correlation to well-known outcrop sections in the region. Recognizing the questionable validity of the use of these series in the subsurface, Mapel and others (1979) used letter designations to subdivide the Mississippian in the area (fig. 10). In effect, these series designations are employed as rock-stratigraphic units (that is, Groups) in the subsurface of the Panhandle. Undercovering this fact is biostratigraphic evidence recently recovered from core in the Hardeman Basin area which indicates that the so-called Osage Series is actually Meramec in age (Ruppel, 1983, 1984).

Hardeman Basin

Because of numerous recent hydrocarbon discoveries, the Mississippian sequence in the Hardeman Basin has been more extensively studied than elsewhere in the Panhandle area. The basic stratigraphy is quite uniform throughout the Hardeman and Hollis Basins. The Chappel Formation forms the base of the sequence, resting directly on Ellenburger Group dolomites (fig. 5). It is sometimes subdivided into an upper Meramec part and a lower Osage part. The Chappel is overlain by oolitic units generally referred to the Ste. Genevieve and St. Louis Formations. There is, however, a great deal of disagreement over correlation of the St. Louis Formation in particular. For purposes of this study, the Chappel/St. Louis boundary has been placed at the base of the lowest shale bed above the highly resistive Chappel (fig. 5). Some workers extend the St. Louis below this point to include all beds of ooids found in the sequence.

The generally highly radioactive and highly resistive, brown to black shales and dark-colored limestones of the Barnett Formation form a persistent marker throughout the Ft. Worth Basin and most of North Texas. The distribution of the Barnett in the Panhandle area is, however, limited almost primarily to Hardeman County (fig. 11) where it reaches a maximum thickness of about 150 ft (46 m). Although generally correlative into Oklahoma, the unit appears to undergo a gradual northward facies change into lighter-colored shales. The Barnett grades westward into the Palo Duro Basin into sandstones (fig. 12) assigned to the Pennsylvanian. The Comyn Formation, which is predominantly composed of carbonate, forms the top of the Mississippian section in the area (fig. 5). The contact with the overlying Pennsylvanian is difficult to place. In the absence of biostratigraphic control its placement is largely arbitrary.

The Chappel Formation has received the greatest amount of study in the Hardeman Basin area because of the discovery of numerous hydrocarbon reservoirs in the unit. It is characterized by rapid lateral variations in lithology. Three basic depositional settings have been recognized comprising at least six lithofacies (Ross, 1982; Ahr and Ross, 1982): (1) relatively deep water, open marine (interbuildup) deposits composed of laminated, cherty spicular wackestones (Allison, 1979; Asquith, 1979), (2) carbonate buildups comprising both core (mudstones and wackestones) and flank (skeletal grainstones and packstones) facies, (3) ooid shoals composed of ooid/skeletal sands (Ahr and Ross, 1982). As indicated previously, the ooid facies is considered part of the St. Louis Formation by some. In any case, it seems clear that the Chappel Formation of the Hardeman Basin records the local development of carbonate buildups in a generally deeper water, open platform, marine setting that eventually shallowed into ooid sand shoals. Recent study of Chappel cores in the Hardeman Basin generally supports interpretations offered by previous workers.

Palo Duro and Dalhart Basins

The Mississippian stratigraphic section exhibits marked changes in thickness and lithology at the western edge of the Hardeman Basin (fig. 12). Because of this, correlation between the Palo Duro and Hardeman Basins is confusing. For example, although the Ste. Genevieve sequence in the Hardeman Basin can be easily traced westward, it is equivalent to the Chester Group in the Palo Duro Basin (fig. 12). These problems in correlation are apparently largely due to the timing of middle Carboniferous deformation in the area. The Mississippian/

Pennsylvanian contact in the Palo Duro Basin is generally defined at the base of the lowest "granite wash" deposits (arkosic sandstones) apparently derived from the erosion along the Amarillo Uplift. Such deposits are not found in the Hardeman Basin until much higher in the stratigraphic section.

Kinderhook Group.--The Kinderhook is largely restricted to the Anadarko Basin (fig. 13) where it is composed of light-colored, mostly fine-grained angular to subrounded, quartz sandstone. It is locally glauconitic and commonly interbedded with green to gray shale. In some cases, it also contains interbeds of light-colored limestone or dolomite, particularly near the periphery of its extent. Although the Kinderhook has been reported in several wells in the Palo Duro and Dalhart Basins, the presence of similar basal Mississippian sandstones in these areas is extremely limited. Such sands are present in a few wells in the extreme northeastern part of the Palo Duro Basin, immediately south of the Amarillo Uplift (fig. 13). This may indicate that these sandstones originally extended over much of the uplift area before being removed by erosion.

Although these sandstones are rare throughout the rest of the Palo Duro Basin, there are basal Mississippian shales present in many wells that may be temporally equivalent. These shales are common in all parts of the basin except in the area of the Texas Arch (fig. 14). However, since the equivalence of these shales to Kinderhook sandstones cannot be established, they are grouped with Osage rocks.

Totten (1956) and Allison (1979) reported Kinderhook-like deposits of sandstone (Misener Sand) and shale in the Hardeman Basin. These deposits, however, appear to be generally thin and only locally developed. Basal Mississippian deposits in this area are probably more properly assigned to the Osage.

Kinderhook rocks apparently represent basal transgressive sediments formed at the beginning of Mississippian deposition in the Panhandle area. Although the exact age of those deposits cannot be determined, relationships with the underlying Woodford Formation (Gutschick and Moreman, 1967; Amsden and others, 1967) indicate that they are early

Mississippian or younger. The distribution of the coarse clastics (fig. 13) suggests a possible source in the present area of the Amarillo Uplift (Gray and Carson Counties, Texas). The absence of basal Mississippian shales along the Texas Arch (fig. 14) indicates that this feature may have had some positive expression well into Mississippian time.

Osage Group.--Osage rocks are the most widespread of all the Mississippian units in the area (fig. 15). These deposits reach thicknesses of nearly 400 ft (122 m) in the Palo Duro Basin. Although the Osage is not easily recognized in the Hardeman/Hollis Basin area, thicknesses of about 400 ft (122 m) have been recorded from the western part of the area in eastern Childress County (fig. 15). The Osage in the Dalhart Basin is generally thinner, having a maximum thickness of only about 175 ft (53 m). Thicknesses of more than 1,000 ft (305 m) are encountered in the Anadarko Basin immediately north of the Amarillo Uplift (Wheeler and Hemphill Counties).

In the Palo Duro, Dalhart, and Anadarko Basins, Osage rocks are gray to brown, commonly argillaceous, cherty limestones and dolomites. Locally they contain significant amounts of gray to green shale. Glauconite and pyrite are minor accessories. Sample log data indicate a mosaic of lithofacies across the Texas Panhandle (fig. 16). In general, pure (shale-free) limestones are present only in the extreme eastern and northeastern parts of the Palo Duro Basin and into the Hardeman Basin to the east. The percent of shale and dolomite generally increases to the west in the Palo Duro Basin (fig. 16). The boundary between relatively pure limestones in the east and shalier dolomites and limestones in the west roughly corresponds to the erosional edge of the underlying Ellenburger Group (fig. 7). In the western and northwestern parts of the basin, the Osage is composed almost entirely of dolomite (fig. 16). Most of the Dalhart and western Anadarko Basins are similarly primarily dolomite.

In the Hardeman and Hollis Basins, the Osage is difficult to recognize. It is usually correlated with the lower part of the Chappel Formation (Allison, 1979; Asquith, 1979). The Chappel is essentially equivalent to the Osage and the Meramec of the rest of the Panhandle area (Ruppel, 1984).

Osage core is available from only the northeastern (Donley County) and extreme eastern (Childress County) parts of the Palo Duro Basin (fig. 2). In the Childress 10 core (fig. 17) the Osage (fig. 18) is composed primarily of alternating layers of brown, laminated wackestone, and skeletal, lime-silt grainstone that are locally silicified. The grainstones contain well-sorted skeletal debris, predominantly echinoderms and bryozoan fragments. In some cases, these grainstones display grading. The wackestones contain abundant laminations of skeletal debris similar to that comprising the grainstones. Relatively thick layers of coarser, skeletal grainstone are locally present in the section (fig. 18). These deposits are commonly contorted and contain numerous truncation surfaces. In some cases, they are heavily burrowed. These grainstones are commonly dolomitized. Also present, although not common, in the Childress 10 core are layers of sedimentary breccia. The nature of the clasts making up these breccias indicates that these deposits were formed by movement following partial lithification of the sediment. Thin layers of dark gray, spiculitic wackestone represent a relatively minor part of the Osage in the Childress 10 well. Silicification, however, seem to be associated with these layers suggesting that sponge spicules were the original source of the silica.

Osage deposits present in the Childress 10 core seem to indicate that relatively deep-water conditions extended from the Hardeman Basin at least as far west as Childress County. The spiculitic wackestones are quite similar to those interpreted as deep water, interbuildup deposits by Allison (1979) and Asquith (1979). These wackestones probably represent in situ deposits formed in a quiet water, perhaps below wave base, open platform setting. The bulk of the Osage in the Childress County area, however, is composed not of in-place deposits, but of transported skeletal debris (Ruppel, 1984). Although finer-grained and thus representing somewhat more distal deposits, these rocks are similar to the limestone breccias associated with carbonate buildups in the Hardeman Basin (Ross, 1981). This indicates that buildup growth may have extended at least as far west toward the Palo Duro Basin as Childress County.

Core from Donley County (fig. 2), immediately south of the Amarillo Uplift, contains a markedly different sequence of Osage lithologies. In the Donley 3 well, only the uppermost part of the Osage has been cored (fig. 19). These deposits comprise (1) alternating layers of red and green spiculitic dolosiltstone and red to green to gray medium- to coarse-grained, skeletal grainstone composed primarily of echinoderms and bryozoans. The dolosiltstones are locally burrowed or contain laminations of skeletal debris. Possible mudcrack structures are present at some horizons. Siliceous sponge spicules are common in the dolosiltstones.

The Osage rocks present in the Donley 3 well appear to characterize an alternation between normal subtidal deposition (grainstones) and supratidal or intertidal conditions (dolosiltstones). These deposits thus suggest at least local shallowing and emergence. Although evidence is equivocal, there are several indications of regional uplift at the end of Osage deposition (Mapel and others, 1979). A rapid change from deep water to shoaling conditions is also indicated at the Osage/Meramec contact in the Childress 10 core.

Due to the lack of core in the interior of the Palo Duro Basin, the exact depositional conditions under which Osage rocks formed in this area are difficult to determine. Regional relationships indicate a general east to west shallowing of environments during deposition of Osage rocks. This implies that the sediments in the Palo Duro Basin were probably formed in relatively shallow, inner platform conditions. Since the interior of the Palo Duro Basin contains predominantly dolomitic rocks, it is tempting to conclude that these deposits represent the shallowest areas of deposition. Such an interpretation cannot be proven without examination of core in the area.

Meramec Group.--These rocks are relatively consistent in thickness, ranging from about 300 ft (91 m) to 350 ft (107 m) except where partially removed by erosion (fig. 20). Although the Meramec and Osage cannot be readily distinguished in the Hardeman and Hollis Basins, an isopach of the entire interval does not reveal any consistent thickening trend in this area.

The top of the Meramec is generally easily recognized in the Panhandle area by a marked increase in resistivity and a gradual shift in SP (fig. 5). This horizon is clearly correlative with the top of the Chappel Formation (in the sense used herein) in the Hardeman Basin (figs. 5 and 12).

The Meramec generally comprises white to buff-colored, fine- to medium-grained limestone. Chert and ooids are locally abundant. Fine-grained quartz sandstone is common near the top of the unit in most wells. In many parts of the area, but particularly in the Dalhart and western Anadarko Basins, the Meramec is divided into three formations: the Ste. Genevieve at the top, the St. Louis, and the lower Spergen-Warsaw (Cunningham, 1969). The Ste. Genevieve is characterized by the presence of quartz sand; it is usually no more than 50 ft (15 m) thick. Although ooids may be present throughout the Meramec, in the St. Louis Formation they are particularly abundant. In the underlying Spergen-Warsaw ooids are much less common. Dark-colored, aphanitic limestone is locally common in the Spergen-Warsaw; fossils are less common than in overlying parts of the Meramec. Dolomite is locally present in the lower St. Louis, but becomes very common in the Spergen-Warsaw (Cunningham, 1969).

In core from the Childress 10 well (fig. 21), the Meramec comprises skeletal grainstones and minor wackestones and mudstones. Echinoderms and bryozoans (both ramose and fenestrate) dominate the fauna. The grainstones are laminated with carbonate mud near the base of the Meramec, but are purer and more massive upward.

Meramec rocks in the Childress 10 area represent the development of a carbonate, skeletal-sand shoal (Ruppel, 1984). Although no core is available for confirmation, sample logs suggest that these shoal facies extended westward throughout the Palo Duro Basin.

Core and sample data indicate that Meramec rocks reflect a general shallowing trend in the area. The presence of limestone and sandstone conglomerates, sandstones, and shales at the Meramec/Chester contact suggests that this shallowing culminated in erosion throughout much of the Panhandle area. The presence of abundant quartz sand in the upper part of the Meramec (Ste. Genevieve Formation) presages the further uplift and erosion that followed.

