GEOLOGY AND HYDROGEOLOGY OF THE PALO DURO BASIN, TEXAS PANHANDLE–A Synthesis of Nuclear Waste Isolation Feasibility Studies, 1977-1987

by

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ABSTRACT

Detailed studies of the geology and hydrogeology of Permian salt beds and adjacent strata in the Palo Duro Basin, Texas Panhandle, began in 1977 when the U.S. Department of Energy (DOE) initiated investigations to consider the salt beds as potential host rocks for a high-level radioactive waste repository. Studies continued until 1987, when the U.S. Congress directed that repository-siting focus solely on Yucca Mountain in Nevada. Knowledge of the geology and hydrogeology of the large and complex Palo Duro Basin was minimal in 1977, in large part because of the previously limited success of hydrocarbon exploration in the basin. The ensuing decade of research produced significant new knowledge of stratigraphy, regional tectonics, geomorphology, hydrocarbon potential, geochemistry, hydrogeologic properties, and ground-water flow in the basin. This report synthesizes and examines the extensive results of these geologic and hydrogeologic studies of the Palo Duro Basin and identifies the perceived scientific issues that remain to be addressed and answered before high-level nuclear waste could be buried in the Texas Panhandle in any future program.

Some geologic and hydrogeologic aspects of the Palo Duro Basin still are not well defined where geologic and hydrogeologic data are sparse and where interpretations differ. Principal among these scientific issues are questions pertaining to: timing of tectonic and diagenetic events affecting Permian sediments and host-rock stability; thermal maturity of hydrocarbon source rocks and the potential for additional discoveries of oil and gas in the basin; recharge rates and relative significance of different sources of recharge to the High Plains aquifer and to the aquifer in the Triassic Dockum Group; ground-water flow velocity through the evaporite section and deep-basin aquifers, influence of fractures on flow velocity, and whether to treat flow as a steady-state phenonemon; origin and age of brine in deeply buried formations

beneath the salt section; origin, paleoenvironment, and rate of formation of playa-lake and other lacustrine basins during the Cenozoic Era; Quaternary and present rates of salt dissolution; and the varying local extent of salt dissolution beneath the Southern High Plains.

INTRODUCTION

Geologic and hydrogeologic studies of the Palo Duro Basin (fig. 1) were conducted from 1977 to 1987 to determine the feasibility of siting a nuclear waste repository in bedded Permian halite. In 1984, a site in Deaf Smith County in the Texas Panhandle was recommended by the U.S. Department of Energy (DOE) for detailed characterization as one of three sites in the United States deemed potentially acceptable for the nation's first repository. As a result of congressional amendments to the Nuclear Waste Policy Act of 1982 (P.L. 97-425) in December, 1987, all activities directed toward site characterization were terminated at all but the Yucca Mountain site in Nevada. Consideration of the Texas Panhandle site may be renewed if the DOE determines the Yucca Mountain site is unsuitable for development as a repository.

The purpose of this report is to compile, synthesize, and examine the results of geologic and hydrogeologic studies of the Palo Duro Basin so that existing scientific information, conclusions, and identified issues and uncertainties will be preserved as part of the public record. The Palo Duro Basin studies provided a comprehensive and detailed analysis of a large and complex sedimentary basin and developed generic conceptual models of basin-scale tectonic, depositional, and geomorphic processes, hydrocarbon potential, and evolution of ground-water flow systems. This report presents interpretations of stratigraphy, tectonics, structure, hydrogeology, geochemistry, and geomorphology of the Palo Duro Basin. The summary is intended to guide future investigations if the area is reconsidered for a repository site. Although comprehensive in scope, the synthesis is by no means a complete account of all work dealing with various aspects of geology and hydrogeology in the basin. The reader is encouraged to consult the numerous publications on the geology and hydrogeology of the Palo Duro Basin, most of which are cited in this report, for details on particular studies.

The Bureau of Economic Geology, a research unit of The University of Texas and the State of Texas, has gathered and interpreted geologic and hydrogeologic data necessary for describing, delineating, and evaluating Permian salt-bearing and bounding strata (table 1) in the Palo Duro Basin and adjacent areas. Other contractors to the DOE, including Stone and Webster Engineering Corporation and Battelle Memorial Institute, prepared published and unpublished reports of studies, some of which are cited in this summary. The research program at the Bureau was divided into broad tasks that addressed (1) geomorphic processes and Quaternary stratigraphy, (2) basin analysis, (3) seismicity and tectonics, and (4) hydrology and geochemistry. Efforts within these tasks included regional investigations as well as support for the DOE in analyzing and interpreting data from test wells and cores obtained as part of site evaluations. Regional studies not only served to delimit possible sites for detailed characterization, but were essential in providing a framework for interpreting the results of local studies.

Investigation of geologic structure, stratigraphy, surface processes, and salt dissolution began in 1977 and hydrogeologic studies began in 1979. Initial tectonic studies were designed to delineate the history of faulting within the basin. Stratigraphic studies focused on salt quality, lateral and vertical continuity of salt units, and analysis of depositional facies in strata below the salt. An erosion-monitoring network was installed to determine erosion rates. Stratigraphic mapping identified areas of salt dissolution and results of dissolution-induced subsidence. Early resource assessment studies focused on the potential for hydrocarbon generation in the Palo Duro Basin. Hydrogeologic studies initially focused on Permian salt-bearing units and underlying deep-basin strata owing to the fact that hydraulic head of ground water is significantly lower in the deep-basin strata than in near-surface aquifers. Movement of radioactive contaminants, if a repository failed to contain wastes, would be downward through the evaporite section to deeper parts of the basin. In 1979 and 1980, two

boreholes were drilled to obtain rock core from evaporite strata for geologic studies. From 1982 to 1984, the DOE drilled ten additional test wells from which core was obtained and in which hydrologic tests were conducted in selected intervals. In this report, these wells are referred to as DOE wells or DOE test wells. Also during the period from 1982 to 1984, investigations were begun to define (1) rates of recharge to the important High Plains aquifer, which overlies the Permian evaporite strata, (2) availability of ground water in Triassic rocks, and (3) the movement of ground water from these aquifer units downward into salt-dissolution zones that occur at the top of the evaporite section. Hydrogeologic studies relied extensively on stratigraphic and diagenetic framework established in previous and concurrent studies. Computer, or numerical, models of ground-water flow were constructed simultaneously with the development of conceptual hydrogeologic models to test explanations of flow paths and hydrologic processes and to quantify travel times from a repository horizon through the evaporite section and deep-basin strata.

This report is organized in four parts. First, the entire stratigraphic section from basement to ground surface (table 1) is reviewed, and basin history from its formation approximately 1,000 mya to the present is discussed. Second, hydrocarbon potential in the basin is reviewed. Third, results of hydrogeologic studies of the ground-water flow systems and flow velocities are synthesized. In this third section, the order of discussion of hydrogeologic units is downward from land surface, following the conceptualized flow path of recharged ground water through the basin. Fourth, results of studies of the salt-dissolution zone at the contact between shallow aquifer systems and the Permian evaporite section are summarized.

PALEOZOIC STRATIGRAPHY AND BASIN EVOLUTION

S. C. Ruppel

The Palo Duro Basin of the southern Texas Panhandle contains as much as 13,000 feet (4,000 m) of sedimentary rock, most of it of Paleozoic age (approximately 245 to 570 mya). Much of the effort devoted to characterizing the basin as a possible nuclear waste repository, thus, has been devoted toward understanding Paleozoic stratigraphy. Especially important in this regard are interpretations of (1) distribution and extent of sedimentary facies, (2) sedimentary processes, and (3) tectonic controls on deposition. These interpretations are crucial to accurate predictions of tectonic stress, salt-dissolution rates, ground-water flow velocity, and possible rate of radionuclide transport. This section summarizes the stratigraphy of the Paleozoic sequence in the Palo Duro Basin.

Regional Setting

The Palo Duro Basin is one of several structural elements constituting the southern Midcontinent (fig. 1). As used by most workers, Palo Duro Basin refers to the broad subsurface structural depression that occupies the southern half of the Texas Panhandle (fig. 1). The present structural configuration of the basin is primarily the result of the interplay of differential subsidence and sedimentation during the Pennsylvanian and Permian Periods. The boundaries of the basin are defined primarily by other structures that formed at this time (fig. 1).

The Palo Duro Basin is separated from the Anadarko Basin to the north by the Amarillo Uplift, a major northwest-southeast trending block of 1,400 m.y. crystalline rocks that exhibits as much as 10,000 ft (3,000 m) of offset to the north. The Bravo Dome and several smaller structural highs separate the Palo Duro Basin from

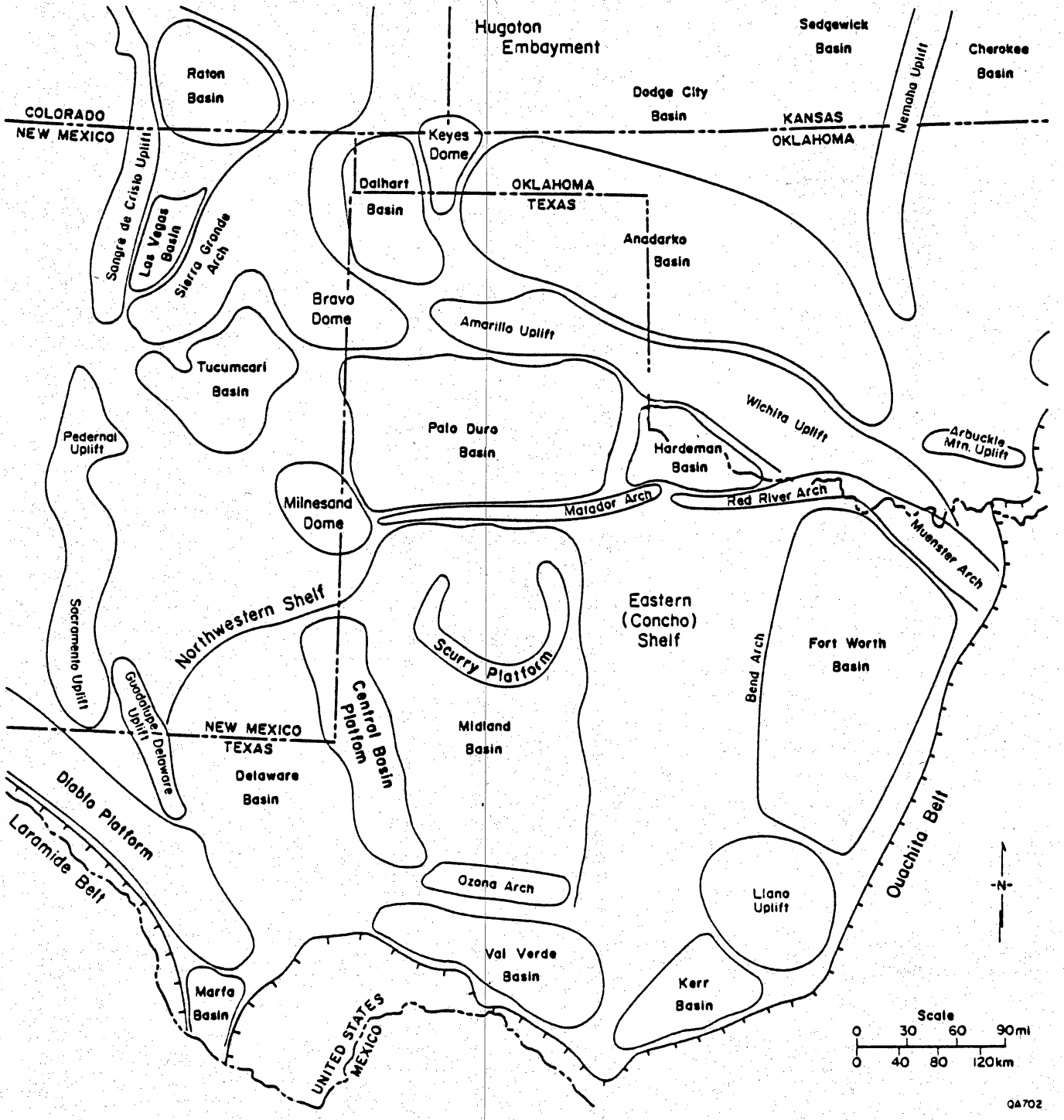


Figure 1. Major subsurface structural elements in the southern Midcontinent. Most of these features were formed during the late Paleozoic. From Ruppel (1983).

the Dalhart Basin to the northwest (fig. 1). The southern margin of the basin is formed by the Matador Arch, a discontinuous series of horst blocks that trend east-west. The eastern and western margins of the basin are less well defined. To the east the Palo Duro Basin is separated from the Hardeman Basin by a low saddle. In contrast with most of the structural features in the region, present structural distinction between the Palo Duro Basin and the Hardeman Basin may be more closely related to pre-Pennsylvanian tectonics and sedimentation than late Paleozoic events (Budnik, 1988). A series of poorly defined basement uplifts separates the Palo Duro Basin from the Tucumcari Basin to the west (fig. 1).

Stratigraphy and Depositional Systems

The Palo Duro Basin contains a relatively thick sequence of Paleozoic rocks (table 1) that exceeds 10,000 ft (3,000 m) in the deepest part of the basin. These rocks overlie Precambrian crystalline rocks and are in turn overlain by rather thin sequences of Mesozoic (principally Triassic) and Cenozoic deposits (see DuBar, this volume).

Precambrian

Precambrian basement rocks in the southern Texas Panhandle comprise felsic volcanics, apparently coeval granites, and younger, interbedded mafic volcanics and sediments (Flawn, 1956; Budnik, 1988).

Much of the Palo Duro Basin is underlain by thick rhyolite of the Panhandle Volcanic Terrane (fig. 2). Recent estimates indicate that these rocks are approximately 1,000 m.y. in age (Thomas and others, 1984). Although the total thickness of these deposits is unknown, nearly 4,000 ft (1,200 m) has been encountered in some

Table 1. Generalized stratigraphic chart and general lithology, Texas Panhandle.