<u>Chester Group</u>.--Rocks assigned to the Chester are significantly more restricted in their areal extent than are underlying Meramec or Osage deposits (fig. 22) due to Late Mississippian/ Early Pennsylvanian erosion. They are confined to the central and eastern parts of the Palo Duro Basin and are present only along the margins of the Dalhart Basin. The maximum

thickness in the Palo Duro Basin is about 300 ft (91 m). By comparison, as much as 1,750 ft (533 m) of Chester has been reported in the Texas part of the Anadarko Basin (Cunningham, 1969).

In the Palo Duro Basin, the Chester Group is primarily composed of white to buff, fine-grained, fossiliferous, oolitic limestone. Fossils include echinoderms, brachiopods, and bryozoans. Chert is relatively rare. Commonly interbedded with these limestones are gray, green, red, and brown laminated, calcareous shales. Thin beds of light-colored, calcareous sandstone are locally present. These clastic deposits are most abundant in an elongate swath through the eastern part of the Palo Duro Basin (fig. 23). Clastic content is also higher in the west-central part of the basin (southwestern Swisher County) and along the Matador Arch (southern Motley County). The lowest shale contents are found in the center of the basin (fig. 23). This area of much purer carbonate is particularly obvious on geophysical logs (fig. 24). The amount of shale and sandstone in the Chester is generally significantly higher north of the Amarillo Uplift (fig. 23). Greatest amounts (nearly 100% shale and sandstone) are found in the northwestern corner of the Texas and Oklahoma Panhandles (Dallam County, Texas, and Texas and Cimarron Counties, Oklahoma).

The contact between the Chester and the underlying Meramec Group is sharp at most places in the Palo Duro and Dalhart Basins. The basal Chester is composed of limestone and quartz conglomerates and quartz sandstones throughout most of the central Palo Duro Basin (Donley, Briscoe, Hall, Floyd Counties). In the eastern (Cottle, Childress, Motley Counties) and west-central (Swisher, Hale Counties) parts of the basin, the contact is marked by basal Chester quartz sandstones and shales or is gradational. In the northern part of the Panhandle, the contact is sharp in the west/northwest and gradational in the east (Anadarko Basin). In the northwest corner of the Texas and Oklahoma Panhandles (Dallam and Hartley Counties, Texas, and Cimarron County, Oklahoma) limestone pebble conglomerates are ubiquitous at the base of the Chester.

Rocks of the Chester Group represent continued shallow-water marine deposition of ooid/skeletal sands. The abundance of clastics, however, contrasts with earlier Mississippian deposits and indicates that sources of terrigenous clastics developed during this time. The sharp contact between Chester and Meramec rocks throughout most of the area indicates that a period of erosion may have preceded Chester deposition. The presence of limestone pebble conglomerates locally in the Palo Duro Basin and commonly in the northwestern Panhandle area is evidence of at least local erosion of the underlying Meramec Group rocks. The distribution of basal Chester lithologies suggests that the uplift which accounted for this erosion was greatest in the northwest and least in the east and west. The apparent trend of the upwarped area was generally northwest-southeast. Interestingly, this trend follows that of the Texas Arch (fig. 7) quite closely.

The distribution of clastics throughout the Chester Group suggests that at least one source of these deposits was to the northwest. Lithofacies mapping in this area (Craig and Connor, 1979) suggests that this source may have been in Colorado. It is difficult to determine how many other sources of terrigenous clastics may have developed during the Chester. Northwest/southeast trending tongues of clastics in the eastern Palo Duro Basin could have been produced by erosion on early, uplifted blocks along the Amarillo Uplift in the Potter/Carson County area. Basal Pennsylvanian clastics also appear to be derived from a such a source (Dutton, 1980, fig. 14). Higher concentrations of terrigenous clastics are also noted in the Chester in southeastern Motley County/southwestern Cottle County (fig. 23), suggesting that the uplift may have occurred along the Matador Arch at this time as well. The development of an area of relatively pure Chester carbonates in the center of the basin may have been a function of distance from terrigenous sources. The distribution pattern of Chester lithofacies is like the Chester lithologies themselves: more similar to Pennsylvanian deposits and styles of deposition than to the underlying Mississippian. This suggests that the forces that acted to shape the area into basins and uplifts in the Pennsylvanian had already become active in the Chester.

Age of Mississippian Rocks

As indicated earlier, conodonts have been recovered from some of the available core in the area. Identifiable fauna were obtained from four cored wells: Donley 3, Childress 10, Hardeman 42, and Hardeman 44 (fig. 2). Childress 10 is perhaps the most significant among these cored wells, because it has long cores in both the Meramec and Osage Groups (fig. 17). Conodonts recovered from these cores reveal that both the Meramec and Osage intervals are Meramec in age. In fact, all conodonts recovered from the Childress 10 core represent middle Meramecian faunas. Even though conodonts were recovered from within about 80 ft (24 m) of the base of the Mississippian, no evidence of Osage or even early Meramec age faunas was recovered. This indicates that little or no Mississippian deposition occurred in this area until Meramec time. Although core available for biostratigraphic analysis in Hardeman County is largely from the upper Chappel Formation (Meramec Group equivalent) only, faunas from these cores are similar to those recovered at Childress 10.

Conodont faunas collected from the Donley 3 core (fig. 19) are significantly older than those observed in Childress and Hardeman Counties. These faunas suggest an early Meramec or even late Osage age for the uppermost part of the Mississippian section in this area. Since the Donley 3 well apparently contains about 200 ft (61 m) of Mississippian rock below the cored interval (fig. 19), it seems likely that true Osage age rocks are present in at least the northern part of the Palo Duro Basin. The age of the basal Mississippian section in the remainder of the basin, however, is not known because of the absence of core. Ruppel (1983, 1984) suggested that Osage rocks are confined to the northern edge of the basin and that most rocks south of the Amarillo-Wichita Uplift are Meramec or younger in age. A more comprehensive treatment of the conodont data and their implications is in preparation.

STRUCTURAL INFLUENCE ON DEPOSITION OF THE SAN ANDRES AND BLAINE FORMATIONS (MIDDLE PERMIAN), POTTER COUNTY

(M. A. Fracasso)

Introduction

The San Andres and partly equivalent Blaine Formations in the Texas Panhandle are composed of cyclic, predominantly carbonate and evaporite deposits with a minor terrigenous clastic component. These deposits have traditionally been considered to represent shallow shelf and supratidal (sabkha) sedimentation during a period of relative structural/tectonic quiescence. Recently, San Andres deposits have been reinterpreted to reflect a primarily subaqueous shallow shelf depositional system (Fracasso and Hovorka, 1984). The very shallow-water, broad, low slope and low relief depositional shelf promoted the development of regional salinity gradients that controlled the distribution and thickness of San Andres facies (Fracasso and Hovorka, 1984). Even minor paleotopographic features present in such systems are expected to have produced sharp local salinity gradients that resulted in anomalous local patterns of facies distribution and thickness (Fracasso, 1984a). Greimel (in press) has presented evidence for lcoal structural control over San Andres sedimentation in the Castro Trough area of the Palo Duro Basin. Fracasso and Hovorka (1984) and Fracasso (1983, 1984a, 1984b) have proposed that vertical change in patterns of San Andres facies distribution and thickness over the Palo Duro Basin area reflect systematic changes in the levels of regional and local basin subsidence rates during deposition.

This report describes patterns of lateral thickness change in a vertical succession of non-salt facies packages of the equivalent San Andres and Blaine Formations in Potter County, Texas Panhandle. Potter County encompasses portions of both the Palo Duro Basin and its northeastern border--the Amarillo-Wichita Uplift (fig. 25). The study is presently limited to non-salt facies units so that complications associated with the possibility of postdepositional salt dissolution over the structurally high basin margin are minimized (Fracasso, 1984a).

Closed-contour thickness anomalies at this areal scale probably reflect syndepositional salinity gradients controlled by local paleotopographic features, rather than regional, basin-wide salinity gradients.

Some correspondence exists between areas of repeated thickness anomalies and underlying basement structures with great relief. This relationship suggests that the expression of minor syndepositional topographic features was in many instances controlled by active differential subsidence over preexisting faults at depth.

Methods

Isopach maps were prepared for a vertical succession of non-salt facies units in the San Andres/Blaine Formations in Potter County and vicinity (figs. 26-32). An ideal San Andres genetic cycle, from base-to-top, is composed of black mudstone, carbonate, anhydrite, and halite facies (Hovorka, 1983; Fracasso and Hovorka, 1984; Hovorka, in press). The non-salt facies thickness comprises the total thickness of the mudstone, carbonate, and anhydrite facies in each unit. The San Andres Formation in the Palo Duro Basin has been divided into three informal genetic sequences (Fracasso, in press), each composed of a number of informal operational units (fig. 33). Operational units are the thinnest stratigraphic intervals that can be correlated and mapped across the entire basin using geophysical logs (Fracasso and Hovorka, 1984). Each operational unit may comprise one or several genetic cycles, depending on their thickness and resolution on geophysical logs.

A primary goal of this report is to demonstrate and precisely delineate areas in Potter County that contain repeated thickness anomalies in the San Andres/Blaine Formation. A grid of labeled squares was superimposed over each isopach map; squares were recorded when approximately half or greater of their area covered a closed-contour thickness anomaly. Grid squares arbitrarily scaled to dimensions of 1.5 mi x 1.5 mi proved to be convenient; these were sometimes divided into quarters (0.75 mi x 0.75 mi) when small anomalies required greater resolution. In this manner, it is possible to tabulate the total number of times each labeled square occurs over thickness anomalies among all the units. This operation was performed separately for closed-contour thick and thin areas. These values were plotted as points in the centers of the squares and then contoured. The resultant map (fig. 34) illustrates the loci of recurrent thickness anomalies in the San Andres/Blaine Formation in Potter County. The contoured number of occurrences provides a rough index of the level of confidence that an enclosed area represents a real locus of persistent non-salt facies thickness anomalies, rather than bad data. I have arbitrarily chosen the value of three occurrences as a thickness anomaly among the seven studied units as the lower limit for contouring. The resultant map of San Andres/Blaine persistent thickness anomalies (fig. 34) can be compared with a structure contour map covering the same area (fig. 35) to evaluate the possibility of correspondence between known structures and areas of recurrent thickness anomalies.

Discussion

Figures 26-32 are isopach maps of a vertical succession of non-salt facies units of the San Andres/Blaine Formation in Potter County, Texas. Each map shows an increasing geometric complexity to the north, as indicated by increases in the overall density of contour lines and the number of closed-contour areas. These northern zones of complex geometry coincide with the underlying Amarillo-Wichita Uplift belt, differing markedly from the broader, more regular trends of thickness change evident in southern Potter County that are characteristic of the interior of the Palo Duro Basin.

Several loci of recurrent thickness anomalies have been identified (fig. 34) using the method outlined above. Areas of persistent thinning are more prominent than areas of persistent thickening, relative to both area covered and number of facies units affected. Figure 35 depicts these thickness anomalies superimposed on a structure-contour map based on the top of the underlying Tubb sandstone (Permian, Leonardian). Several correlations of San Andres/Blaine thickness anomalies with Tubb structures are apparent. Thin zones A and C (fig. 34) are situated over prominent structural highs, the X-L Dome and eastern arm of the

John Ray Dome, respectively. Thin zone B (fig. 34) is centered just southeast of the John Ray Dome, and partially overlaps its southeastern margin. Thick zone I lies over a marked depression on the John Ray Dome, and thick zones III and IV are situated over the downthrown sides of prominent faults. Thick zone II overlies the John Ray Dome (figs. 34 and 35). Zones of persistent thinning in the San Andres/Blaine tend to overlie structural highs, whereas persistent thicks overlie structural lows, with only two exceptions--thin zone B and thick zone II (figs. 34 and 35).

These persistent thickness anomalies cannot be simply explained by postdepositional differential compaction over underlying high and low structures. If this were the case, then all units overlying a given structure should be similarly affected--yet they are not. Likewise, syndepositional, continuous, passive differential subsidence over inactive older structures (the "post-tectonic adjustments" of Johnson, 1978, p. 57), and passive infill of static depositional relief (inherited from an earlier episode of structural activity) are both unlikely explanations of the persistent thickness anomalies. In either case, all the units overlying a given structure should exhibit similar thickness anomalies. Moreover, successive units would be expected to show gradually decreasing effects up-section, as either the structural stresses dissipate with time, or the relief fills respectively. Neither of these situations are observed in the San Andres/Blaine stratigraphic succession. The thicknesses of units overlying a given structure are never all affected similarly, nor is there evidence of a vertical trend toward decreasing number of thickness anomalies, or reduction of their areal extent.

Based on the irregular pattern of vertical recurrence of thickness anomalies over given loci, it appears that their development was controlled by intermittent minor changes in local rates of subsidence across underlying structures during deposition. These local differential subsidence rates probably originated in response to low amplitude changes in externally-controlled regional or local stress fields that episodically reactivated fault-bounded structures at depth.

Differential subsidence over buried structures may have controlled the accumulation of overlying sediments by three mechansisms, which may have acted alone or in combination. These mechanisms are: 1. control of effective sediment aggradation rates, 2. control of local water volume by development of topographic relief, and 3. control of sediment transport and accumulation by development of topographic relief. Sediment aggradation may have controlled the thickness and vertical sequence of facies in subaqueous, shallow-shelf carbonate-evaporite depositional systems (Fracasso, 1984a; Fracasso and Hovorka, 1984). If average rates of sedimentation for equivalent facies are the same over broad areas, then the effective aggradation rates will be greater over structural highs that subside less rapidly than adjacent areas. These greater effective rates of sediment accumulation would produce localized zones of shallower water and restricted circulation over the structural highs. The salinity gradients thereby induced would have inhibited non-halite facies sedimentation and promoted halite deposition over the structural highs, resulting in thinner non-halite facies relative to the adjacent, more rapidly subsiding structural lows. This mechanism may have been particularly effective in areas such as Potter County, that were far removed from the southerly normal-marine water source. In the most distal areas, the regional shallow-shelf circulation system probably operated at or near its threshold capacity to maintain near-normal marine salinities for even the short, initial transgressive phase of each cycle. Minor changes of water depth in such environments would have had profound effects on local circulation, brine salinity, and resultant facies thickness.

Differential subsidence of high amplitude over short time spans may have produced a variable local depositional topography, with structural lows overlain by topographic depressions, and structural highs overlain by topographic rises. The large volumes of water in topographic depressions may be expected to produce thick non-salt facies accumulations for three reasons. The prolonged maintenance of near-normal marine salinities in large-volume water bodies would support populations of skeletal organisms for longer times, resulting in thicker carbonate accumulations. These extended periods of near-normal marine salinities may also have allowed enhanced dissolution of underlying halite, producing thicker basal siliciclastic black mudstone insoluble residues (Fracasso, 1984a). More fundamentally, evaporation of larger volumes of marine brine should produce proportionately greater thicknesses of all chemically precipitated facies, including portions of the carbonate, and the anhydrite components of the non-halite facies units.

In the subaqueous, very shallow-water, shelf depositional system proposed for the San Andres/Blaine Formation (Fracasso and Hovorka, 1984), most of the shelf sediments were probably deposited above effective wave base. This renders it quite likely that the sediments were periodically current reworked. Given the existence of some structurally controlled depositional topographic relief, the net effect of current agitation would have been to winnow sediments from the higher energy topographic rises, and deposit them in adjacent, lower energy depressions. This would result in thicker sediment accumulations over structural lows, and thinner accumulations over structural highs.

Although the effects of these mechanisms for controlling the thickness of non-halite facies deposits cannot be differentiated at present, I suspect that local depositional relief was never great enough for difference in local water volume (mechanism 2) to have been a significant factor. Therefore, the most likely factors that influenced the thickness of non-halite facies units are variable effective sediment aggradation rates (mechanism 1), and sediment transport in areas of local depositional relief (mechanism 3), both controlled by intermittent differential subsidence across underlying, fault-bounded structures.

The overall trend of increasingly complex non-halite facies geometry to the north in Potter County is present in all units except unit 3 of the lower San Andres, which pinches out along the southern border of the Amarillo-Wichita Uplift. There is apparently no well-defined segregation of successive units with similar geometries into packages corresponding to the lower, middle and upper San Andres genetic sequences established in the basin interior (Fracasso, in press; Fracasso and Hovorka, 1984). This suggests that the Amarillo-Wichita Uplift belt was persistently structurally active throughout San Andres/Blaine Formation deposition, and does not display the episodic pattern of structural activity characteristic of the Palo Duro Basin interior.

CORE ANALYSIS OF THE DOCKUM GROUP, DOE-GRUY FEDERAL #1 GRABBE, SWISHER COUNTY

(D. A. Johns)

Introduction

The DOE-Gruy Federal #1 Grabbe well in northeastern Swisher County penetrated and cored the entire Triassic interval represented by the Dockum Group. The interval is 553 ft thick in this well from which 405 ft of core was recovered (73% recovery) (fig. 36). Lithologies identified in core include sandstone, siltstone, mudstone, conglomerate, and claystone. Sandstone is the dominant lithology (47%). Medium to coarse sand is commonly found in the lower part of the Dockum while fine to very fine sand dominates the middle and upper sections. Siltstone is the second most abundant lithology (24%) and is evenly distributed. Mudstone (18%) occurs throughout the section, but almost half of the total is in a thick bed in the upper part of the Dockum core. Pebble and granule conglomerate (6%) are most abundant and thickest in the middle of the formation. Most of the claystone lithology (4%) is found in a single bed in the lower Dockum section.

In Swisher County, the Dockum Group overlies the Permian Dewey Lake Formation and underlies the Tertiary Ogallala Formation. Although the upper contact was not recovered, the first Ogallala sediments, light brown, calcareous, poorly sorted sandstone with opal-filled fractures, contrast sharply with the orange-red, ripple-laminated, mudstones and siltstones at the top of the recovered Dockum. On geophysical logs this contact is characterized by a sharp drop in gamma-ray response and typically an increase in resistivity and sonic travel time. The basal contact is not as easily delineated. For the Grabbe well, the contact is based primarily on geophysical log correlations within the Palo Duro Basin and supported by lithologic changes in the core. The contact is at the base of a red-brown, bimodal, faintly ripple-laminated, very fine-grained sandstone containing pervasive pedogenic structures. Unfortunately, there was no core recovery for the next 18 ft until red-brown, parallel to ripple-laminated mudstones and siltstones of the Dewey Lake were encountered, so the nature of the basal contact is unknown. The contact shows a sharp drop in gamma-ray response compared to the underlying sediments.

Lithologic Packages

For the purposes of this discussion the Dockum is divided into five separate vertical sequences. These sequences encompass sediments of either similar lithology, lithologic associations or depositional style, and the boundaries mark what are considered to be significant changes in lithology and depositional environment.

First Sequence

Sixty-five feet of red-brown, very fine-grained sandstone and siltstone with minor claystone characterized by fine grain size, soft sediment and pedogenically disrupted fabrics, and small-scale ripple laminations make up the lower sequence (from 780 to 715 ft). Three significant lithofacies are recognized from bottom to top: sandstone, interbedded sandstone and siltstone with mudstone, and micaceous muddy siltstone.

The lower 27 ft of the first sequence is a very fine sandstone exhibiting pedogenic structures (G. A. Kocurek, personal communication, 1984; Ahlbrandt and Fryberger, 1981) in the form of irregular, discontinuous 1 to 2 mm thick bands of clay illuviation overprinting subtle small-scale ripple laminations, burrows and small scours. Oxidized, heavy mineral placers define ripple forms. Well-rounded, frosted, medium to coarse sand grains, similar to those characteristic of the Dewey Lake and older Permian units, form lag deposits at 769 ft and are sparsely scattered throughout the interval. Bedding is indistinct and appears massive. Small angular, thin mud clasts are common.

The sandstone grades upward into parallel laminated, interbedded sandstone and mudstone with some climbing ripples and fluid escape structures capped by about two feet of clayey mudstone and siltstone in which bedding is very contorted, probably by soft sediment processes.

The top of the first sequence consists of 38 ft of red-brown, micaceous, muddy siltstone which becomes less muddy upward. The lower part is ripple- and parallel-laminated, contains rip-up clasts, thin zones of contorted bedding, water escape structures, burrows, and desiccation cracks. The upper portion displays contorted bedding at the base, grading upward into parallel and finely ripple-laminated siltstone. Micas are observed on bedding planes near the top. Thin, very light gray, sand-rich scours with coarse sand lags are more numerous upwards.

Second Sequence

Two distinctly different sandstones and a mudstone/claystone top make up the 79-ft thick second lithologic package (715 to 636 ft). Missing core separates the sandstones (660 to 697 ft) and also the contact with the first sequence.

The lower sandstone is pale reddish brown, (10R 5/4), poorly consolidated, poorly sorted, silty, coarse sandstone grading up to better sorted medium sandstone. Structures are large-scale crossbedding and horizontal lamination; plant debris and mud clasts up to 5 cm across are common on bedding planes.

The upper sandstone is poorly sorted, coarse- to medium-grained with maroon and white color banding parallel to bedding. It is crossbedded with some scour surfaces marked by coarser-grained sand. This rapidly grades into a dark brown-purple, intensely burrowed mudstone/claystone with scattered small reduction spots. No fossils have been found.

Third Sequence

Four significant lithofacies make up the 223 ft (636 to 413 ft) thick third sequence: the basal siltstone, conglomerate, sandstone, and siltstone/mudstone. The interval contains most of the conglomerates and the thickest sandstones found in the core.

Basal Siltstone

The thin (9 ft thick) basal section has a scoured base of caliche pebble (0.5 cm) conglomerate, 2.5 inches thick, grading into brown ripple-laminated siltstone with ochre-colored bands parallel to bedding. Clay content increases upward whereas ripple size decreases.

Conglomerates

Dockum conglomerates, up to 4.5 ft thick, consists of granules, pebbles or cobbles of mudstone, siltstone, or carbonate (caliche), and, rarely, plant debris. They may be composed of mixtures of all the above rock types or exclusively one rock type, depending on the substrata on which the conglomerate developed. Those composed entirely of carbonate rock fragments commonly show stylolite development parallel to bedding. Larger rock fragments, up to 5.5 cm across, are typically flat or elongate, while smaller ones are more spherical. Imbrication of clasts is common.

Conglomerates have three modes of occurrence: (1) at the base of sandstones, (2) within sandstones, or (3) as isolated lenses in mudstones or siltstones. Those at the base of sandstones are typically massive, clast-supported units, while the others are typically stratified, matrix-supported conglomerates. Graded bedding is common, mostly in one or more fining-upward sequences. However, inverse grading of thin beds was found at 625, 509, and 488 ft.

Sandstones

Sandstone is the major lithofacies of the third sequence. These deposits are reduced, generally calcite cemented, highly micaceous (a characteristic of Dockum sandstones) and contain small clay rip-up clasts, plant debris and clay drapes along bedding planes. Coarse- and medium-grained sandstones are in the lower half of the interval, in beds generally less than 10 ft thick, while fine-to-very-fine-grained sandstones occur in the upper half in beds up to 27 ft thick. The base of sandstone beds, when visible, is typically scoured with a lag of mudstone, siltstone, caliche clasts, or coarser sand grains. In the lower parts of sandstones, bedding-types are massive; horizontal stratification, tabular and trough crossbeds grade up into ripple and climbing ripple laminations. Load structures are common where sandstones directly overlie mudstones or siltstones.

Siltstones and Mudstones

All the sandstones are overlain by siltstones or mudstones from four- to nine-feet thick. Contacts are gradational, but abrupt associations are known (575, 480, and 457 ft). Siltstones are gray-green in color or rarely brownish-white, whereas mudstones are dark brown to red brown with ochre-colored laminae. In some cases, mudstones have a reduced gray-green zone where adjacent to sandstone beds. Climbing-ripple or low-angle, ripple lamination, rare large ripples, and some sand or mud laminae are found in siltstone. Many have faulting, load casts, and soft sediment folds. A siltstone at 477 ft contains a scour 8 cm deep into which clasts of the surrounding, still soft sediment collapsed and deformed. Mudstones exhibit low-angle to parallel lamination with burrows and possible root traces. Desiccation cracks, some resulting in brecciation, calcite nodules, and "streaks" cross-cutting and parallel to bedding are also associated with mudstones.

Fourth Sequence

Massive calcite-cemented mudstone 44 ft thick (414 to 370 ft) makes up the fourth interval. The base is gradational with the underlying siltstone; the top is relatively sharp but still transitional with the overlying siltstone. Color lightens upward from dark red-brown to pale red-brown. No original sedimentary structures have been identified, but the poor condition of the core hampers recognition of such features. Small calcite nodules (approximately 1 cm) are common throughout the sequence.

The base of the mudstone is massive; gray-green, calcitic, rounded mudstone intraclasts are present as fracture fills. Neither the tops of the fractures or zones of mud clasts, the source for those filling the fractures, have been found. The lower interval grades into mudstone with pervasive, whitish, calcite-rich veinlets and abundant, yellow-brown, circular (spherical?) spots. The veinlets are similarly oriented and become increasingly common in the same direction, but it is impossible to determine if this is true throughout this interval, because the core is broken. The overlying interval, 18 ft thick, is similar to the bottom interval, but is lighter in color and has scattered large reduction spots. The upper 12 ft of the mudstone is massively calcite-cemented and continues a nodular, brecciated fabric and numerous fractures and burrows filled with finer-grained, calcitic mudstone and mudstone intraclasts.

Fifth Sequence

Two conglomerate-sandstone-siltstone sequences and a basal siltstone characterize the 83-ft thick fifth sequence (370 to 287 ft). The top 60 ft of the Dockum was not recovered. The gradational base of the sequence is marked by increasing silt and mica content.