SYSTEM	SERIES	Group	Formation	General Lithology and Depositional Setting
QUATERNARY	HOLOCENE			Playa deposits
	PLEISTOCENE		Tule	Eolian and lacustrine clastics, freshwater limestones
			Blackwater Draw	
TERTIARY	PLIOCENE		Blanco	Fluvial, eolian, and lacustrine clastics
	MIOCENE		Ogallala	
CRETACEOUS	COMANCHE	Washita	Duck Creek	Marine sandstones, limestones, and shales
		Fredericksburg	Kiamichi	
			Edwards	
			Walnut	
		Trinity	Paluxy	Transgressive sand and gravel
JURASSIC*	UPPER		Morrison	Marine sandstone and shale
TRIASSIC	UPPER (?)	Dockum	Trujillo: Chinle / Santa Rosa	Fluvial-deltaic and lacustrine clastics
			Tecovas	
PERMIAN	OCHOA		Dewey Lake	Brine-pool salt, anhydrite, red beds, and peritidal carbonates
			Alibates	
	GUADALUPE	Artesia	Salado	
			Tansill	
			Yates	
			Seven Rivers	
			Queen	
			Grayburg	
		Pease River	San Andres (Blaine)	
			Glorieta	
	LEONARD	Clear Fork	upper Clear Fork	
			Tubb	
			lower Clear Fork	
			Red Cave	
		Wichita-Albany		
	WOLFCAMP		"Brown dolomite"	Shelf and shelf-margin carbonate, basinal shale, and deltaic sandstone
PENNSYLVANIAN	VIRGIL	Cisco		
	MISSOURI	Canyon		
	DESMOINES	Strawn		
	ATOKA	Bend		
	MORROW			
MISSISSIPPIAN	CHESTER			Shelf carbonate and chert
	MERAMEC			
	OSAGE			
	KINDERHOOK*			
DEVONIAN*		Hunton		Shelf carbonates and clastics
SILURIAN*				
ORDOVICIAN	CINCINNATI*	Viola		
	CHAMPLAIN*	Simpson		
	CANADA	Ellenburger		
CAMBRIAN				Shallow marine sandstone
PRECAMBRIAN				Igneous and metamorphic

* Unit absent or insignificant in the Palo Duro Basin

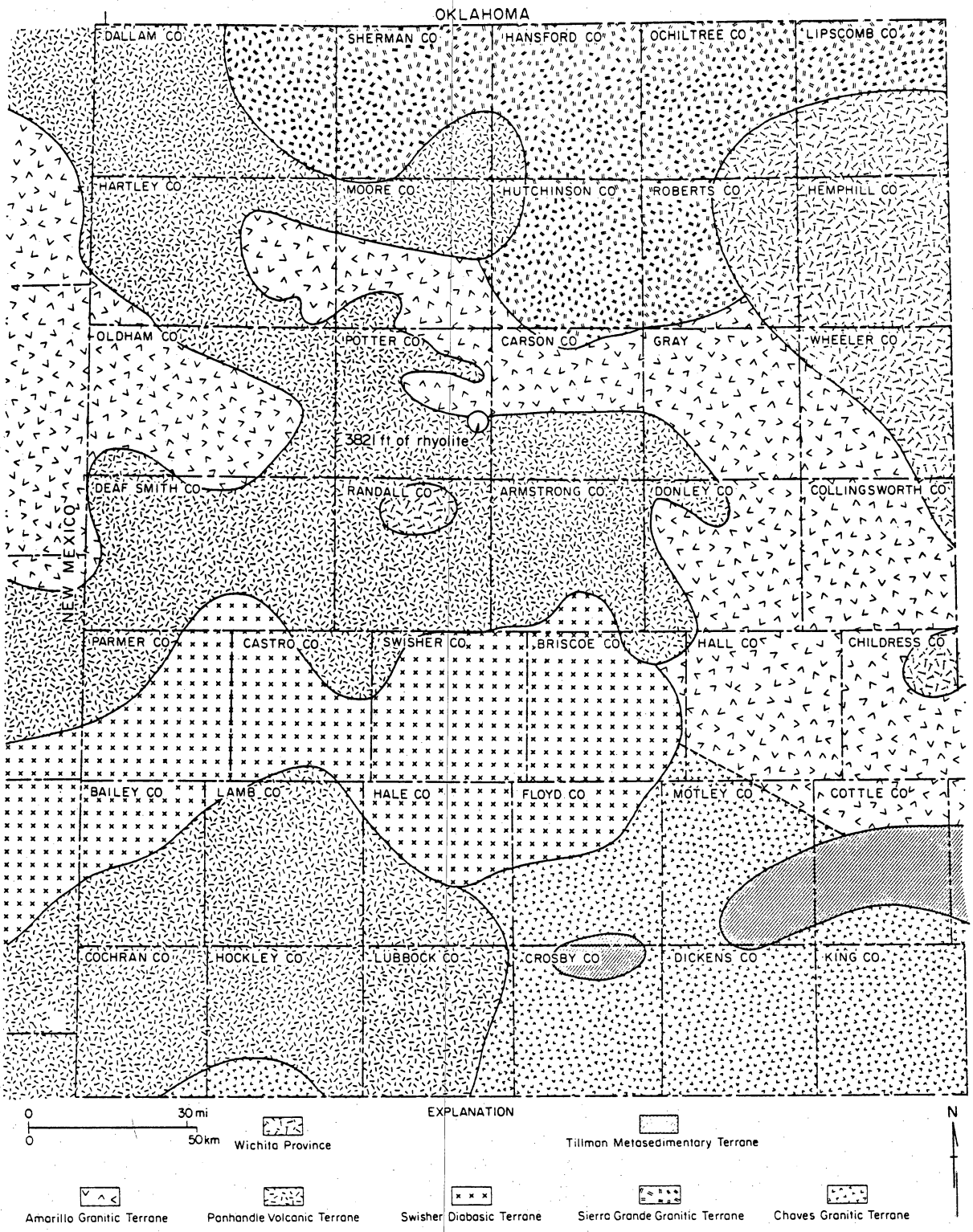


Figure 2. Distribution of Precambrian basement terranes in the Texas Panhandle. From Muehlberger and others (1967).

drill holes (Budnik, 1988). Gravity lows correlate with the known distribution of these rocks suggesting a very thick sequence may be present (Goldstein, 1982). Granites (Amarillo Granitic Terrane) of approximately the same age underlie the northern and eastern margins of the Palo Duro Basin (fig. 2). Regional studies suggest that these rhyolites and granites may be part of an extensive volcanic/plutonic province that extends from New Mexico to Ohio (Van Schmus and Bickford, 1981).

A younger sequence of mafic rocks (Swisher Diabasic terrane) overlies older volcanic/granitic terranes in the central part of the basin (fig. 2). The presence of interbedded, fossiliferous sediments in this sequence suggests that these rocks may be at least in part as young as Paleozoic (Roth, 1960).

Lower Paleozoic

Lower Paleozoic rocks in the Palo Duro Basin consist of siliciclastics of possible Cambrian age and dolostones of probable Early Ordovician age. Unnamed Cambrian(?) siliciclastics overlie Precambrian crystalline rocks in the eastern and southern parts of the Palo Duro Basin (fig. 3). These deposits, which seldom exceed thicknesses of 50 ft (15 m), may be equivalent to similar basal siliciclastics deposited across the emergent Texas craton in the middle to late Cambrian (Barnes and others, 1959; Ruppel, 1985). Such deposits are thought to have formed in a nearshore marine setting (Cornish, 1975; Dutton, 1980a; Ruppel, 1985).

Overlying the basal Cambrian siliciclastics are thicker and more continuous dolostones of the Ellenburger Group (fig. 4). These rocks comprise cryptalgal and burrowed dolomudstones that were deposited in a broad shallow sea that extended across much of the North American craton during the Early Ordovician (Ruppel, 1985; Kerans, in press). The Ellenburger is absent in much of the Palo Duro Basin along the Texas Arch, a northwest-trending structural positive (Adams, 1954). The

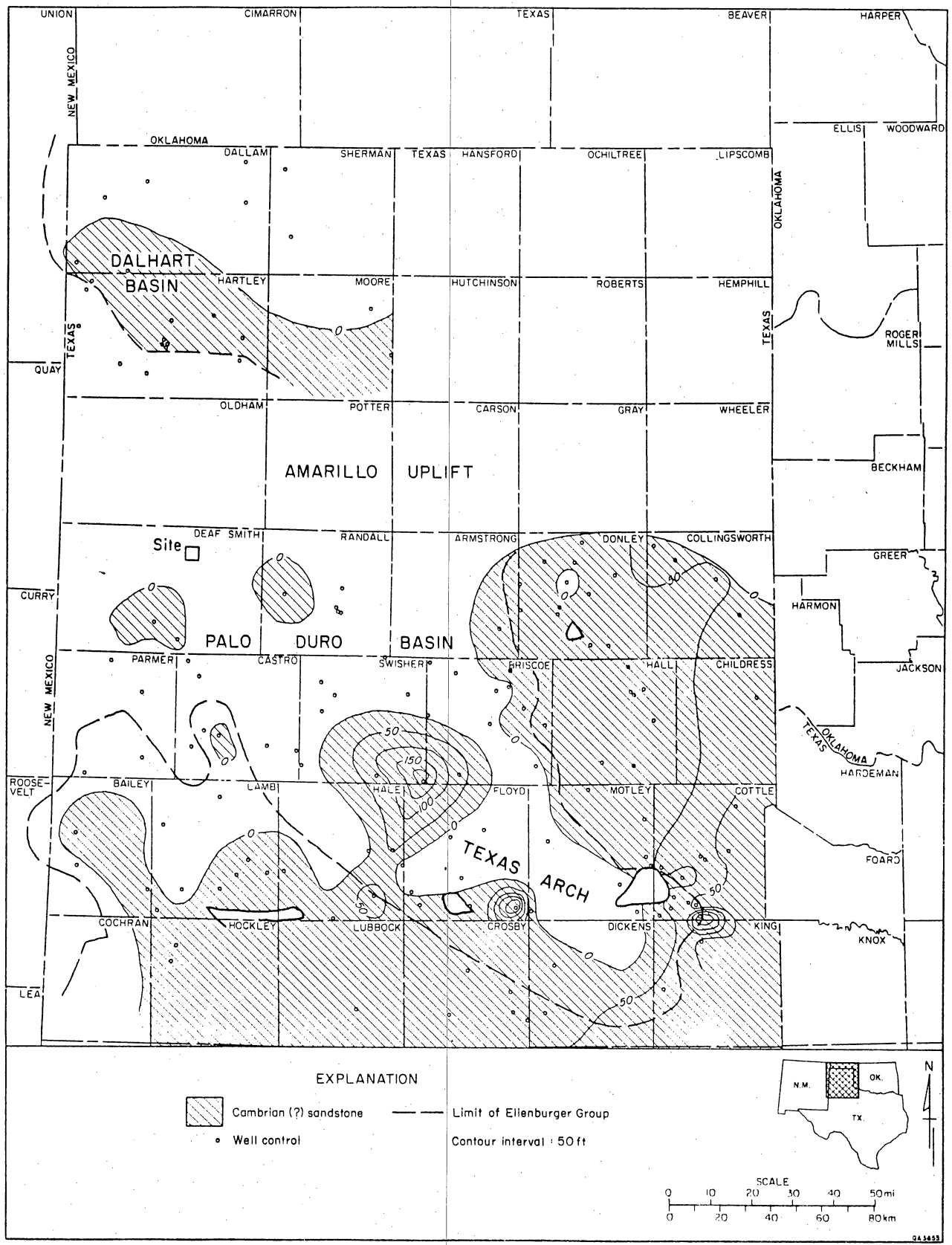


Figure 3. Isopach of Cambrian(?) rocks in the Texas Panhandle. From Ruppel (1985).

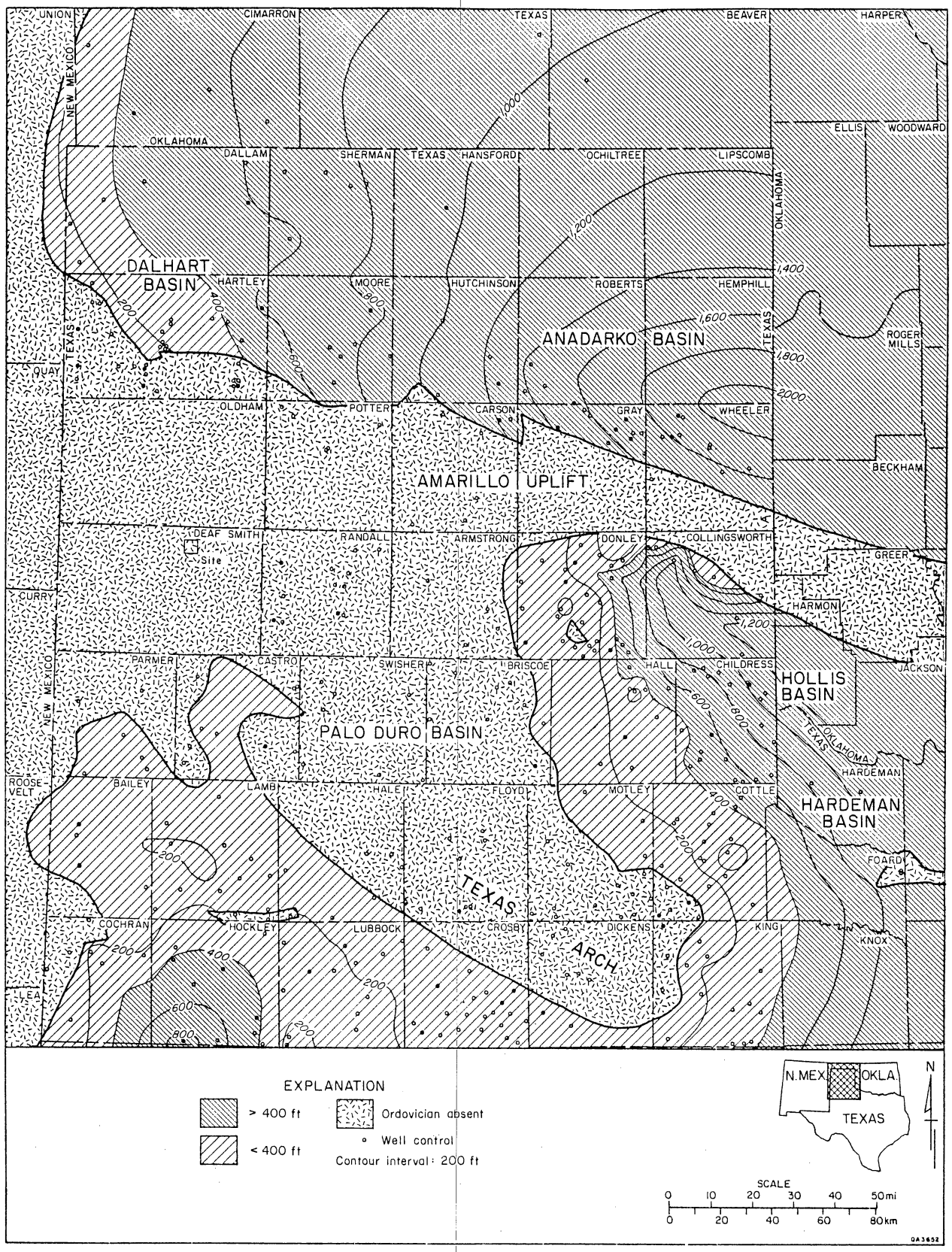


Figure 4. Thickness of Ellenburger Group in the Texas Panhandle. From Ruppel (1985).

Ellenburger sequence thickens away from this feature, especially to the northeast toward the Amarillo Uplift (fig. 4).

Middle Ordovician through Devonian rocks are absent from the Palo Duro Basin. A relatively complete sequence of these rocks is present, however, to the north in the Anadarko Basin (fig. 5). Less complete sequences are found in the Midland Basin to the south and in the Hardeman Basin to the east (fig. 1).

Mississippian

Mississippian rocks comprise a relatively thick sequence of predominantly carbonate rocks that covers much of the Texas Panhandle. These rocks are absent only along the Amarillo-Wichita Uplift and associated structures that trend northwest-southeast through the area (fig. 6).

Mississippian rocks can be subdivided into four lithologically distinct parts. Although the ages of these rocks are poorly known, the four lithologic units have commonly been referred to using standard Mississippian series names (table 1).