The base of this sequence consists of red-brown, ripple- and cross-laminated siltstone with coarse caliche lag in some ripples and scours. It is abruptly overlain by 3 ft of graded, crossbedded, imbricated caliche and mudstone intraclast conglomerate having a fine sand matrix. This grades into gray-green, ripple and climbing ripple-laminated, fine sandstone with mud-rich interbeds and abundant, soft sediment folds and faults. The overlying red-brown, cross-bedded, ripple and climbing ripple-laminated, coarse siltstone with soft sediment deformation structures is gradational with the sandstone. Unusual purplish-red reduction zones having lighter colored (more reduced?) centers are common parallel to and cross-cutting bedding. Calcite cement appears concentrated in the reduced zones. At 333 ft, a fracture containing and surrounded by calcite nodules crosses but does not otherwise affect one of the reduction zones.

In erosional contact with the siltstone is three feet of clast- and-matrix-supported caliche, siltstone, and mudstone intraclast conglomerate that rapidly grades into dark gray-green, horizontal and cross-laminated, very micaceous, fine sandstone with small clay clasts along bedding planes. This becomes climbing ripple-laminated toward the top. The top of the recovered Dockum is brecciated, ripple-laminated siltstone grading into mudstone with sandstone lenses.

Depositional Interpretations

First Sequence

The lower part of the first sequence is interpreted to represent eolian flat deposition grading upward into a mudflat or possibly marginal lacustrine setting. Pettijohn and others (1973) and Folk (1974) characterized eolian sediments as (1) well sorted, (2) lacking micaceous minerals, 3) having a dull opaque or matte surface on quartz grains, and 4) bimodal. No distinctly eolian ripple forms (Kocurek and Dott, 1981) were identified, but the homogeneous, fine grain size, presence of bimodal lags containing well-rounded, frosted grains, and the influence of pedogenic processes suggest that eolian flat processes were important in the basal section.

Noncharacteristic ripples, small scours, and small, thin, angular, mudstone intraclasts in the first sequence indicate aqueous deposition, probably by sheet sand flow which in desert environments produces ripple laminated sand covered by a veneer of mud later reworked into mudclasts (Sneh, 1983; Smoot, 1983). Similar features have been found by Hubert and Hyde (1982) in semiarid sheet-flow deposits in the Upper Triassic red beds of Nova Scotia. A marginal lacustrine environment is suggested by interbedded sandstone and mudstone containing fluid escape and soft sediment deformation structures similar to those present in the Eocene Wilkins Peak Member of the Green River Formation (Smoot, 1983). The muddy siltstone exhibiting ripple and parallel lamination, burrows, and desiccation cracks in the upper part of the first sequence probably represents shallow lacustrine deposition (Smoot, 1983). The siltstone at the top of the sequence is interpreted as a delta-front deposit because it overlies more muddy siltstone and has contorted bedding, ripple lamination, and contains micas on bedding planes suggesting fluvial input. The numerous sand-rich lenses with scoured bases in this sequence are probably frontal splays.

Second Sequence

The two sandstones of the second sequence are of fluvial origin; they are characterized by fining-upward grain sizes, horizontal and inclined cross-stratification, and mud clasts and plant debris on bedding planes. Similar features are found in modern fluvial channel environments (Harms and Fahnestock, 1965; Bernard and others, 1970; McGowen and Garner, 1975). The upper sandstone fines upward into mudstone and claystone interpreted as lacustrine based on their dark brown color, the presence of numerous burrows and their general similarity to units described in the Dockum by McGowen and others (1979) and Seni (personal communication).

Third Sequence

Sediments of the third sequence were deposited in deltaic, fluvial, and interdeltaic or overbank environments. The lower conglomerate and ripple-laminated siltstone (636 to 628 ft) are delta-front sediments deposited by a prograding river or stream system. This is indicated by the grain size, that is coarser than the underlying lacustrine rocks, and by similar relationships noted by McGowen and others (1979). Clast-supported conglomerates are interpreted as channel lag/fill or basal crevasse splay deposits based on their scoured bases and gradational relationships with overlying sandstones. The stratified, matrix-supported conglomerates are interpreted as channel bar or small channel fill deposits primarily based on their occurrence within sandstone beds. These are similar to those described by McGowen and Garner (1975) in coarse-grained point bars. In the lower part of the sequence, sandstones are interpreted to be delta-front or distributary channel sediments because of their generally thin beds, soft sediment deformation structures and primary fluvial sedimentary structures. The thick, fining upward, cross-stratified to climbing ripple-laminated sandstones in the upper part of the sequence are believed to be fluvial channel-fill and channel-bar deposits similar to those described in the second sequence. Descriptions of upper channel bar and crevasse deposits by Coleman and Gagliano (1965), Bernard and others (1970), Visher (1972) and McGowen and Garner (1975) strongly resemble those of the third sequence. These are recognized by their relationship with underlying sandstones, fine grain sizes, climbing ripple lamination, sand and mud laminae, intraformational scours and soft sediment deformation structures. Because of their position in the section, some siltsones in the lower part of the sequence may be delta-front deposits. Overbank and interdeltaic pond environments are interpreted for thin, finely ripple-laminated to parallel-laminated burrowed mudstones with caliche nodules based on similar deposits of the Brazos River (Bernard and others, 1970).

Fourth Sequence

The mudstones of the fourth sequence strongly resemble the dolomitic mudstones described by Smoot (1983) in the Eocene Green River Formation. Smoot interpreted the mudstones to have been deposited by sheet floods on a subareally exposed mudflat which underwent varying degrees of reworking by floodwaters. Identical conditions are believed responsible for the Triassic mudstones of the fourth sequence. Subareal exposure is suggested by calichification of the mudstone, mud cracks, and the desiccation breccia at the top of the interval which destroyed original sedimentary structures. Local reworking by floodwaters or lake waters is indicated by rounded and angular calcitic mudstone intraclasts deposited in desiccation fractures. The transitional base of this sequence suggests a gradual increase in lake size and the resulting thickness (over 40 ft) suggests a prolonged lacustrine setting.

Fifth Sequence

Siltstone at the base of the fifth sequence is interpreted to be a delta front deposit due to its position overlying lacustrine mudstone and its ripple- and cross-lamination. The two fining-upward sequences of conglomerate-sandstone-siltstone exhibit structures characteristic of fluvial environments similar to those previously discussed: horizontal and cross stratification to ripple lamination upward. The overlying siltstones and mudstones are believed to be interdeltaic deposits because of their relation to the underlying fluvial deposits but the lack of core recovery above these rocks makes interpretation difficult.

Conclusions

The core from the DOE-Gruy Federal #1 Grabbe exhibits classic Triassic Dockum Group sediments. These deposits record a gradual shift in climate and depositional style from arid or semiarid, eolian deposits of the Upper Permian and Lower Triasic to more humid lacustrine and fluvial/deltaic deposits of the Upper Traissic. Three major lacustrine-deltaic-fluvial-lacustrine cycles are noted: (1) top part of sequence 1 to top part of sequence 2, (2) top of sequence 2 to top of sequence 3, and (3) bottom of sequence 4 to top of sequence 5. Fine-grained lacustrine sediments exhibit features indicating: (1) undisturbed, subaqueous deposition--fine parallel lamination, (2) disturbed, subaqueous deposition--bioturbation, (3) periods of prolonged subareal exposure--calichification and brecciation. Most lacustrine intervals show evidence of prograding deltaic clastics by increasing grain sizes and amplitude of sedimentary structures upward. Fluvial sediments typical of the Dockum Group sediments exhibit classic fining-upward grain sizes and decreasing amplitude of sedimentary structures.

EVIDENCE FOR SYNDEPOSITIONAL AND POSTDEPOSITIONAL STRUCTURAL CONTROLS ON THE SEDIMENTS OF THE DOCKUM GROUP, PALO DURO BASIN

(D. A. Johns)

Structural elements in the Palo Duro Basin influence the accumulation, distribution, and trends of Dockum clastic sediments. The total isopach map (fig. 37) reflects basin configuration by thickening to the southwest, the deepest portion of the basin according to the base of Dockum structure map (fig. 38). The isopach map also shows sediment thicks and thins which correspond to structural lows and highs, particularly in Lamb County. These are not believed to be artifacts of preservation because the same correlation is seen on an isopach of the lower Dockum (fig. 39), which is not affected by later erosion in Lamb County and because similar structure is found on the basement (R. T. Budnik, personal communication). In addition, percent sandstone shows the structural highs to be generally sand poor (figs. 40 and 41), suggesting they were topographically high and so collected less sand than surrounding areas. Similar sandstone trends for the upper and lower Dockum (figs. 40 and 41) show an easterly sediment source for the Palo Duro Basin. However, the upper Dockum appears to have a western component exhibited by a band of 10 to 20 percent sandstone along the western boundary of the area (fig. 41) which has no obvious eastern source. This is suggesting a gradual westward shift of source areas for Dockum sediments.

Evidence indicating postdepositional structural modification of Dockum strata can be found on the subsurface maps. The structure and total isopach maps (figs. 37 and 38) show anomalously thick sediment preserved in structurally low areas along the Amarillo Uplift while sandstone trends (fig. 40) suggest some of these areas were lows during deposition, their unusual thickness and corresponding lows and sediment thicks in the overlying Ogallala Formation (Seni, 1980) indicate downwarping continued following deposition. In addition, most sandstone trends in the basin appear to be dip-oriented systems (figs. 40 and 41) but trends in Deaf Smith, Randall and Armstrong Counties appear strike oriented when compared to the structure on the base of the Dockum (fig. 38). Deposits from the northern edge of this system were cored in the DOE-Stone and Webster #1 J. Friemel well and tentatively interpreted to be primarily of fluvial origin (Stratigraphy CSR, 1983), which would occur in dip-oriented depositional systems. It is believed that uplift following deposition of Mesozoic strata (Budnik, personal communication; McGookey, personal communication) has sufficiently modified original basin geometry to give these sandstone trends their apparent strike orientation.

Further evidence for uplift is deduced from the structure contour map on top of the Dockum Group (fig. 42). It is important to remember that the top of the Dockum has been truncated by erosion. The structure contours trend north/northeast. The trend is the same for parts of the Dockum overlain by Cretaceous units as it is for that overlain by the Mio-Pliocene Ogallala Formation. This trend also parallels that of the present land surface. This implies that west-to-east gradient, similar to that existing today, was developed on the top of the Dockum prior to deposition of the Cretaceous units. The shift toward a western sediment source may have been the precursor of later uplift and tilting of the Dockum depositional basin.

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FIGURE CAPTIONS

Figure 1. Stratigraphic column of the Texas Panhandle area.

Figure 2. Locality map showing lines of section and core wells.

Figure 3. Structure on top of the Ellenburger Group (Lower Ordovician) in the Texas Panhandle area. Faults are interpretive based, for the most part, on well control only.

Figure 4. Structure on top of the Mississippian System in the Texas Panhandle area. Faults are interpretive based, for the most part, on well control only.

Figure 5. Typical pre-Pennsylvanian sections in the Dalhart, Palo Duro, and Hardeman Basin. Lithologies based on analysis of cuttings.

Figure 6. Thickness of Cambrian (?) basal clastic deposits in the Texas Panhandle area.

Figure 7. Thickness of the Ellenburger Group (Lower Ordovician) in the Texas Panhandle area.

Figure 8. Mississippian subcrop, Texas Panhandle area. Note that post-Ellenburger, pre-Mississippian strata are confined to the Anadarko and Midland Basins.

Figure 9. Thickness of Mississippian System in the Texas Panhandle area.

Figure 10. Common schemes of stratigraphic nomenclature applied to Mississippian rocks in the Texas Panhandle area. Meramec formations used in the Dalhart and Anadarko Basins are, in some cases, extended into the Palo Duro Basin with varying degrees of accuracy. The Ste. Genevieve is probably most consistent in its application. Usage in the Hardeman Basin varies considerably. Some workers place the St. Louis at the top of the Meramec; others consider the top of the Chappel as the top of the Meramec.

Figure 11. Thickness and distribution of the Barnett Formation in the Texas Panhandle area. The unit thickens eastward into the northern Ft. Worth Basin area (Montague County, Texas) to greater than 1,000 ft (305 m) (Henry, 1982).

Figure 12. East-west cross section of pre-Pennsylvanian strata through the Palo Duro and Hardeman Basins. Location of line of section is indicated in figure 1. Note particularly the facies changes that occur in Childress County, especially in the upper part of the Mississippian

sequence. This indicates that the present structural boundary between the two basins generally parallels paleoenvironmental boundaries. Note also that the top of the Missipssippian (top Chester) in the Palo Duro Basin correlates with top of the Ste. Genevieve in the Hardeman Basin. This discrepancy is in part due to differing terminologies developed by workers in the two basins, but also reflects the facies (conformable) nature at the "Mississippian-Pennsylvanian" boundary in the area. Lithologies are based on analysis of well cuttings except for Hardeman 42 for which core was also studied.

Figure 13. Thickness of basal Mississippian sandstones (Kinderhook deposits) in the Texas Panhandle.

Figure 14. Distribution of basal Mississippian shales (Kinderhook?) in the Palo Duro Basin area. Note the absence of these deposits along the Texas Arch (where Ellenburger deposits have been removed).

Figure 15. Thickness of the Osage Group. The Osage is not easily recognized in the Hardeman Basin.

Figure 16. Lithofacies map of the Osage Group. Data from sample logs.

Figure 17. Mississippian section and core intervals in the Childress 10 well (location shown in figure 2). Mississippian Group designations reflect common terminology applied to the section in the area. Conodont zones are based on identification of fossils from cored intervals. These data indicate that both the Meramec and Osage in the Childress 10 well are Meramecian in age.

Figure 18. Core analysis of the Osage Group (lower Chappel) in the Childress 10 well.

Figure 19. Pre-Pennsylvanian section and cored interval in the Donley 3 well (location shown in figure 2). Conodonts recovered from this core suggest a late Osage or early Meramec age for these deposits.

Figure 20. Thickness of the Meramec Group in the Texas Panhandle area.

Figure 21. Core analysis of the Meramec Group in the Childress 10 well (location shown in figure 2).

Figure 22. Thickness of Chester Group in the Texas Panhandle area.

hatched areas denote persistent thicks. Tubb sandstone structure-contour map is unpublished data generously provided by Roy T. Budnik, Bureau of Economic Geology, The University of Texas at Austin, 1984.

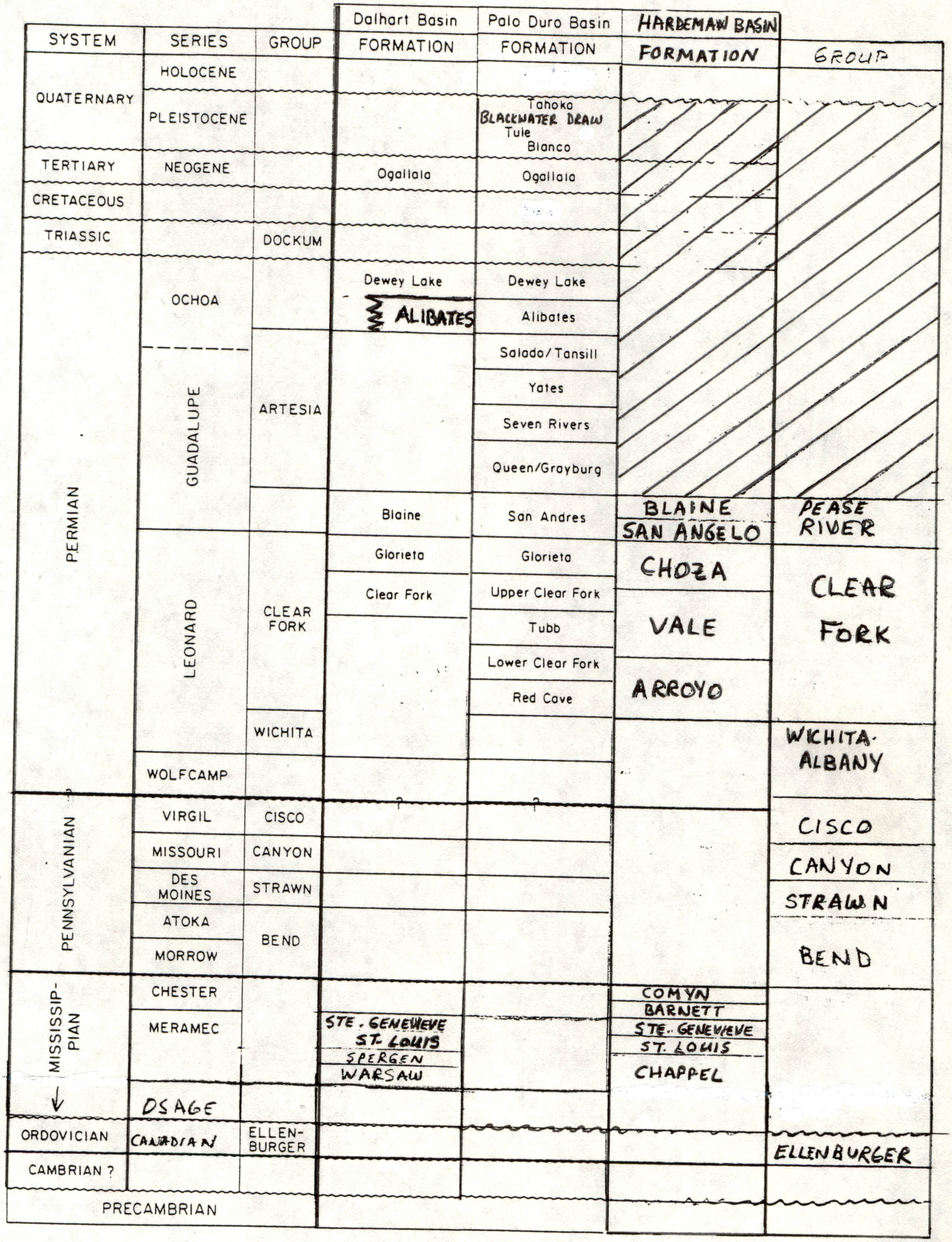

Figure 1. Stratigraphic column of the Palo Duro, Hardeman, and Dalhart Basins. After Handford and Dutton (1980).

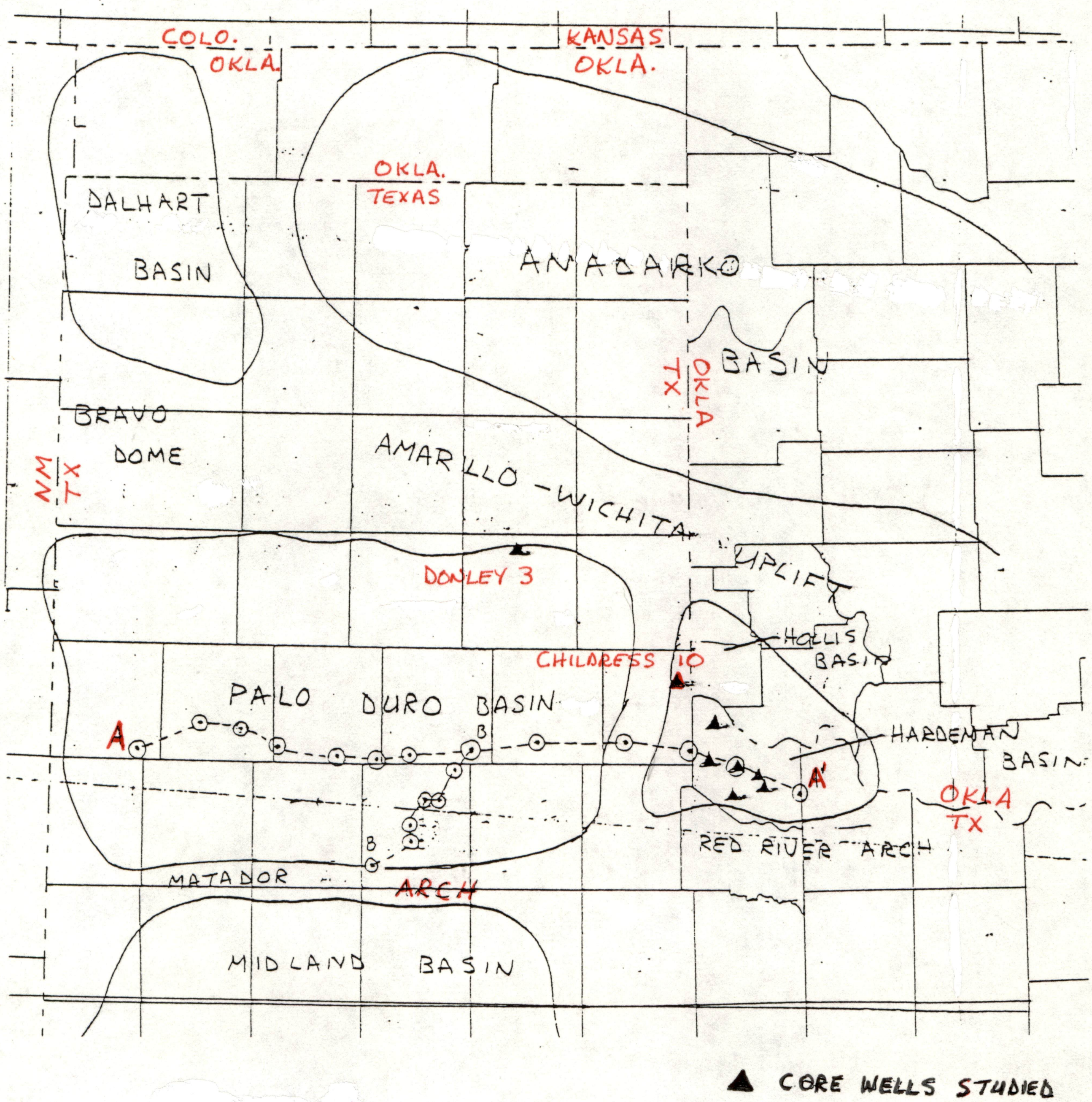

Figure 2. Location map showing lines of section and core wells.

See Figures 3 through 9
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SYSTEM	SERIES	DALHART / ANADARKO BASINS (Cunningham, 1961)	PALO DURO BASIN (Nicholson, 1960)	HARDEMAN BASIN	MID-CONTINENT REGION (MAPEL ET AL., 1977)
MISSISSIPPIAN	CHESTER	CHESTER	CHESTER	COMYN	D
				BARNETT	
	MERAMEC	STE. GENEVIEVE	MERAMEC	STE. GENEVIEVE	C
		ST. LOUIS		ST. LOUIS	
		SPERGEN		CHAPPEL	
		WARSAW			
	OSAGE	OSAGE	OSAGE		B
	KINDERHOOK	KINDER-HOOK			A

Figure 10. Common schemes of stratigraphic nomenclature applied to Mississippian rocks in the Texas Panhandle area. Meramec formations used in the Dalhart and Anadarko Basins are, in some cases, extended into the Palo Duro Basin with varying degrees of accuracy. The Ste. Genevieve is probably most consistent in its application. Usage in the Hardeman Basin varies considerably. Some workers place the St. Louis at the top of the Meramec; others consider the top of the Chappel as the top of the Meramec.

See Figures 11 through 16
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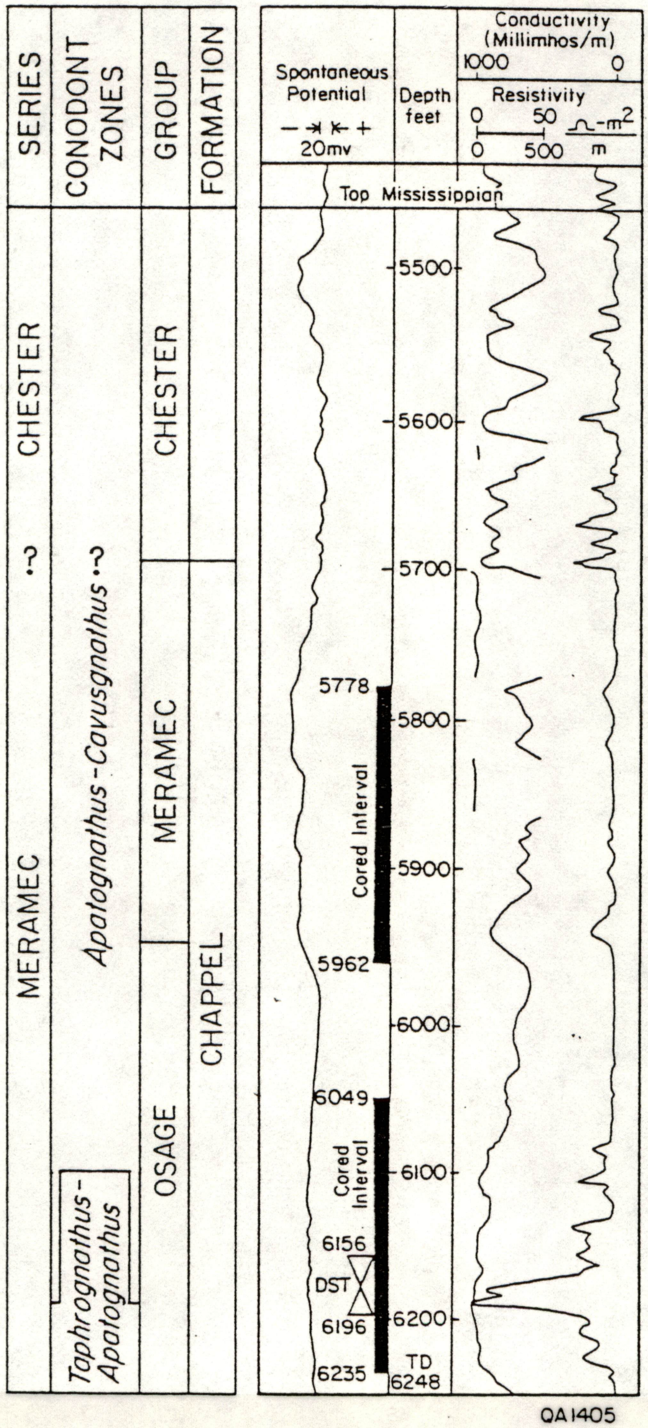

Figure 17. Mississippian section and core intervals in the Childress 10 well (location shown in figure 2). Mississippian Group designations reflect common terminology applied to the section in the area. Conodont zones are based on identification of fossils from cored intervals. These data indicate that both the Meramec and Osage in the Childress 10 well are Meramecian in age.