Kinderhookian

Basal Mississippian deposits in the northern Texas Panhandle consist of shales and quartzose sandstones that are commonly referred to as Kinderhook (Ruppel, 1985). Although rare in the Palo Duro Basin, these deposits reach thicknesses greater than 150 ft (45 m) in the northern Panhandle (fig. 7). The distribution of these rocks suggests that they were derived from exposed Precambrian basement along the Texas Arch in the vicinity of what is now the Amarillo Uplift (fig. 7). The age of these rocks is unknown. However, stratigraphic relationships with overlying and underlying rocks in the area suggest that they are Late Devonian to Early Mississippian (Kinderhookian) near-shore marine equivalents of basinal shales of the

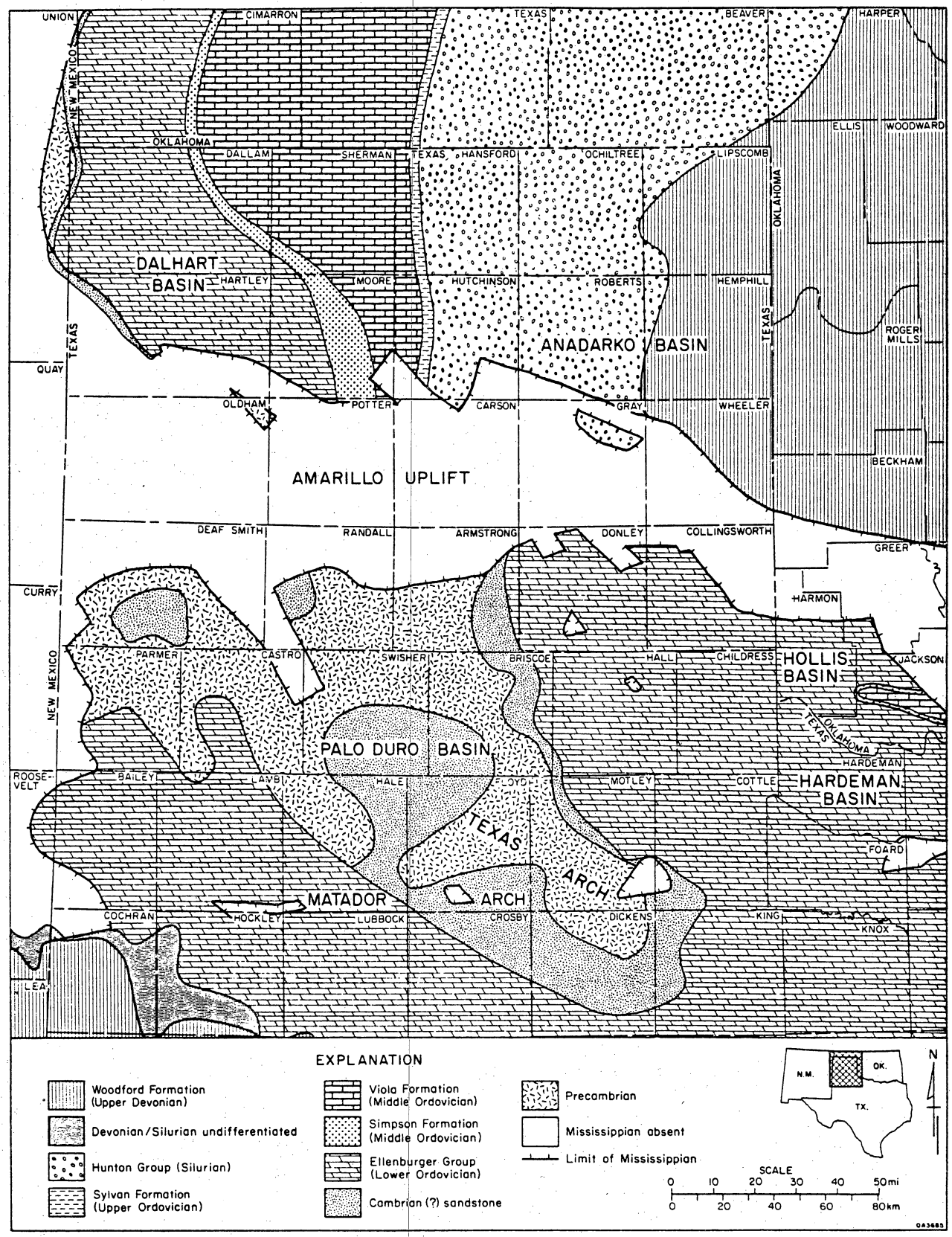


Figure 5. Mississippian subcrop, Texas Panhandle. From Ruppel (1985).

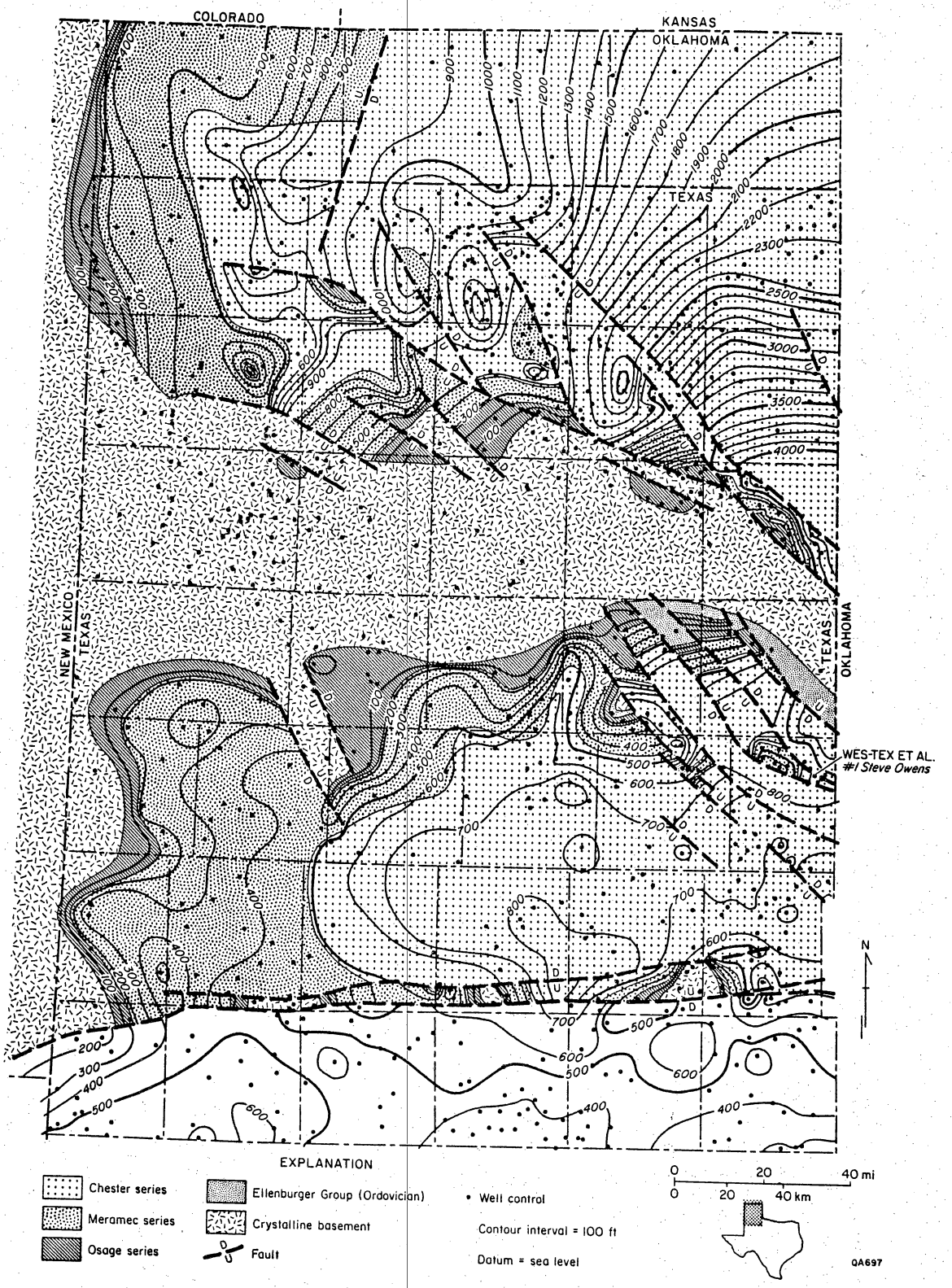


Figure 6. Thickness of the Mississippian system and distribution of Mississippian Series in the Texas Panhandle. From Ruppel (1983).

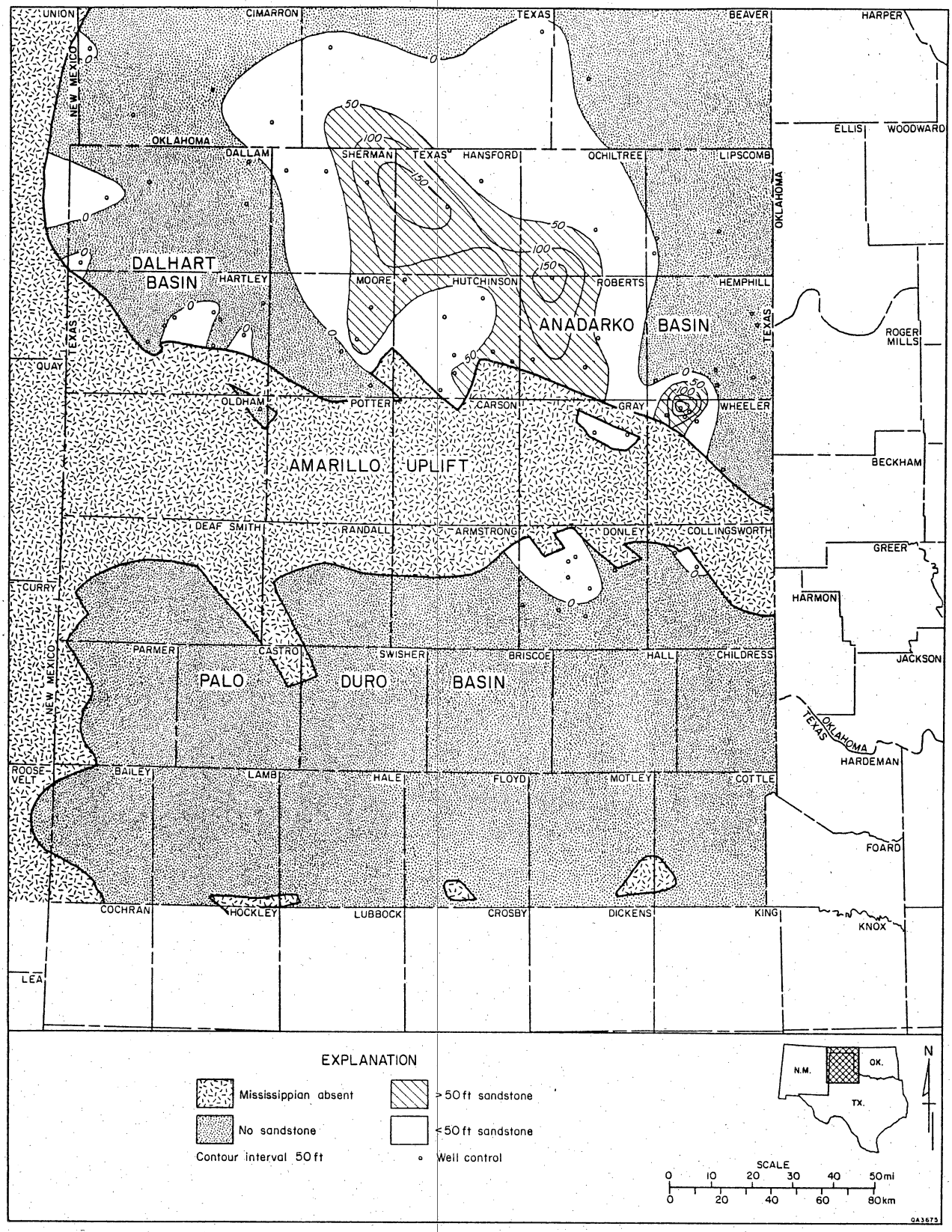


Figure 7. Thickness of basal Mississippian (Kinderhook) sandstones, Texas Panhandle. From Ruppel (1985).

Woodford Formation, which were deposited in Oklahoma and the northeastern Texas Panhandle.

Thin shale deposits form the base of the Mississippian section along the margins of the Texas Arch in the Palo Duro Basin (fig. 8). The distribution of these deposits suggests that they may be partly correlative with Kinderhook sandstones to the north. Basal siliciclastics are generally absent east of the Palo Duro Basin.

Osagean

Basal carbonate rocks in the Mississippian section in the Texas Panhandle are termed Osage (Ruppel, 1985). These rocks, which reach thicknesses of about 300 ft (91 m) in the eastern part of the Palo Duro Basin (fig. 9), directly overlie Precambrian basement on the Texas Arch and Kinderhook sandstones and shales along the eastern margin. For the most part, these rocks are argillaceous, cherty limestones and dolostones; dolostones are progressively more abundant to the west (Ruppel, 1985). The presence of interbedded skeletal grainstone and mudcracked and cryptalgally laminated mudstone documents alternating subtidal and intertidal to supratidal conditions.

Osage rocks in the eastern part of the Palo Duro Basin comprise spiculitic mudstone and carbonate turbidites that were deposited in a deep, quiet-water, outer ramp setting (Ruppel, 1984, 1985). Carbonate buildups are common especially in the Hardeman Basin to the east (Ahr and Ross, 1982). Available biostratigraphic data indicate that these Osage rocks are Meramecian rather than Osagean in age (Ruppel and Lemmer, 1986).

Meramecian

Rocks termed Meramec in the Texas Panhandle are primarily light-colored, skeletal grainstone and packstone deposited in high-energy, shoal environments (Ruppel, 1985).

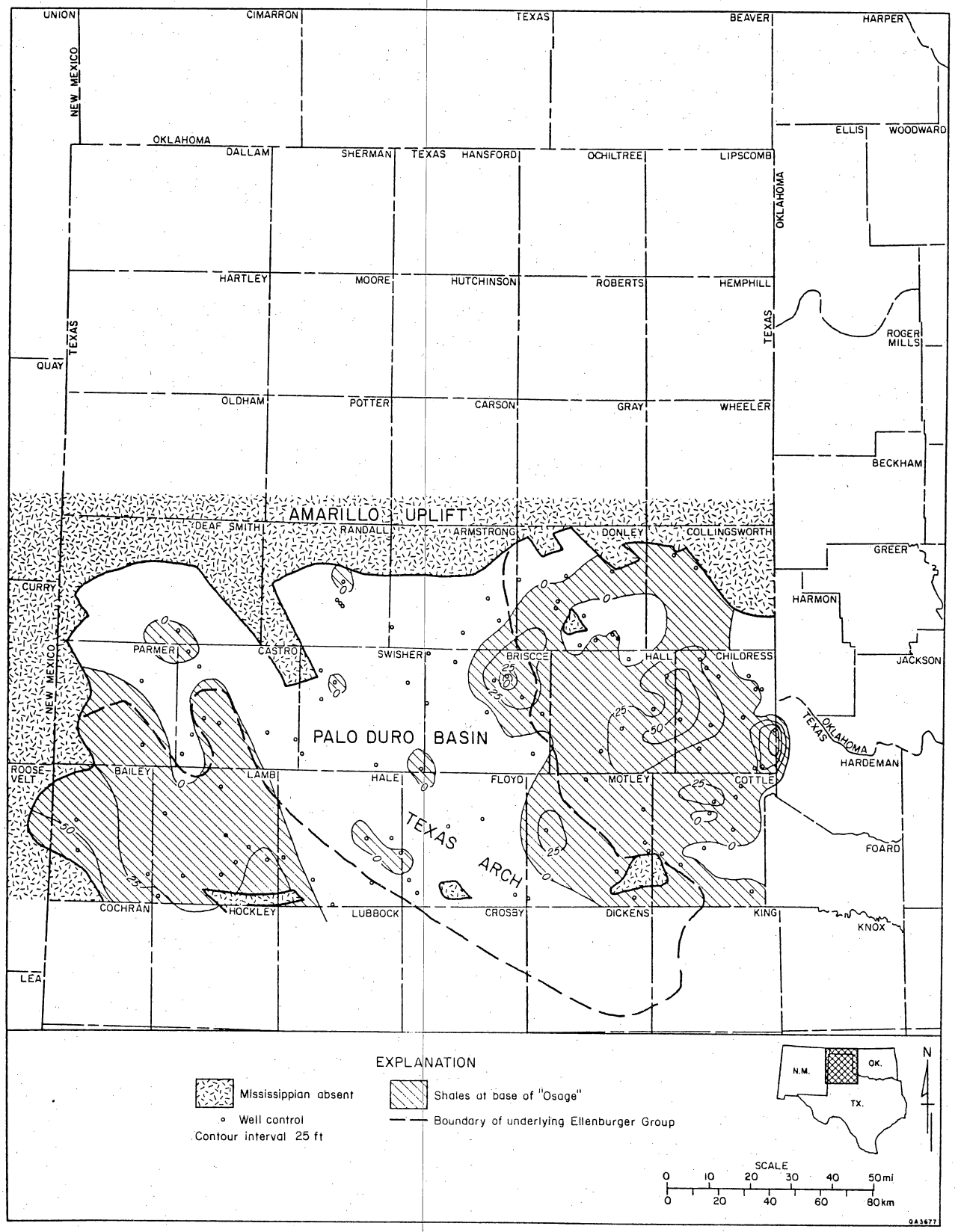


Figure 8. Thickness of basal Mississippian shales along the Texas Arch in the Palo Duro Basin. From Ruppel (1985).