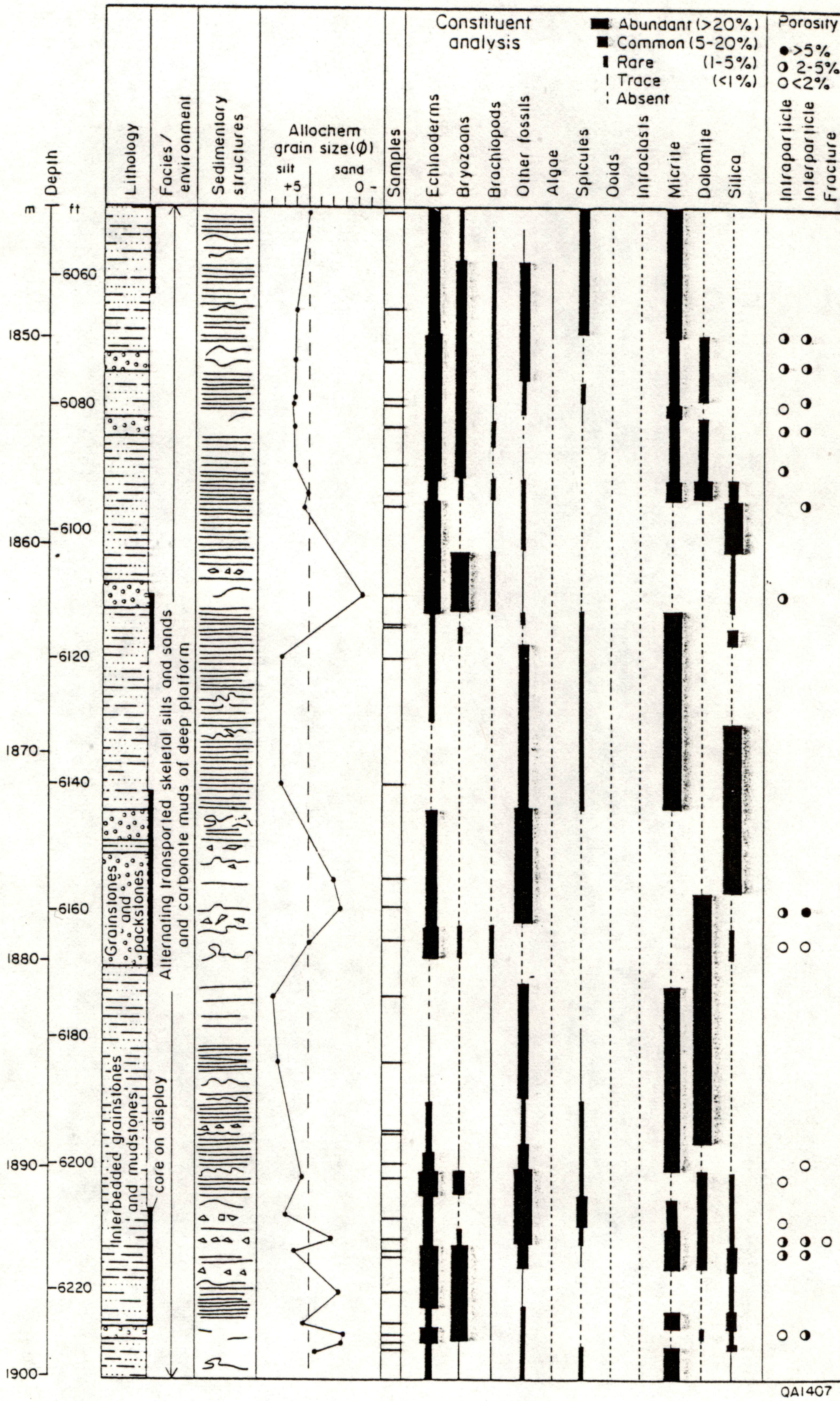

Figure 18. Core analysis of the Osage Group (lower Chappel) in the Childress 10 well.

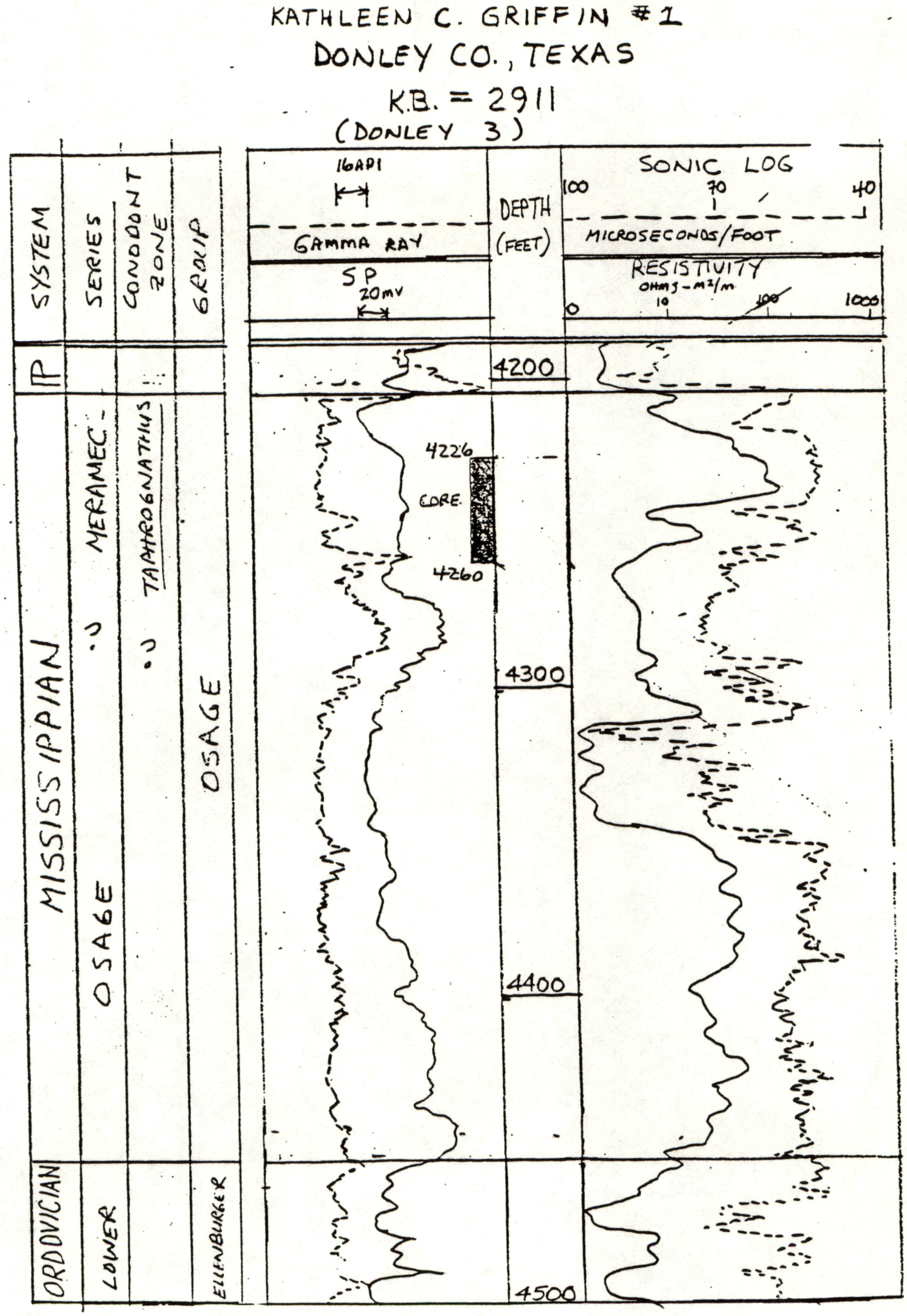

Figure 19. Pre-Pennsylvanian section and cored interval in the Donley 3 well (location shown in figure 2). Conodonts recovered from this core suggest a late Osage or early Meramec age for these deposits.

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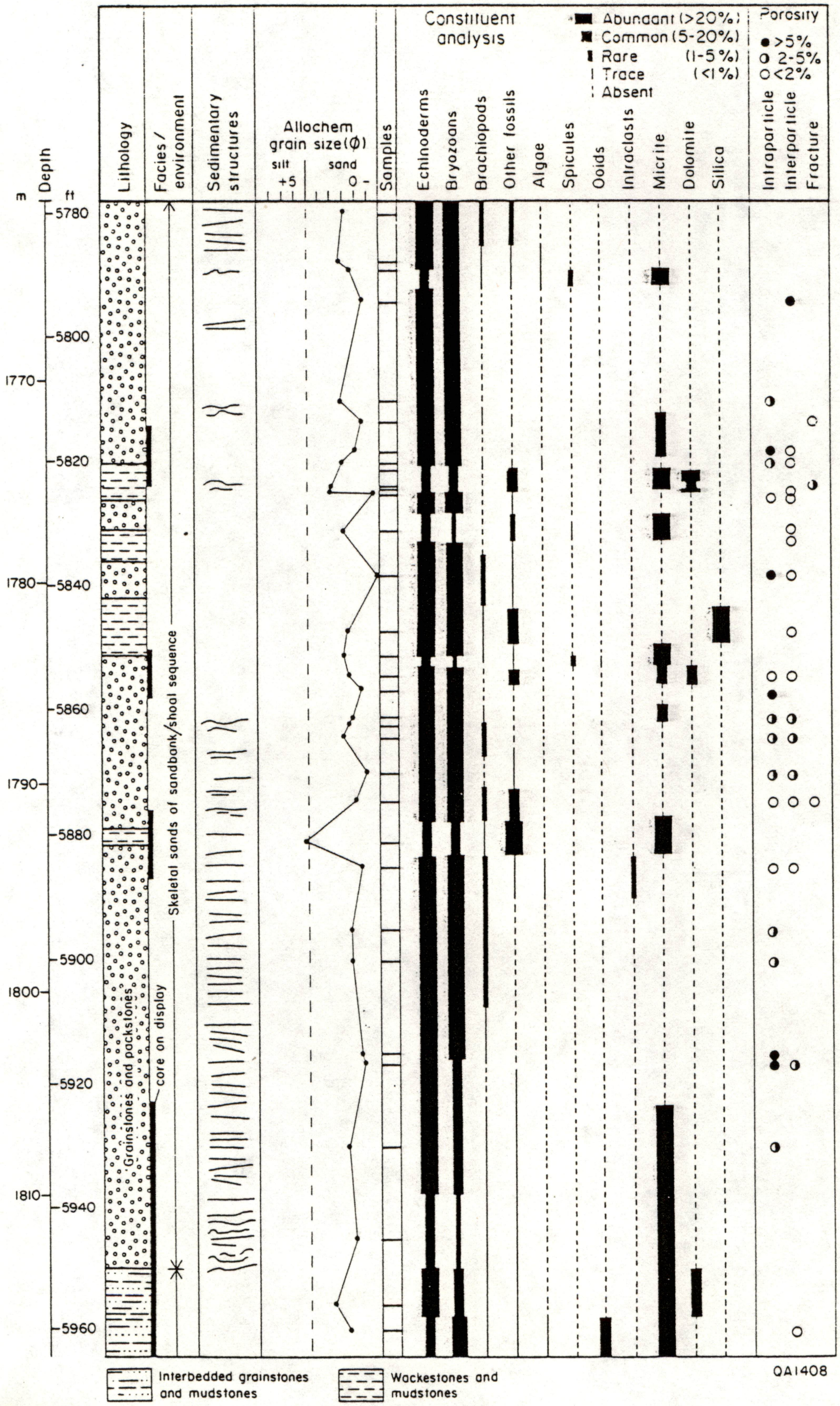

Figure 21. Core analysis of the Meramec Group in the Childress 10 well (location shown in figure 2).