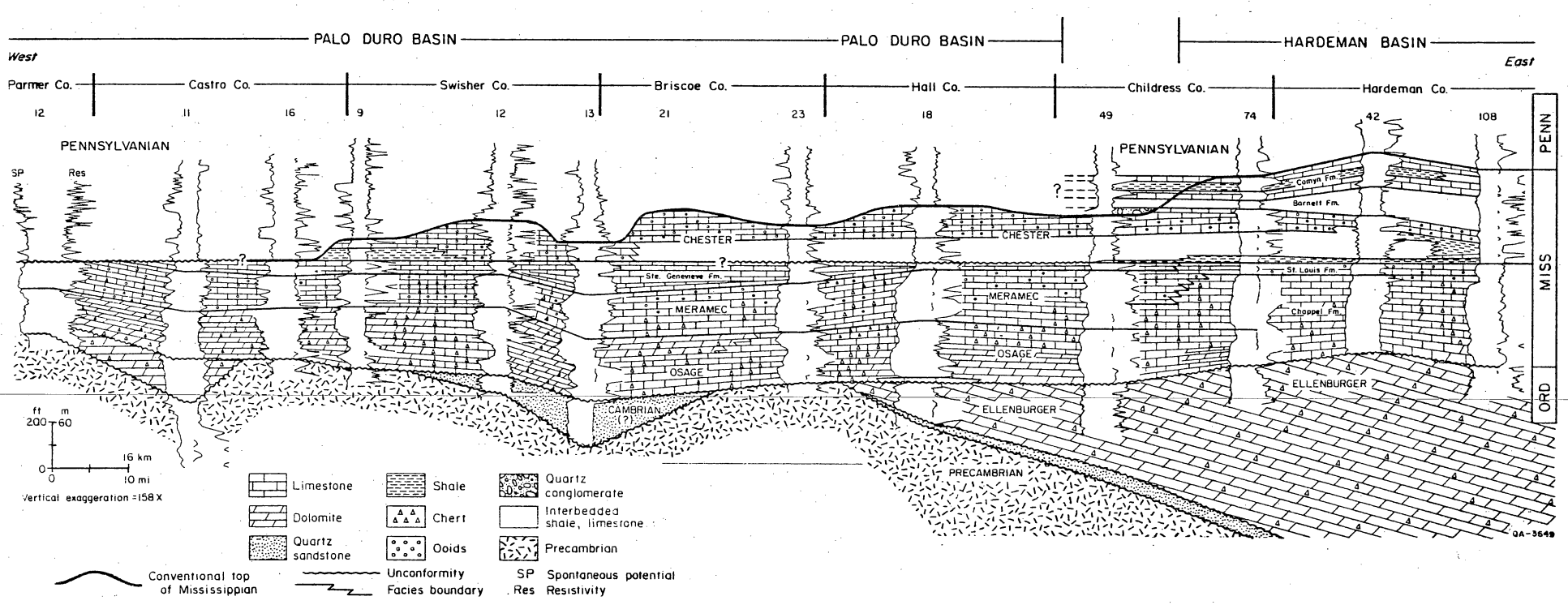


Figure 9. West-east cross section of Mississippian and older strata in southern Texas Panhandle. From Ruppel (1985).

Ooid grainstone is especially common in the upper part of the sequence, which was truncated by erosion along the axis of the Texas Arch prior to Chesterian deposition (Ruppel, 1985). Like the underlying Osage, Meramec rocks form a relatively thick and continuous deposit across most of the southern Panhandle (fig. 9).

Chesterian

Rocks assigned to the Chester Series largely comprise light-colored oolitic limestone and varicolored shales and sandstones (Ruppel, 1985). The distribution of the siliciclastic deposits indicates that they were derived from uplifted source rocks to the northwest, the Ancestral Rockies of New Mexico and Colorado. During the Chesterian, the Palo Duro Basin was characterized by shallow-water carbonate deposition punctuated by periodic influx of siliciclastics, predominantly from the northwest, which marked the beginning of late Carboniferous tectonic activity in the region (Ruppel, 1985). Much of the Chester section in the Panhandle was subsequently removed during continued tectonic uplift in the Pennsylvanian (figs. 7 and 10).

Pennsylvanian

Pennsylvanian rocks are present throughout nearly the entire Palo Duro Basin. Pennsylvanian and younger rocks in the Texas Panhandle exhibit a great deal more lateral lithologic diversity and variability than is apparent in underlying Lower Paleozoic deposits. This variability is the result of crustal instability and deformation associated with the Ouachita orogeny and the formation of the ancestral Rocky Mountains during the late Paleozoic collision between North America and Africa-South America (Budnik, 1986). Basins and structural highs produced by this deformation controlled depositional styles and patterns in the southern midcontinent during the Pennsylvanian and early Permian.

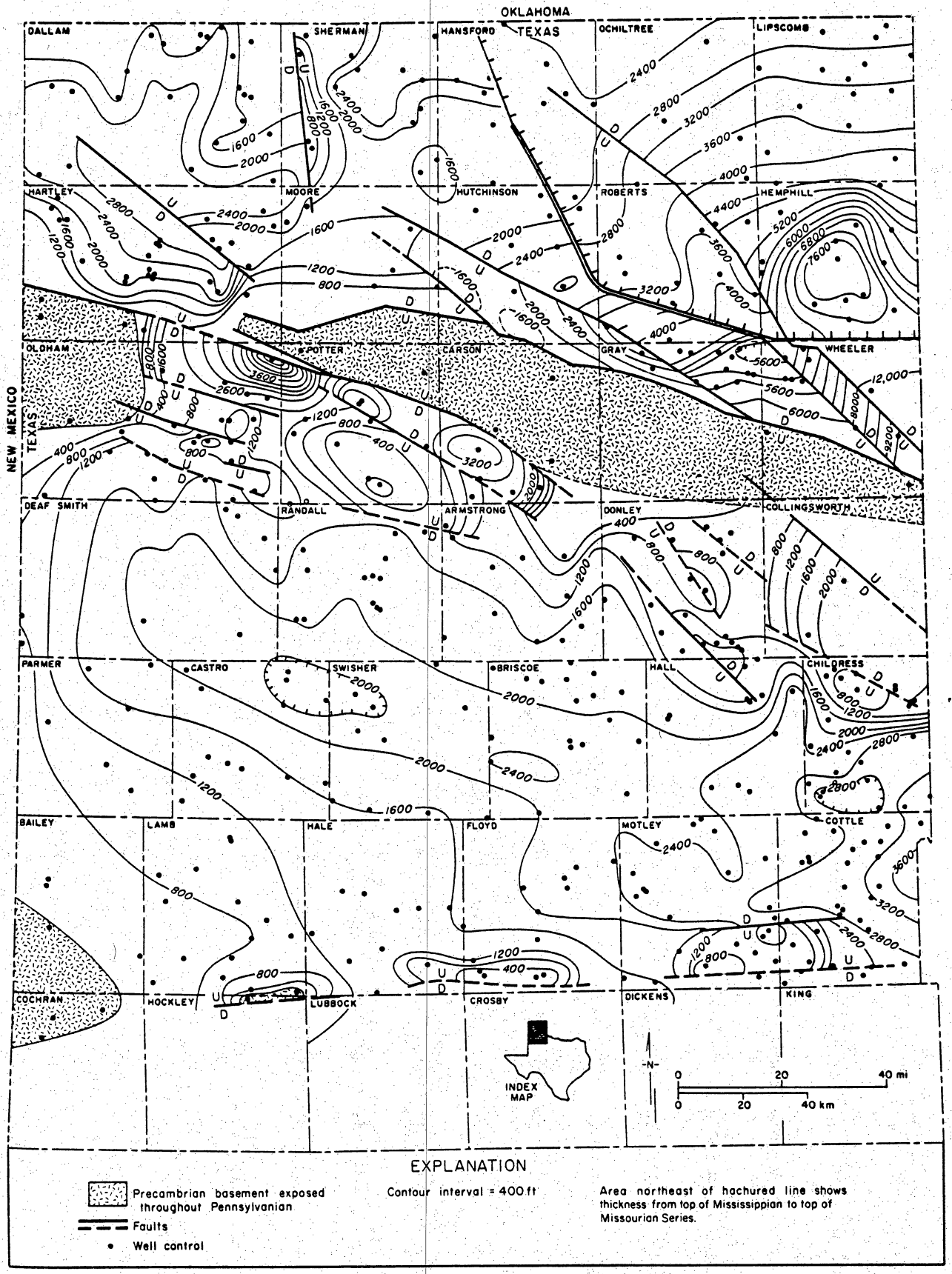


Figure 10. Thickness of Pennsylvanian rocks in the Texas Panhandle. From Dutton (1982).

Pennsylvanian deposits in the Palo Duro Basin constitute a thick sequence of rocks (fig. 10) comprising three basic lithologies: (1) coarse-grained siliciclastics (arkosic quartz sandstones and conglomerates, which are commonly referred to as granite wash), (2) dark-colored, organic-rich shales, and (3) light-colored, grain-rich carbonates (fig. 11). These rocks were deposited in (1) deltaic, (2) basinal, and (3) shallow platform environments, respectively (Dutton, 1980a; Handford, 1980a).

Lower Pennsylvanian rocks (Morrow and Atoka Series) are primarily characterized by interbedded coarse-grained siliciclastics, shales, and shallow-water carbonates. These rocks, which were deposited on a relatively low relief platform, are extensive across most of the basin. Coarse-grained clastics are increasingly dominant, however, toward the northern margin of the basin (fig. 11). These deposits were derived from uplifted blocks of Precambrian basement (Amarillo-Wichita Uplift, Bravo Dome) which began to shed sediment in the early Pennsylvanian (Dutton, 1980a). Thickness patterns indicate that these clastics were carried well into the basin by well-defined distributional systems that were probably part of high-constructive delta systems (fig. 12).

In the middle and upper parts of the Pennsylvanian sequence, the major rock types are much more clearly segregated into discrete geographic areas as a result of intensifying deformation and structural and topographic differentiation of the area. Shales are most abundant in the central part of the basin along a north-south to northwest-southeast-trending axis. Carbonates are abundant in the eastern and western parts of the basin, on either side of the axis of shale deposition (fig. 11). The thickest accumulations of carbonate occur in elongate trends near the margins of the shale depocenter (fig. 13). The distribution of dolomite closely parallels these thickness trends (fig. 14). Upper Pennsylvanian coarse-clastics are more restricted to the northern basin margin compared with the lower Pennsylvanian (fig. 15).

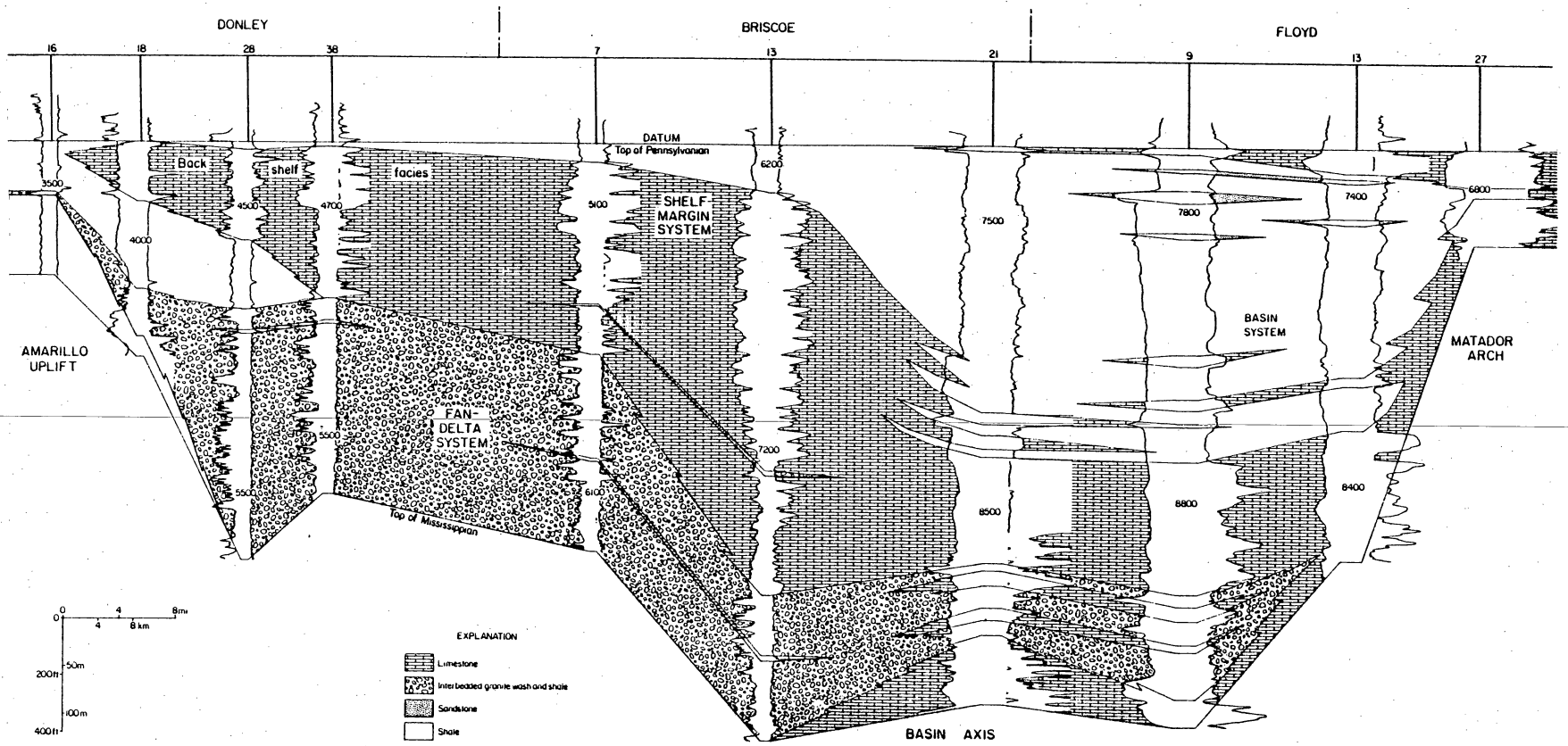


Figure 11. North-south cross section through the Palo Duro Basin illustrating major Pennsylvanian lithofacies. From Dutton (1980a).

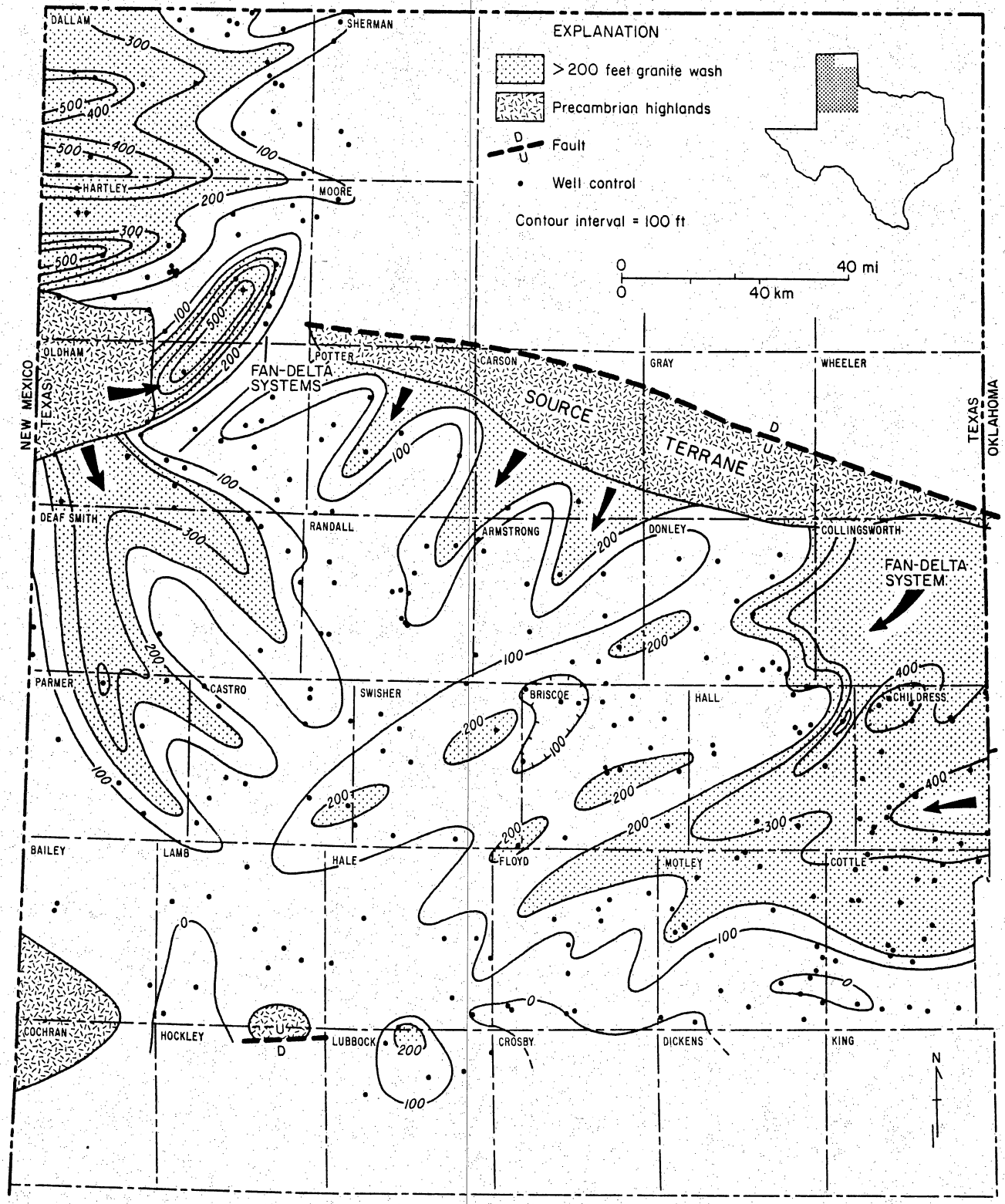


Figure 12. Thickness of coarse-grained siliciclastics in the lower Pennsylvanian. From Dutton (1980a).

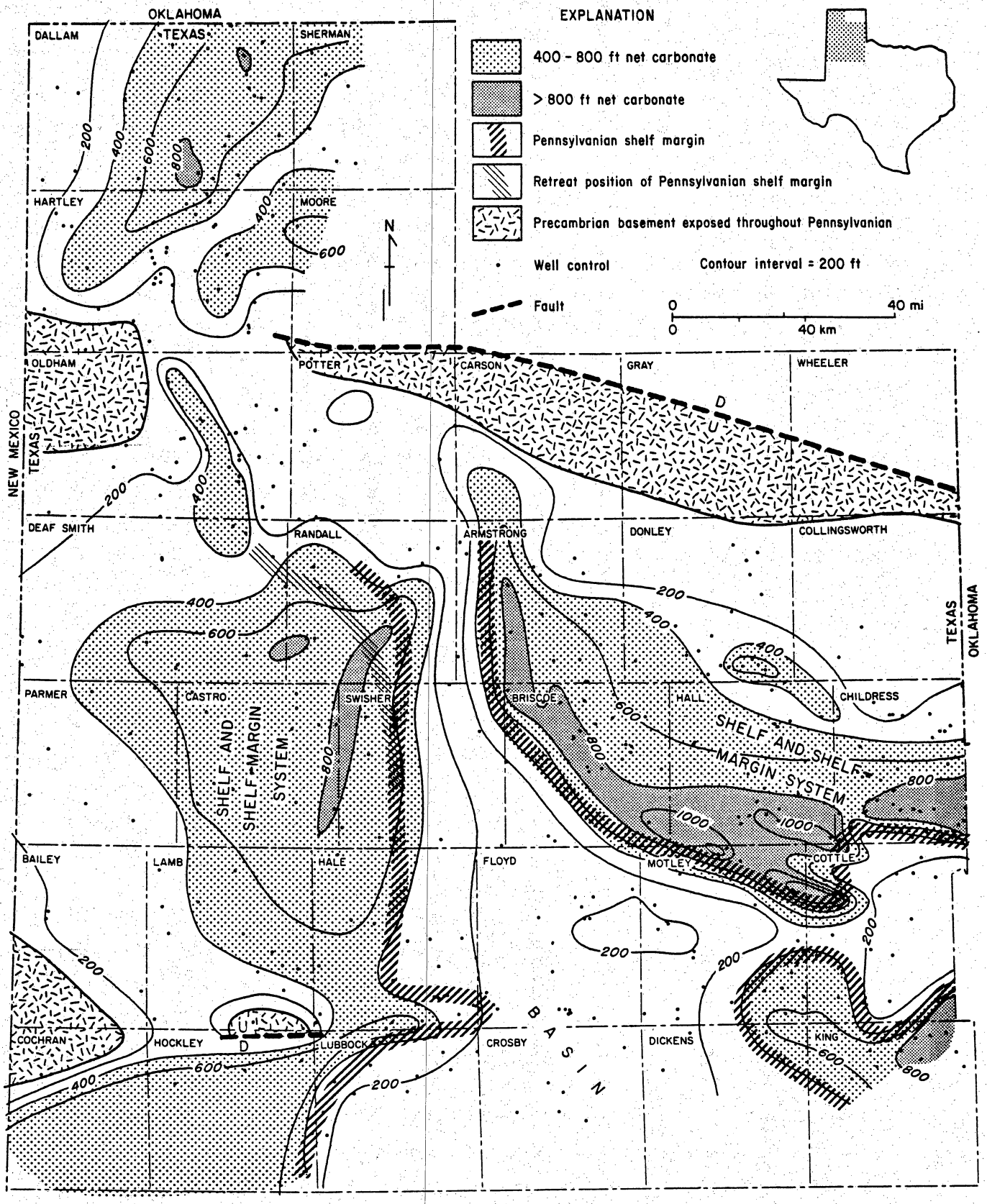


Figure 13. Distribution of platform carbonates in the upper Pennsylvanian. From Dutton (1982).

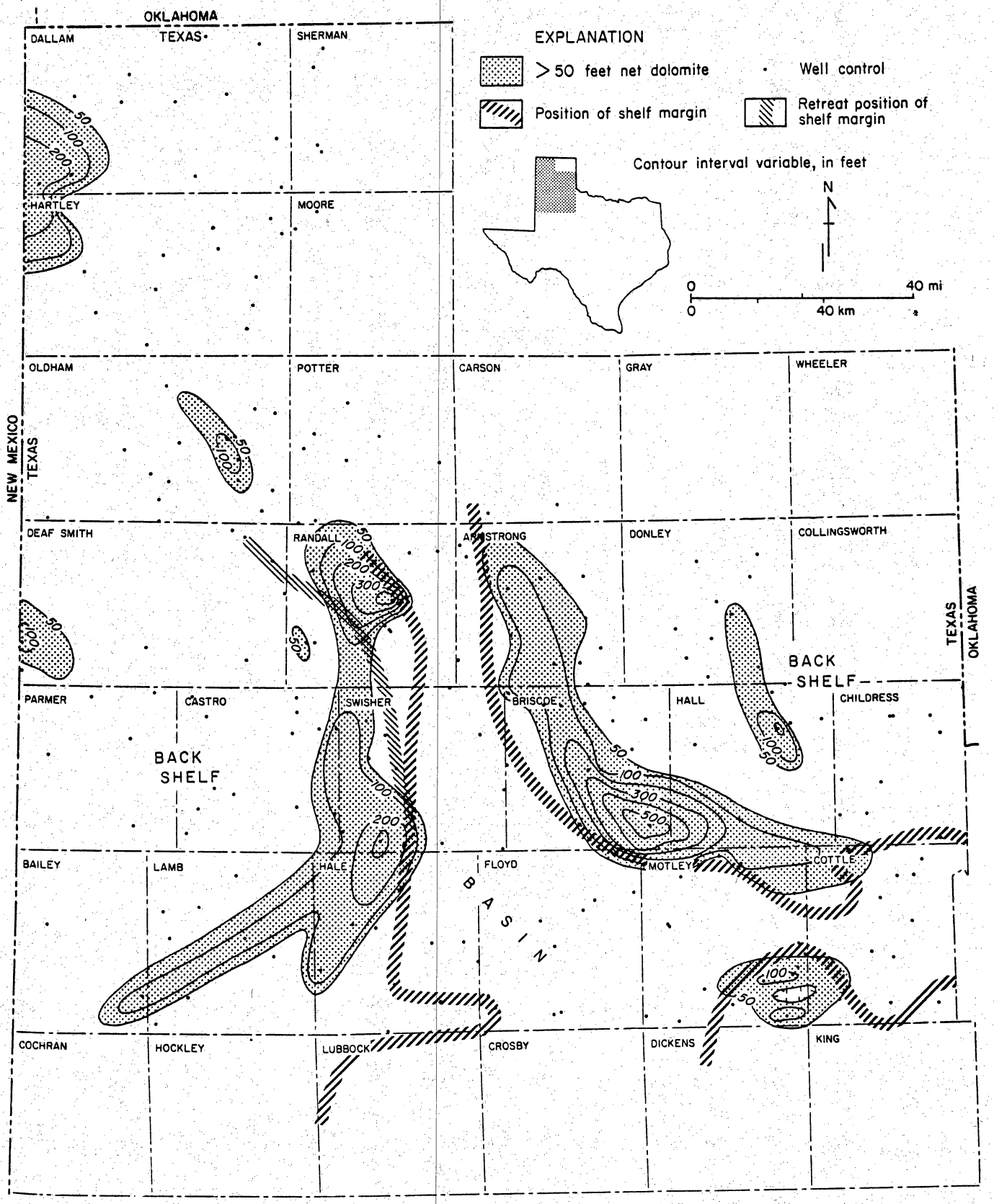


Figure 14. Thickness of dolomite in the upper Pennsylvanian. From Dutton (1982).

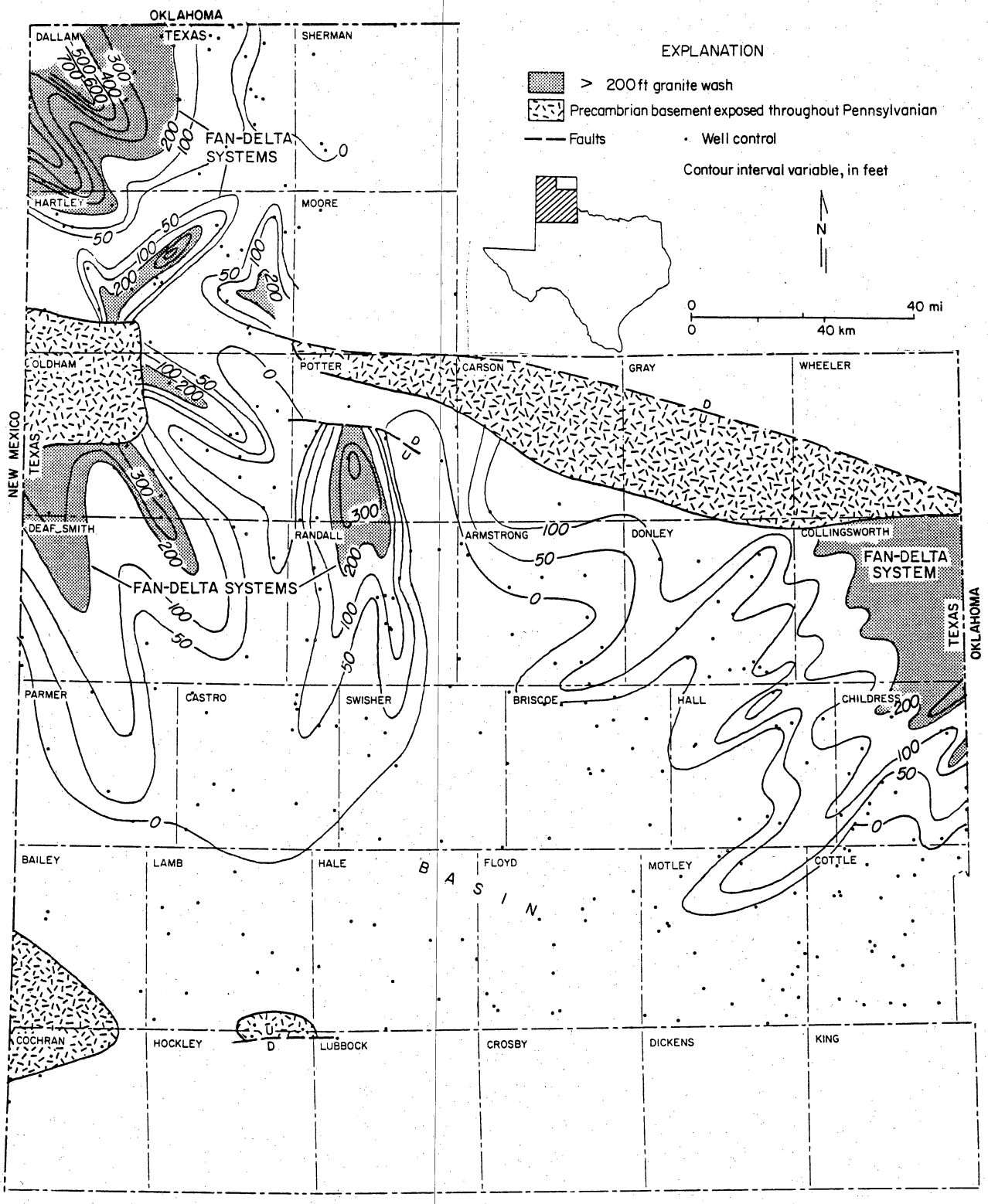


Figure 15. Thickness of coarse-grained siliciclastics in the upper Pennsylvanian. From Dutton (1982).