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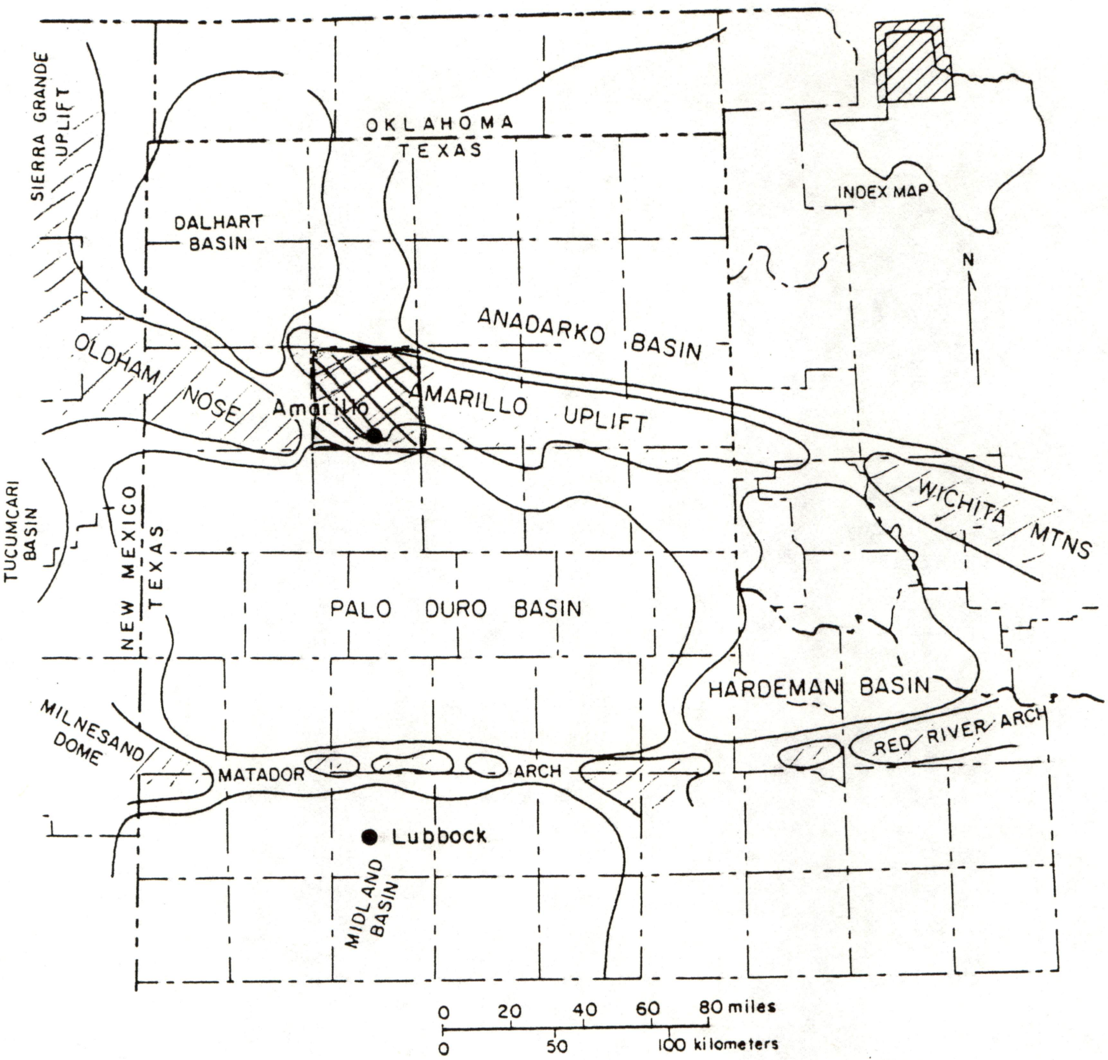

Figure 25. Location map showing Potter County (hatched lines) and major structures, Texas Panhandle. After Nicholson, 1960.

See Figures 26 through 36
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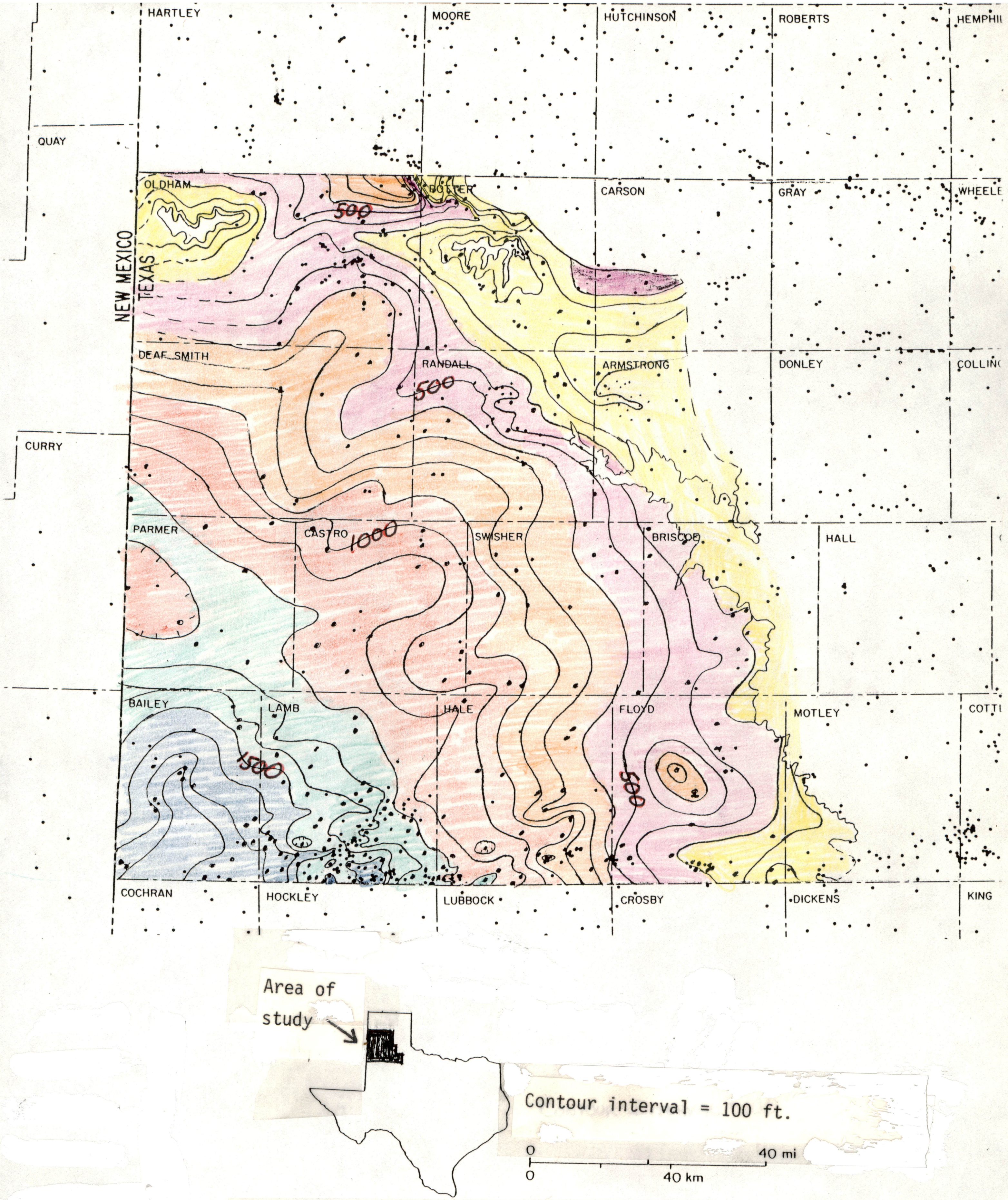


Figure 37. Isopach map of the Dockum Group.

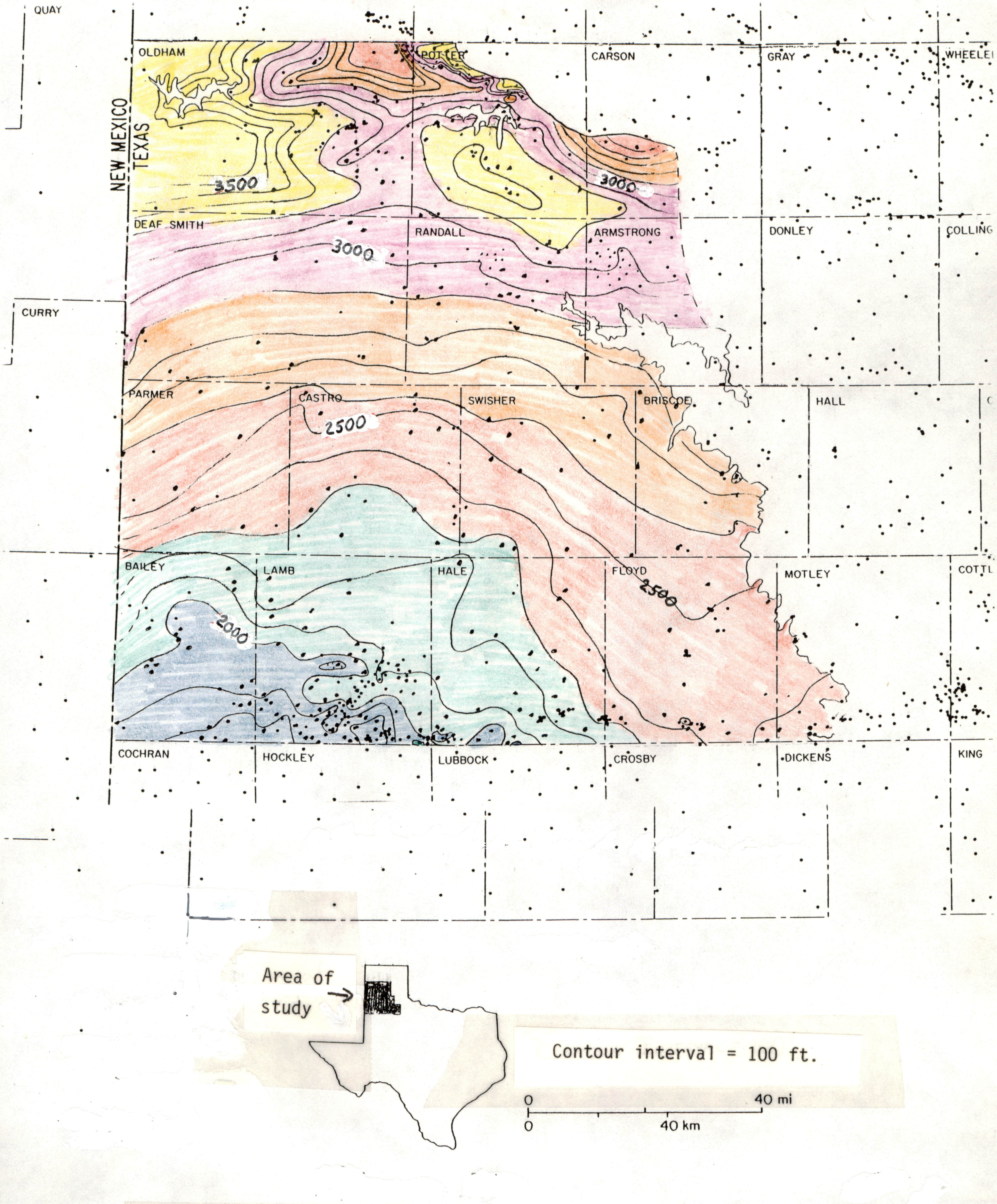


Figure 38. Structure contour map on base of the Dockum Group.

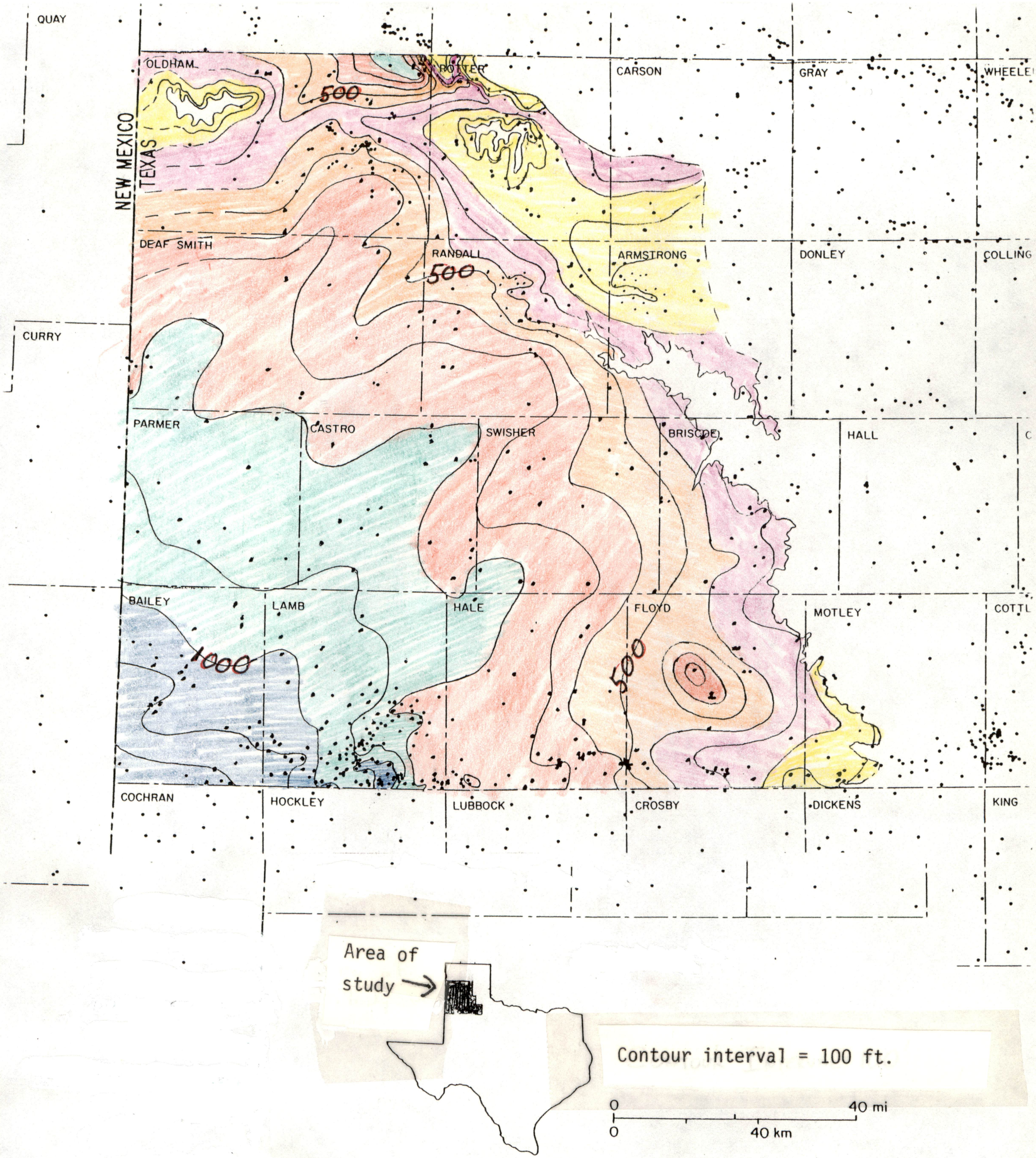


Figure 39. Isopach map of the lower Dockum.