Dutton (1980a) and Dutton and others (1982) interpreted these middle and upper Pennsylvanian facies trends to represent the development of a deep-water basin that was flanked by elongate carbonate buildups. The depth of this basin is poorly constrained. Rose (1986b) suggested water depths on the order of 1,000 ft (305 m) at the time of maximum relief during the Late Pennsylvanian. Crevello and others (1985) and Dutton (personal communication, 1988) interpreted a somewhat shallower depth (approximately 600 ft [200 meters]).

The composition of the carbonate buildups forming the platform margins along the basin is poorly known. Crevello and others (1985) described carbonate debris flows from along the eastern margin of the basin. These rocks contain phylloid algae, *Tubiphytes*, encrusting tubular foramifera, and a variety of shallow marine organisms suggesting that they were derived from the platform margin. The composition of these rocks is similar to that of Pennsylvanian and lower Permian rocks elsewhere in the southern midcontinent (see Wilson [1975] for a summary).

Permian

Wolfcampian

Lowermost Wolfcamp facies reflect a continuation of depositional styles begun in the middle and late Pennsylvanian. Thickness trends (fig. 16) and the distribution of shales and flanking platform carbonates suggests, however, that the basin axis trended more northward and extended further northwestward than during the Pennsylvanian (fig. 17). The eastern platform margin prograded progressively westward during the Wolfcampian due to the continued influx of coarse-grained clastics from the east and the associated deposition of aggrading carbonate bank systems (figs. 18 and 19; Handford and Fredericks, 1980). The basin began to be filled with terrigeneous and carbonate detritus early in the Permian. According to Handford and Fredericks (1980)

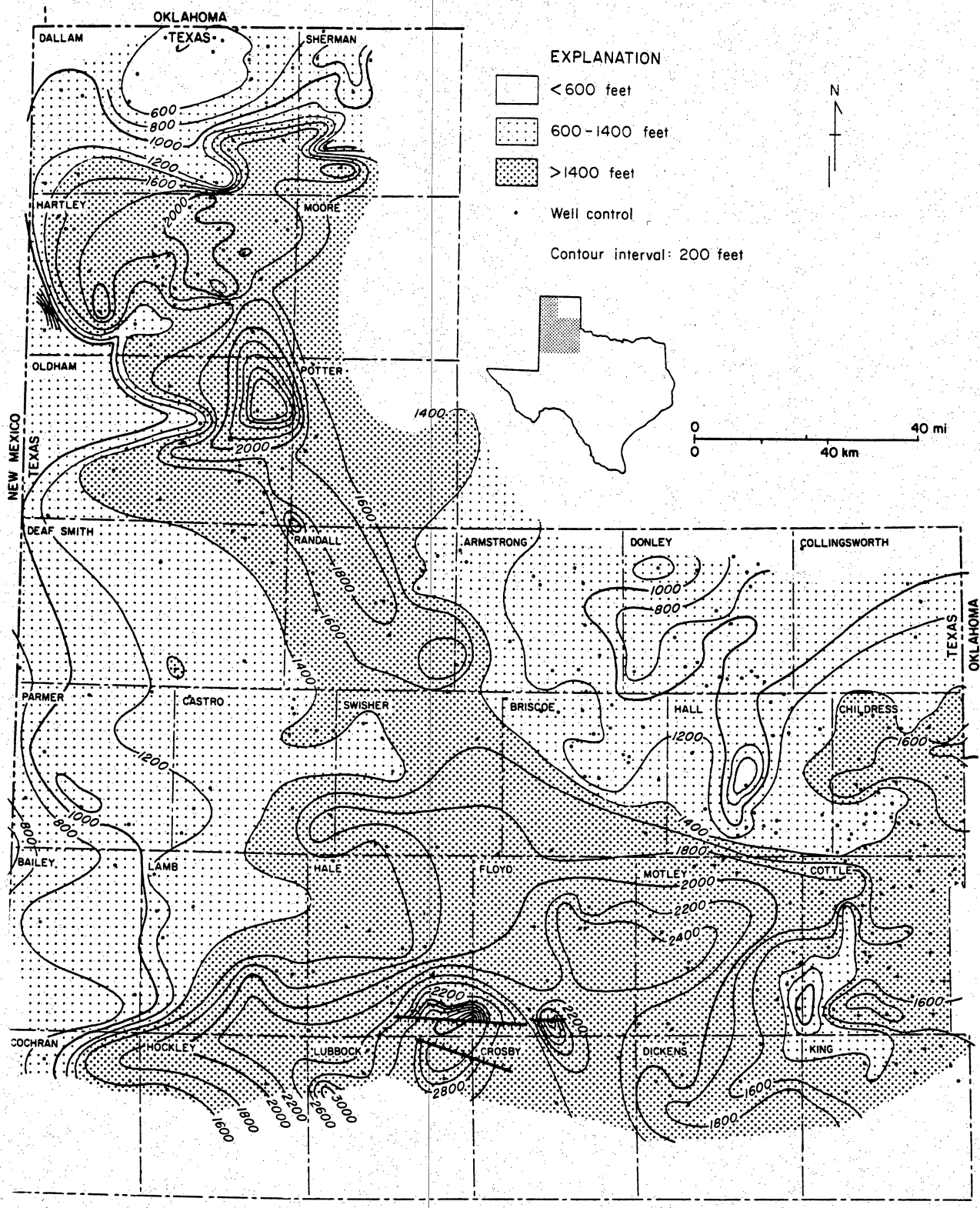


Figure 16. Thickness of Wolfcampian strata in the Texas Panhandle. From Dutton (1982).

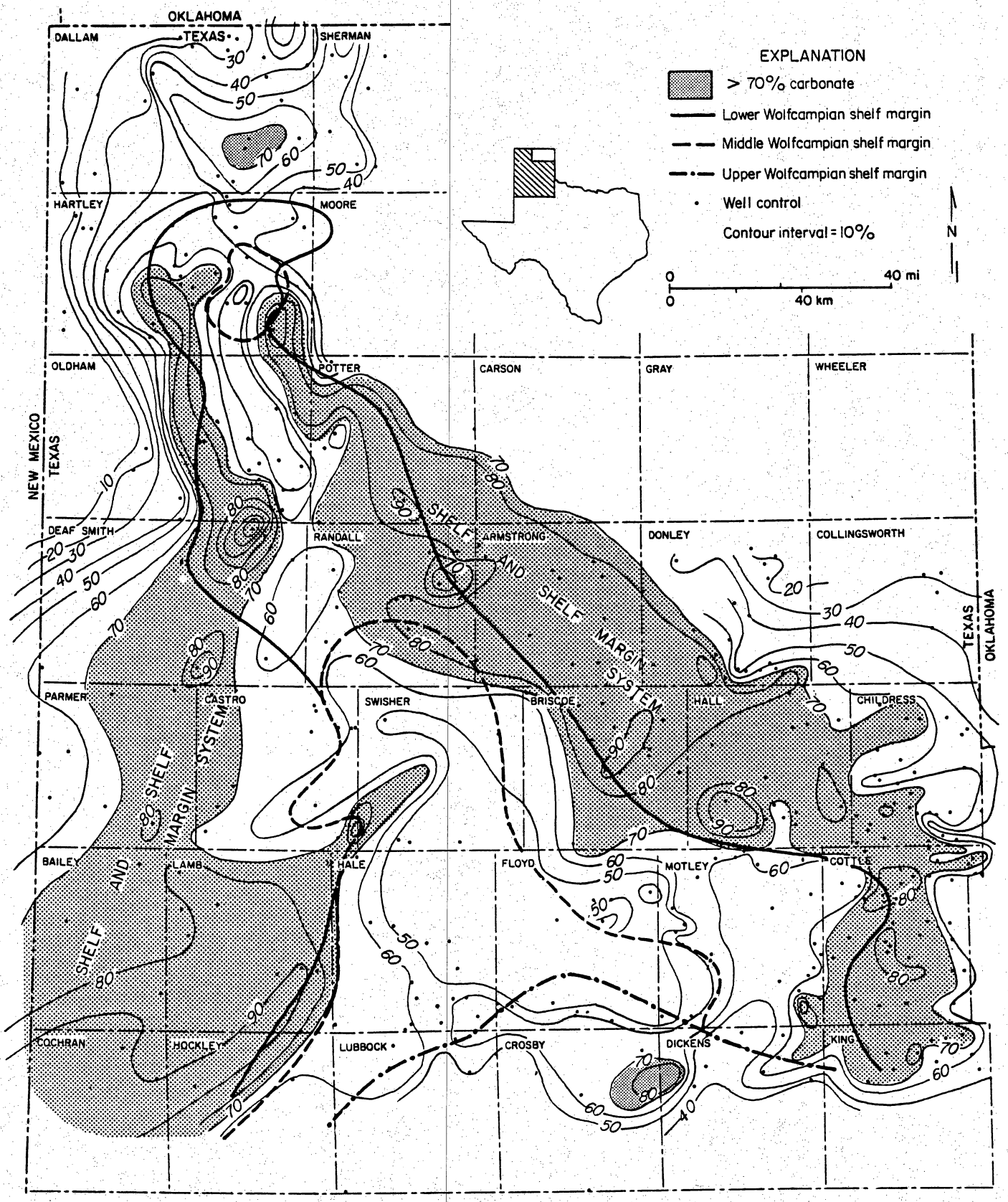


Figure 17. Distribution of platform carbonates in the lower Permian (Wolfcampian). From Dutton (1982).

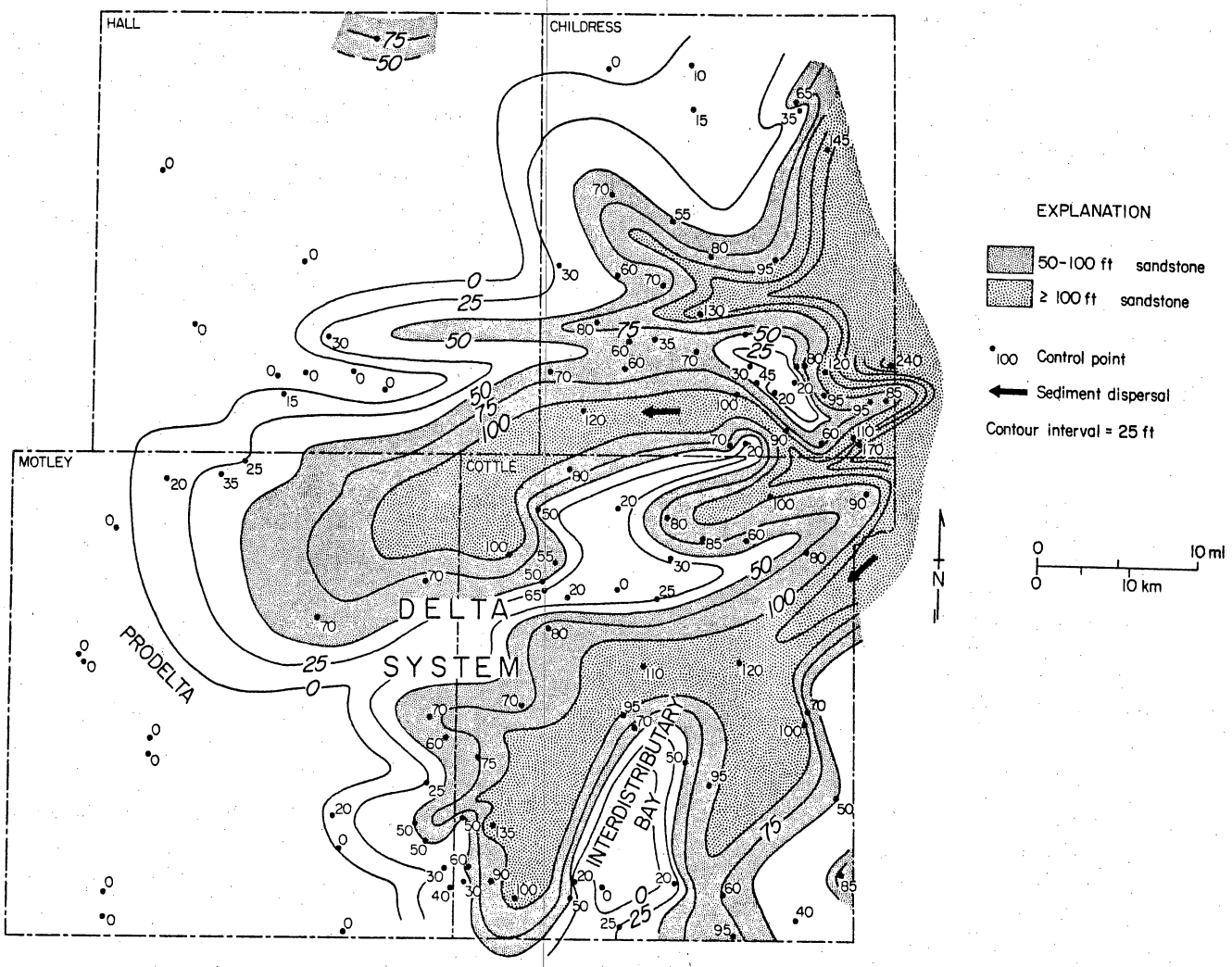


Figure 18. Thickness trends of Wolfcampian deltaic sandstones in the eastern Palo Duro Basin. From Handford (1980a).

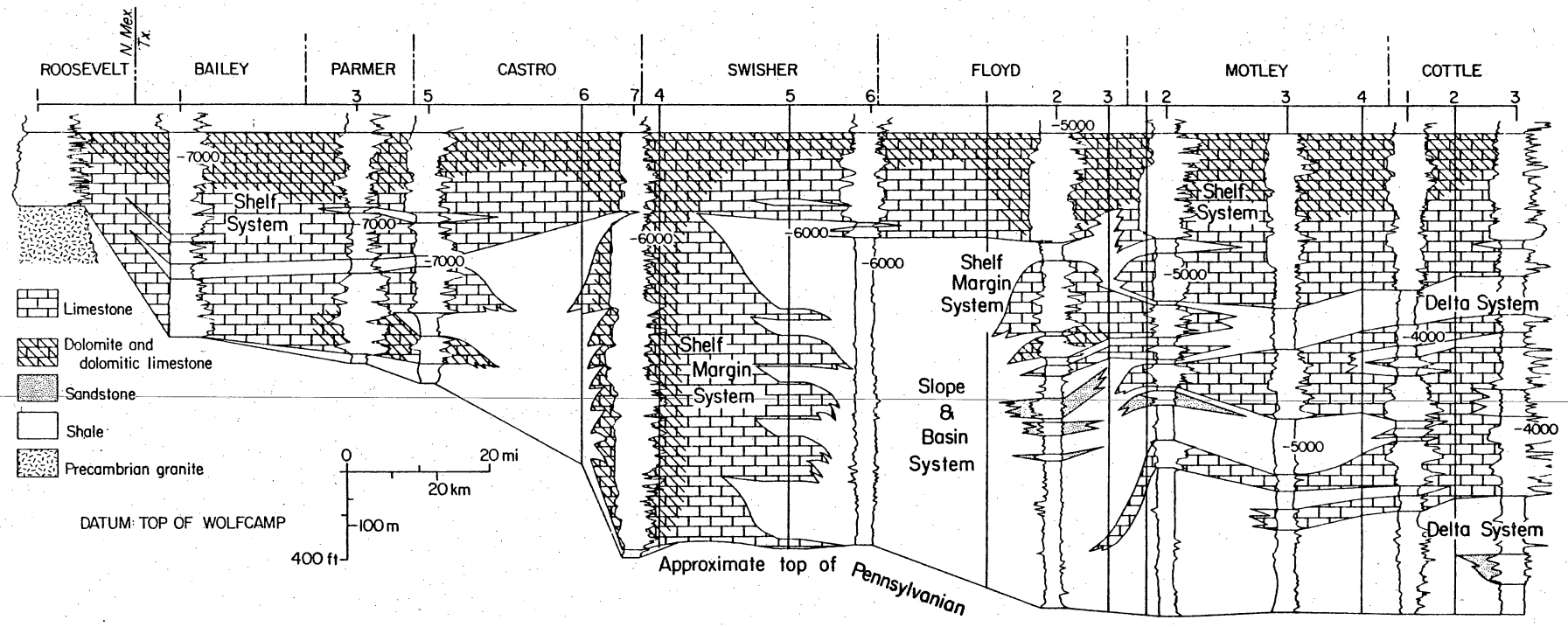


Figure 19. East-west cross section through the southern Texas Panhandle showing the distribution of Wolfcamp facies. From Handford (1980a).

platform to basin relief was probably no more than 400 ft (122 m) during the Wolfcampian Epoch. Although there is evidence of erosion at the base of the Wolfcamp section along the margins of the Palo Duro Basin, most of the area is characterized by continuous deposition across the Pennsylvanian/Permian boundary (Conti and others, 1988). Preliminary core study indicates that platform margin carbonates are similar to those described for the Pennsylvanian.

Upper Wolfcampian sediment documents the filling of the deeper-water basin and the development of a very low relief carbonate platform across the entire southern Panhandle. These rocks are composed primarily of shallow-water marine carbonates (ooid grainstone, skeletal wackestone, and mudstone) and greatly reduced quantities of both coarse- and fine-grained siliciclastics. The upper Wolfcamp section is extensively dolomitized and contains anhydrite which increases in abundance upward through the section (Handford, 1980; Conti and others, 1988). Dolomite is also pervasive in platform margin carbonates flanking the central basin (fig. 20). The sequence represents an upward-shallowing trend from open, shallow-water marine to supratidal.

Leonardian to Ochoan

By the end of Wolfcampian deposition, the southern Texas Panhandle area was a flat, relatively featureless platform. Deposition throughout the remainder of the Permian was predominantly a function of variations in supply of terrigenous clastics and changes in water depth. Structural controls (principally differential subsidence) were restricted to minor changes in paleoenvironmental setting and sediment accumulation rate due to differential subsidence along the extreme southern (Matador Arch) and northern (Amarillo Uplift) margins of the basin.

The post-Wolfcampian Permian section in the Palo Duro Basin comprises a thick sequence (fig. 21) of cyclic deposits of alternating shallow-water carbonates, evaporites, and siliciclastics deposited in distinct transgressive-regressive events. Each of these

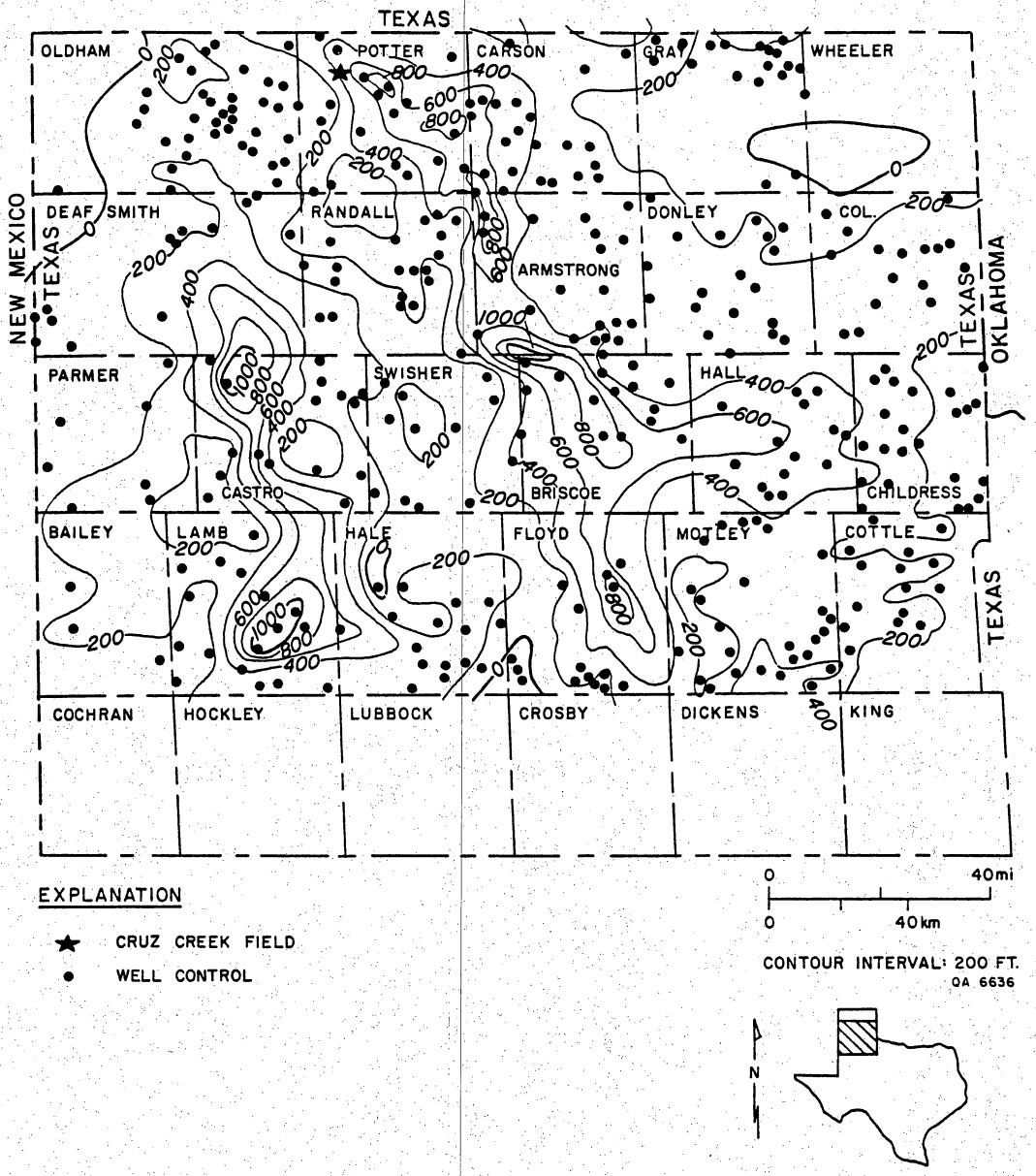


Figure 20. Thickness of dolomite in the lower Permian (Wolfcamp). From Conti and others (1988).

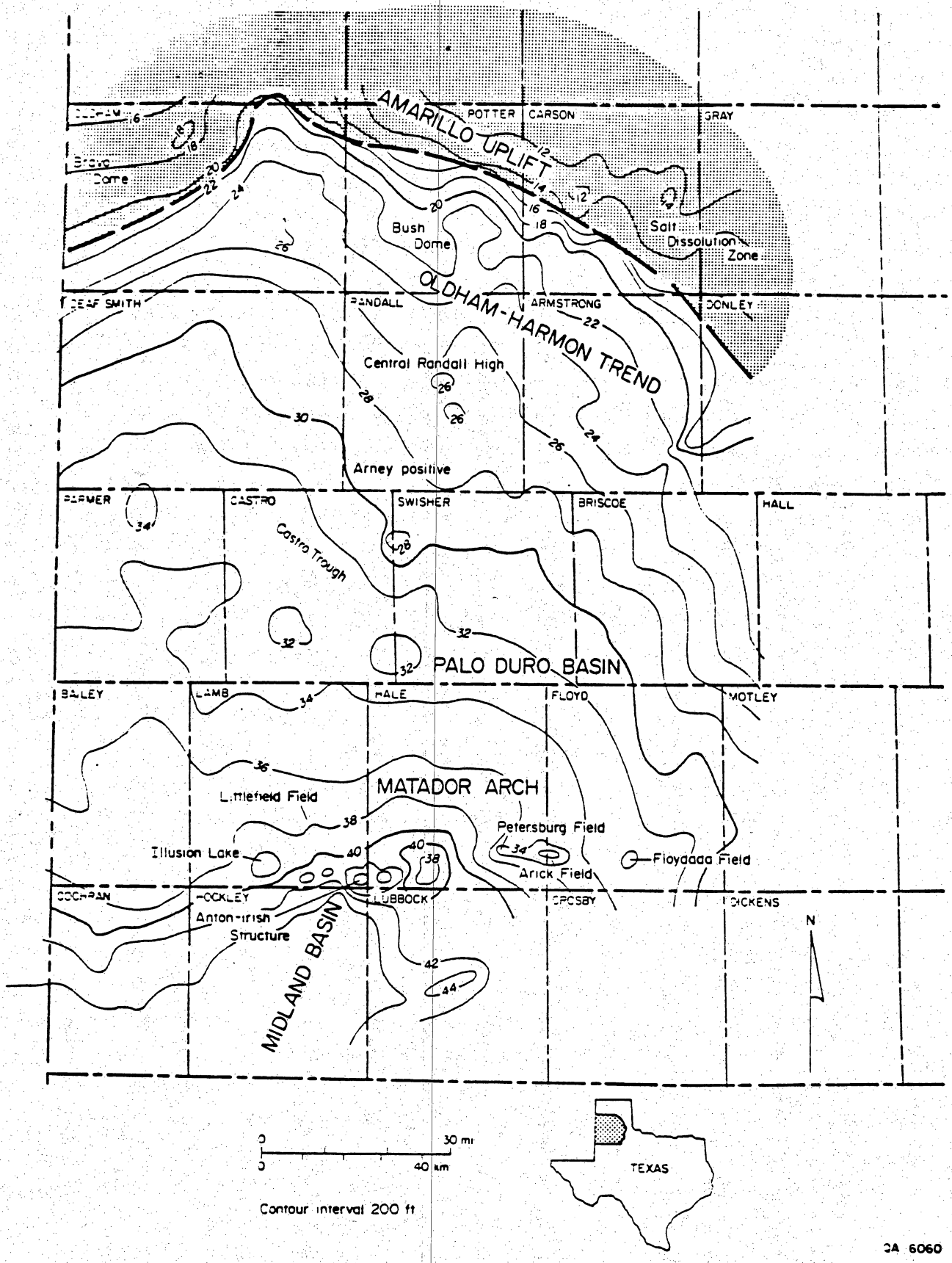


Figure 21. Thickness of middle and upper Permian (top of Tubb Formation to top of Alibates Formation) strata in the Palo Duro Basin. From Budnik (1988).

sequences represents a response to variations in siliciclastic influx (predominantly from the north) and relative sea level change. Seven major cycles can be recognized.

Leonardian Series

1. Wichita Group/Red Cave Formation
2. lower Clear Fork/Tubb Formation
3. upper Clear Fork/Glorieta Formation

Guadalupian Series

4. San Andres Formation/Queen-Grayburg Formation
5. Seven Rivers Formation/Yates Formation

Ochoan Series

6. Salado Formation/Tansill Formation
7. Alibates Formation/Dewey Lake Formation

Each major cycle is a lithologic couplet composed of a lower, more basinward sequence and an upper, more landward sequence. The lower and upper parts of these couplets, thus, represent the distal and proximal ends of the facies tract. In lower couplets, distal facies are marine carbonates (mostly dolomitized). The Wichita Group (fig. 22), which forms the base of the lowermost couplet, for example, comprises peritidal to supratidal dolostones (mudstone and peloidal grainstone) and nodular and bedded anhydrite (Green, 1985; Conti and others, 1988). Lower parts of each couplet typically grade upward and updip into progressively more evaporitic deposits (fig. 23). In upsection couplets, distal facies comprise progressively more restricted deposits. The Salado Formation, which forms the base of the Salado/Tansill couplet, is primarily halite (McGillis and Presley, 1981).

The upper member of each couplet is composed of terrigenous siliciclastic deposits, principally red beds. In the Palo Duro Basin these deposits comprise primarily fine-

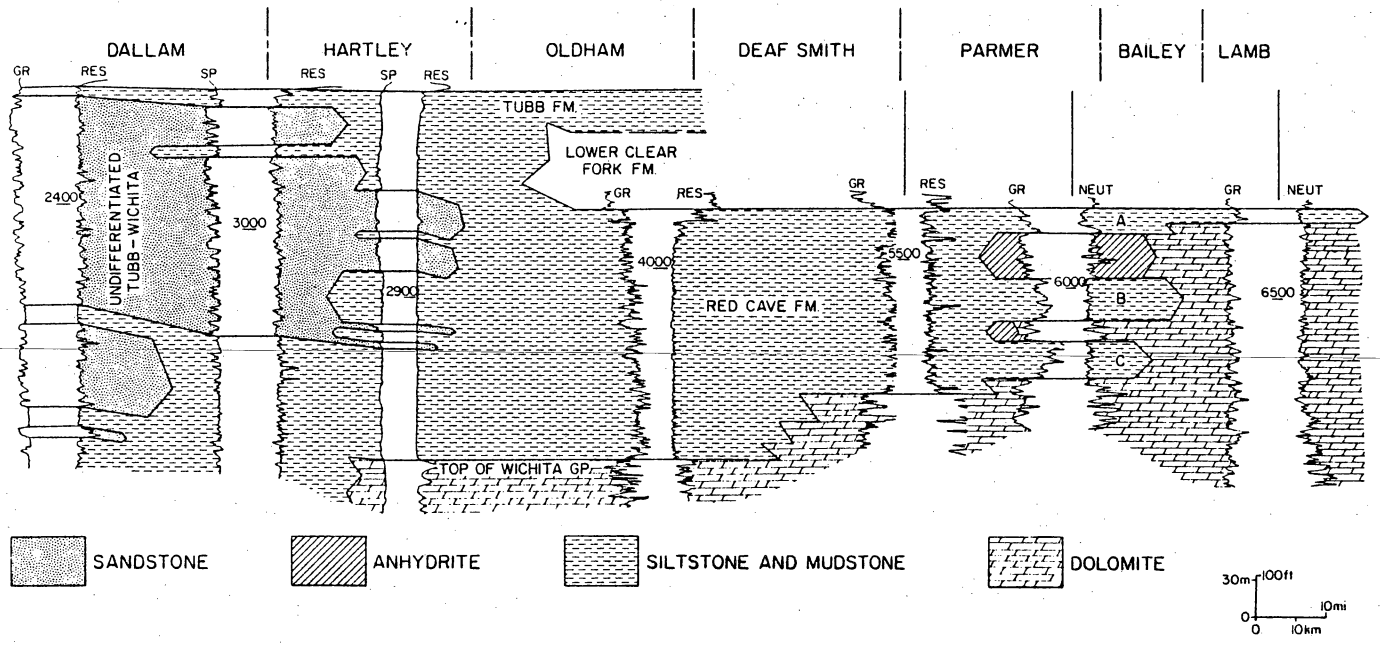


Figure 22. Cross section depicting Red Cave and Wichita rocks in the Palo Duro Basin. From Handford and Fredericks (1980).