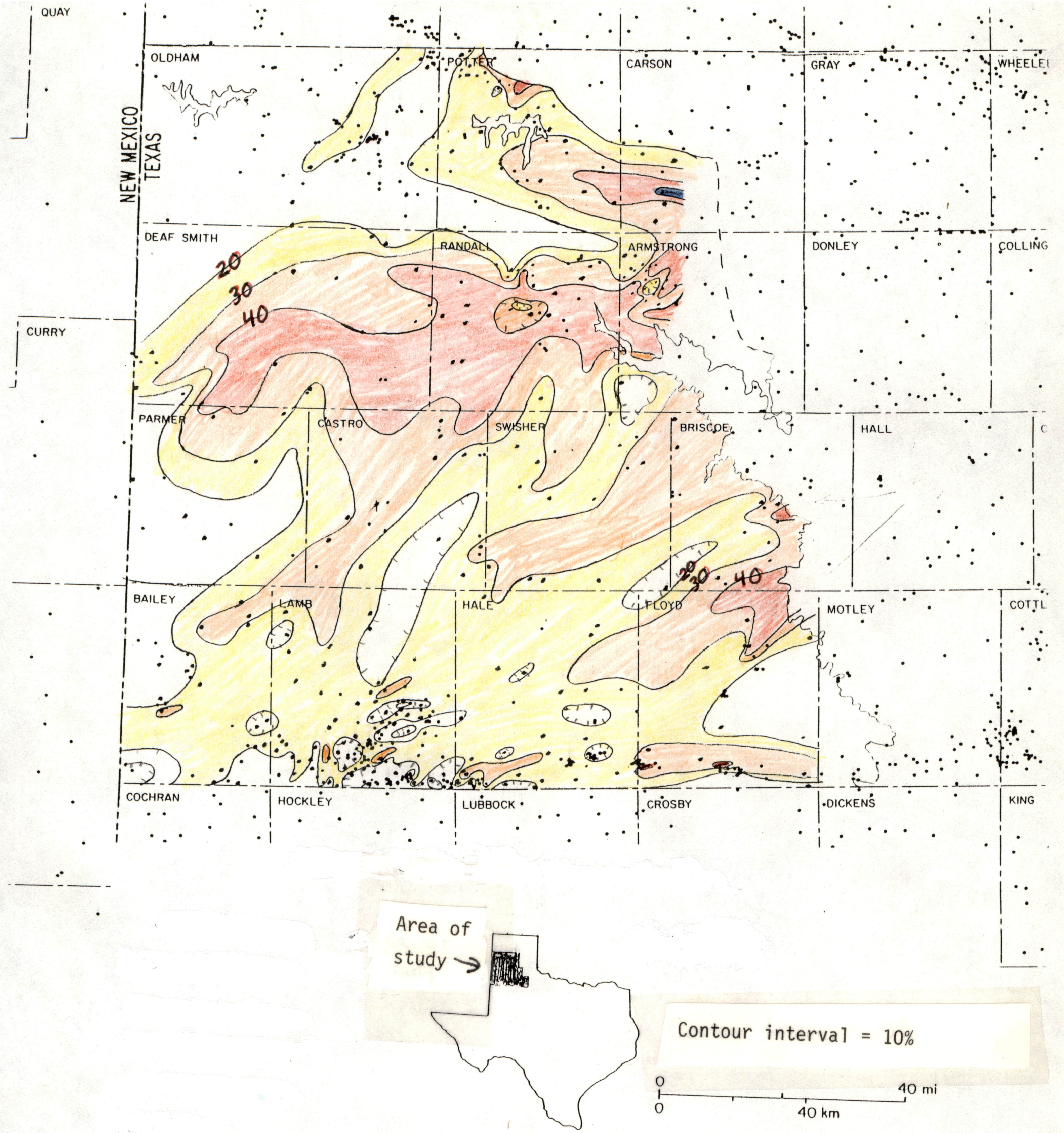


Figure 40. Percent sandstone map of the lower Dockum.

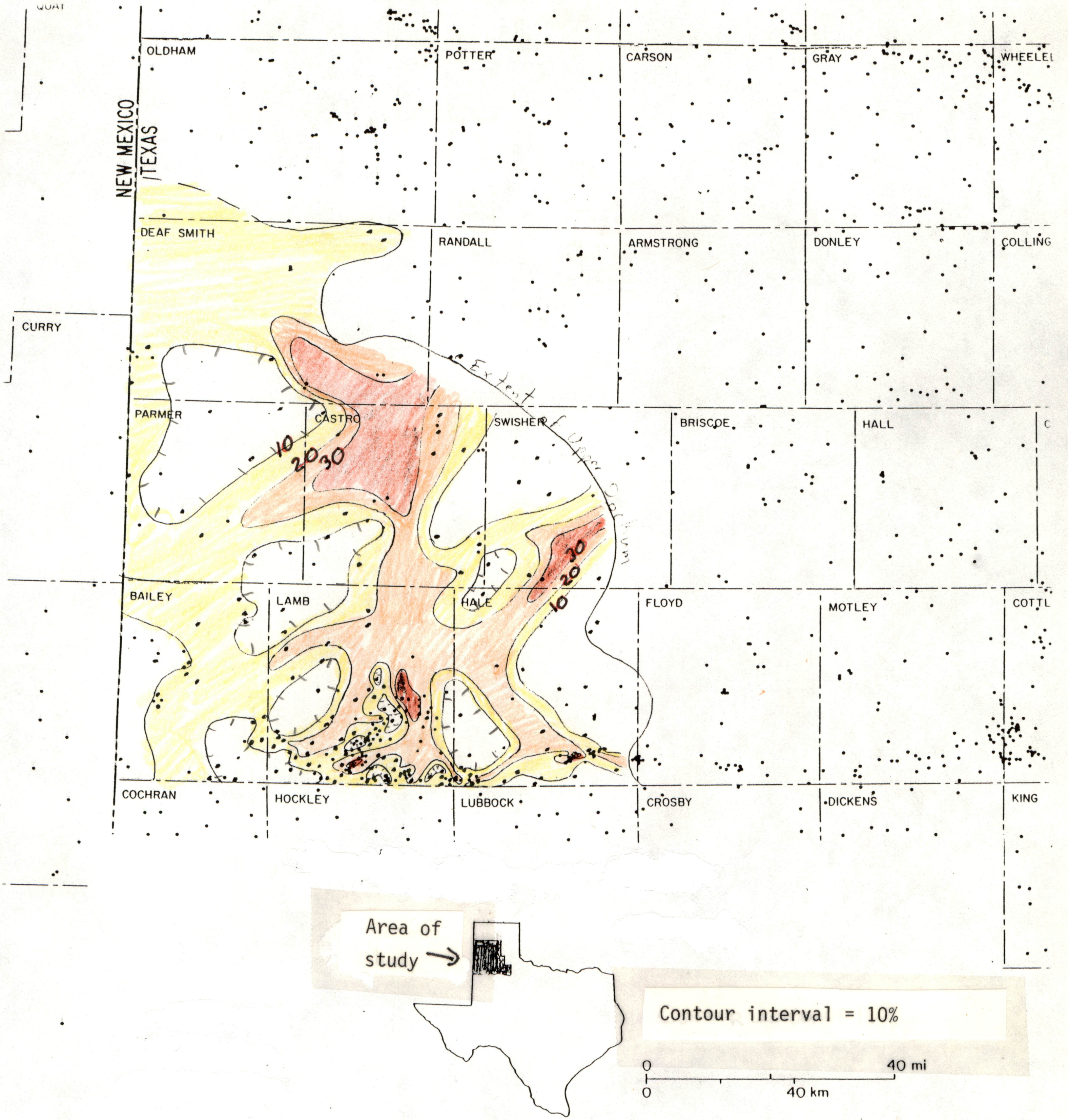
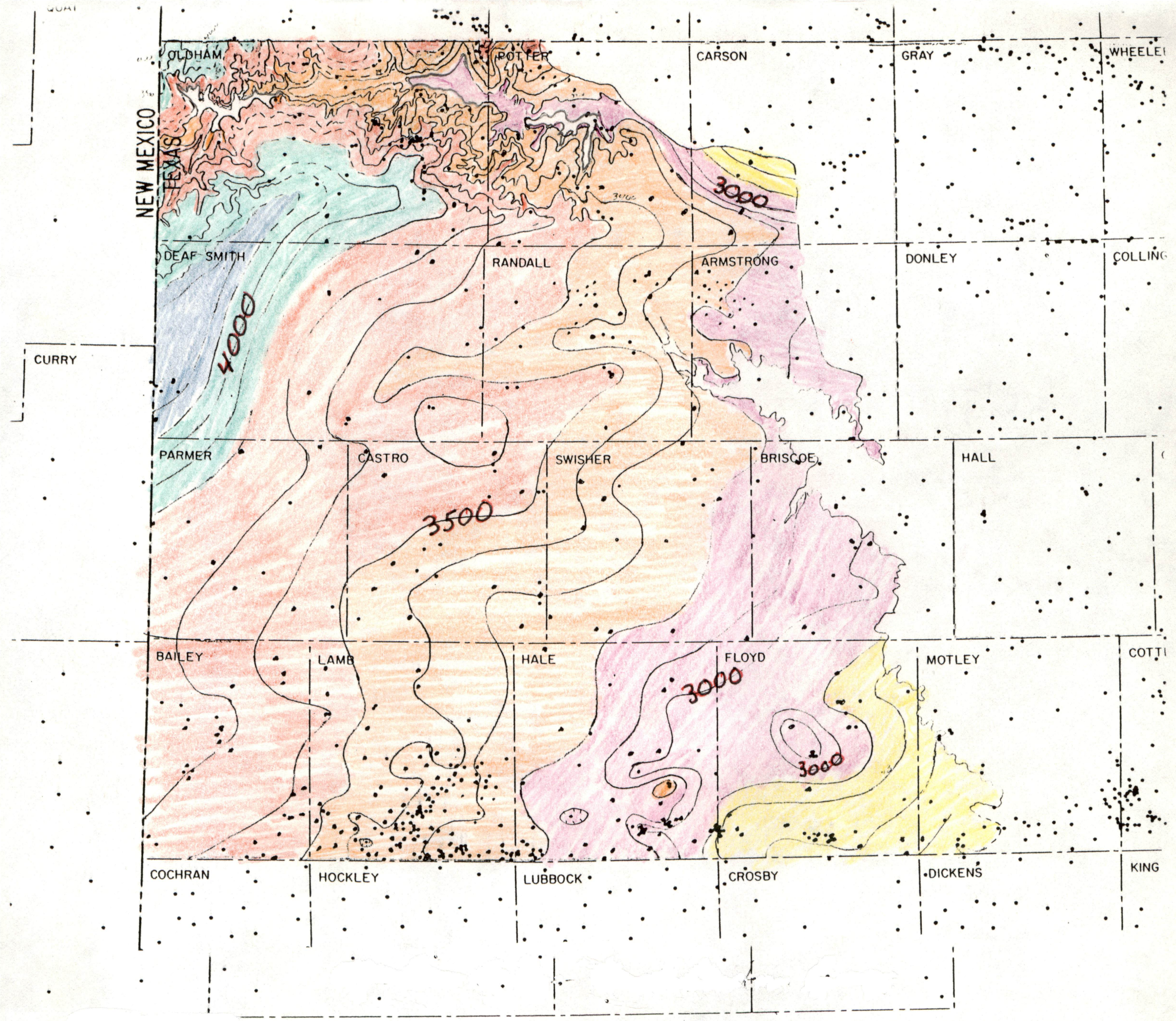
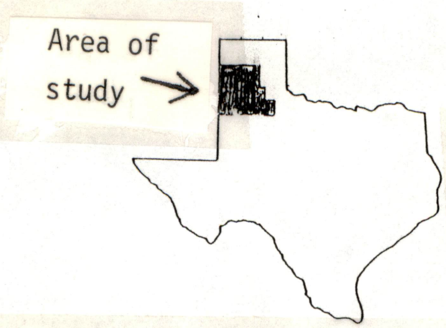


Figure 41. Percent sandstone map of the upper Dockum.





Contour interval = 100 ft.

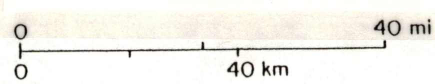


Figure 42. Structure contour map on top of the Dockum Group.