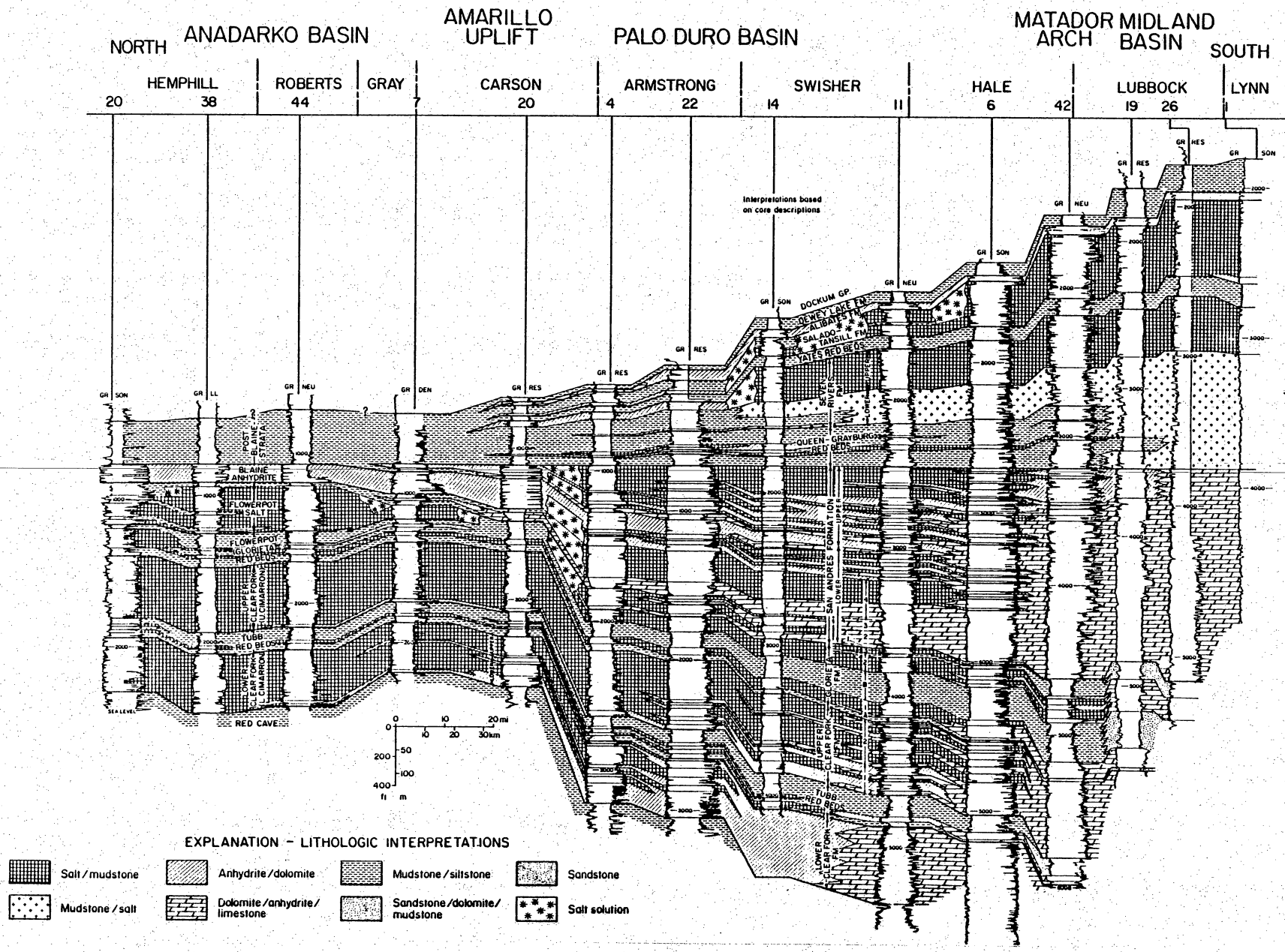


Figure 23. Cross section illustrating Leonardian to Ochoan deposits in the Palo Duro Basin. From Presley (1987).

grained sandstone and coarse siltstone (Handford and Fredericks, 1980; McGillis and Presley, 1981; Presley and McGillis, 1982; Presley, 1987; Nance, 1988).

Each major cyclic sequence is itself composed of numerous second-order cycles. Like the major cycles, these cycles, which are most easily recognized in the lower, non-clastic parts of the couplet, generally consist of alternations of basal carbonate, anhydrite, and halite. The composition varies, however, depending on whether deposition occurred in a more proximal or more distal part of the facies tract. Lower Leonard cycles contain rare halite whereas Ochoan cycles contain rare carbonates (Presley, 1987).

Second-order cycles have been recognized throughout the Leonard-Ochoan section and range in thickness from about 20 to 280 ft (6 to 85 m). They are best-developed in the San Andres Formation where at least 29 such cycles have been reported (Fracasso and Hovorka, 1986). The San Andres has received especially close scrutiny because it contains salt (halite) beds identified as a possible repository by the U.S. Department of Energy.

An ideal San Andres cycle consists of (1) basal siliciclastic mudstone, (2) shallow water skeletal limestone, (3) dolostone, (4) nodular anhydrite, (5) bedded anhydrite, and (6) halite (fig. 24; Fracasso and Hovorka, 1986). Cycles in the lower San Andres are complete and relatively thick (70 to 300 ft [21 to 91 m]). Upper San Andres cycles are typically more incomplete (halite is commonly absent) and thinner (20 to 80 ft [6 to 25 m]).

Leonard to Ochoan cyclic sequences in the Palo Duro Basin are the result of the migration of nearshore marine to terrestrial depositional tracts across the relatively flat platform that characterized the southern Texas Panhandle during the middle and Late Permian (Presley, 1987). These cyclical facies shifts document seven major rises and falls in relative sea level during the middle and late Permian. Following each rise, more basinward depositional environments moved rapidly landward. Subsequent

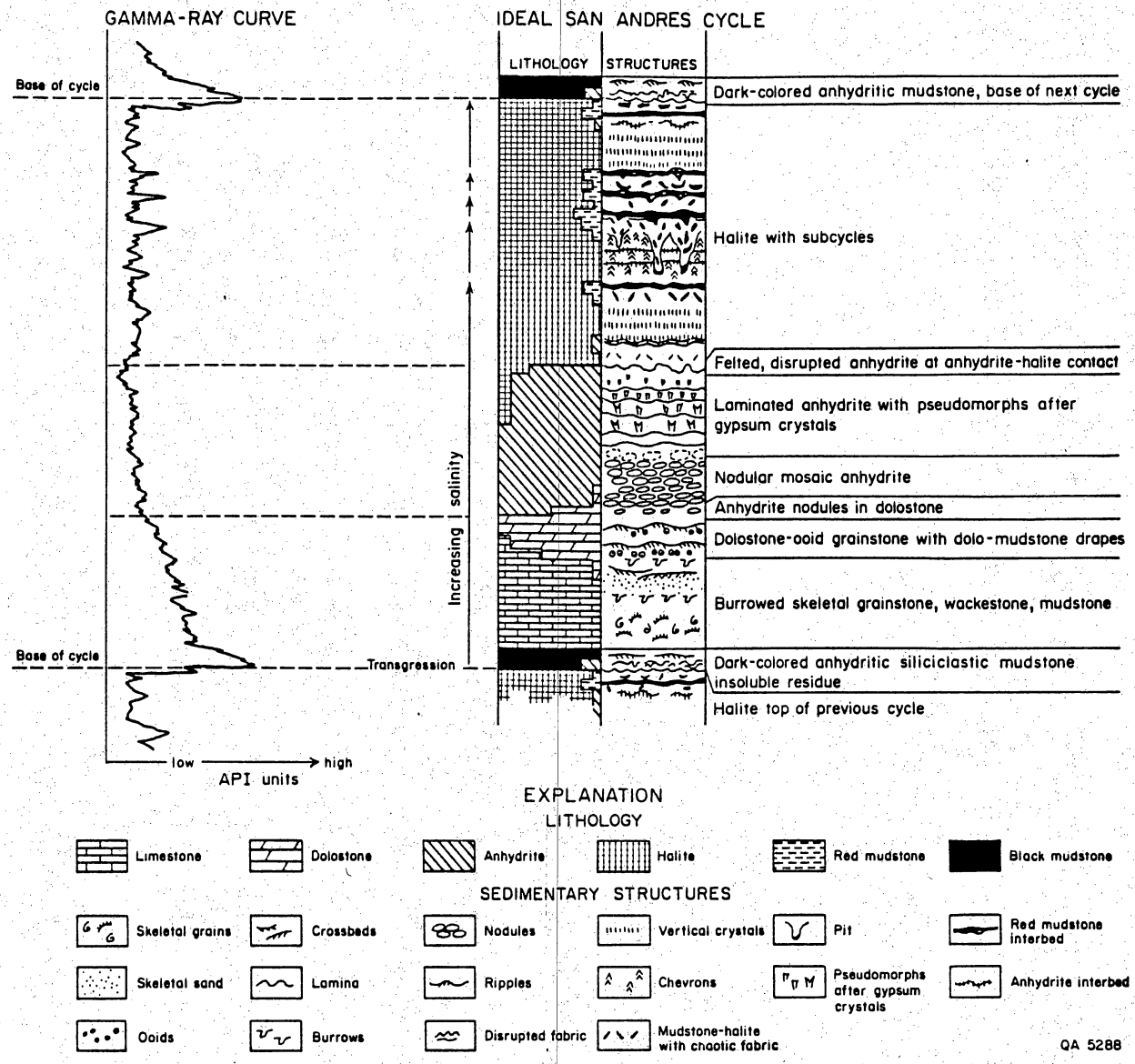


Figure 24. Vertical facies sequence in an ideal second-order cycle in the San Andres Formation. From Fracasso and Hovorka (1986).

progradation and aggradation of terrigenous siliciclastics, evaporites (halite and anhydrite, and carbonates) resulted in a general shallowing trend and basinward shift in paleoenvironments through time.

Carbonates, which form the base of many cyclic sequences, were deposited in a shallow water, inner platform setting (fig. 25). Vertical and lateral (updip) trends in mineralogy (limestone to dolostone) and texture (skeletal grainstone/packstone to peloidal mudstone) indicate a progressive increase in salinity from normal marine to hypersaline (Fracasso and Hovorka, 1986).

Evaporites accumulated updip from carbonates in shallow brine seas formed by a progressive increase in restriction and evaporative concentration of sea water brine (Presley and McGillis, 1982; Hovorka, 1987; Presley, 1987). Anhydrite, precipitated as gypsum, formed on the outer, more seaward reaches of the brine sea while halite precipitated updip in more restricted conditions (Fracasso and Hovorka, 1986; Hovorka, 1987). Fluctuations in subsidence, and the influx of marine and meteoric waters produced minor cyclicity in both gypsum- and halite-precipitating environments. Incomplete cycles in the San Andres are in large part the result of partial to complete dissolution of halite caused by the influx of unsaturated brines prior to the beginning of the next cycle. Basal siliciclastic mudstone, which is reworked residue of terrigenous clay previously contained in the halite, is in many cases the only record of the previous existence of halite in these cycles.

Terrigenous clastics accumulated at the proximal end of the paleoenvironmental tract as mixed eolian and fluvial deposits (Presley and McGillis, 1982; Presley, 1987; Nance, 1988). Deposition of these sediments is probably a response to drops in relative sea level.

The geographic extent of each of these paleoenvironmental systems and their patterns of migration through time was a function of basin geometry, subsidence, and relative sea level change. Leonardian first-order cycles document a southward migration

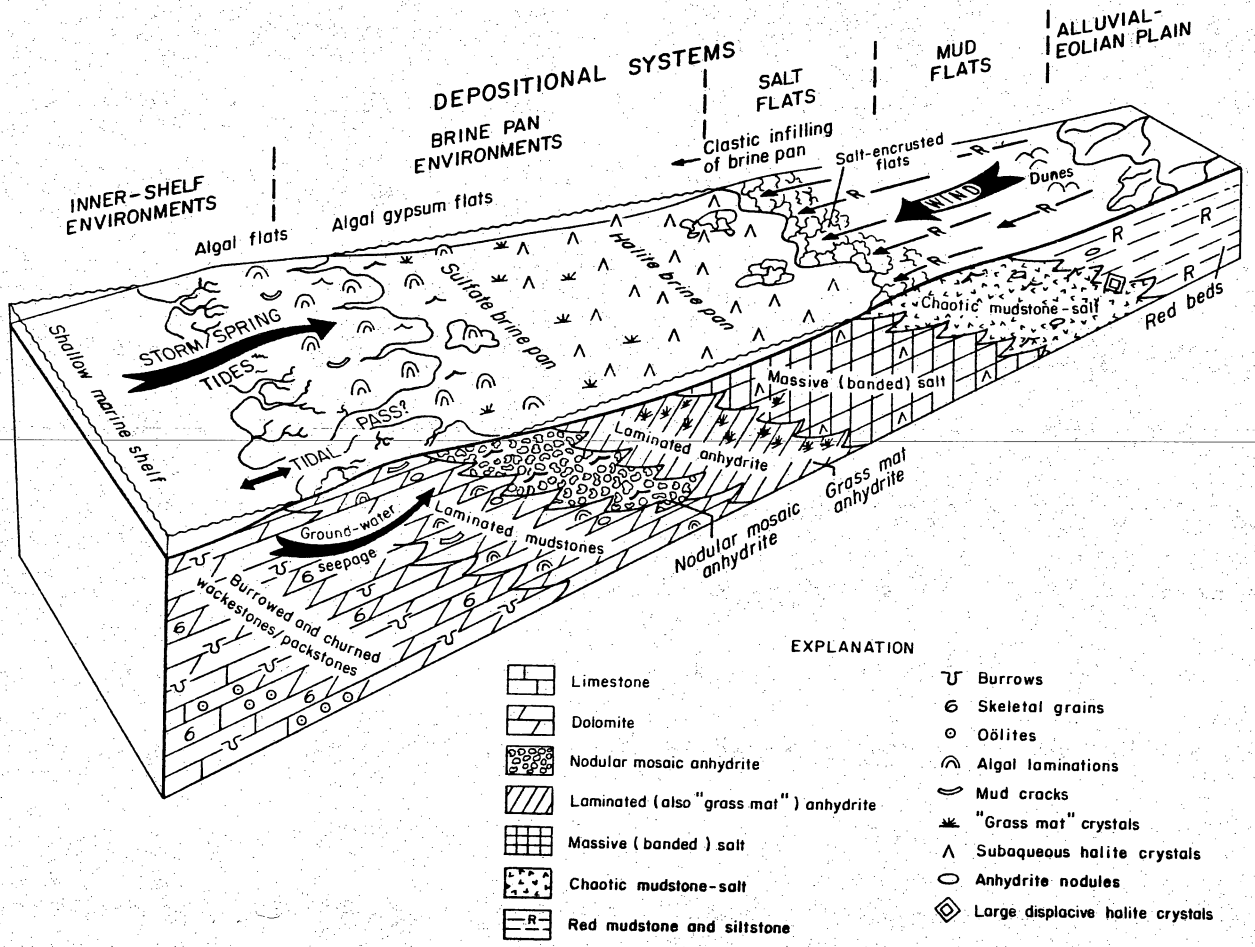


Figure 25. Paleoenvironmental reconstruction of Palo Duro Basin area during the post-Wolfcampian Permian. From Presley and McGillis (1982).

of shallow water carbonate and evaporite-precipitating environments (fig. 25). This migration, which was caused by increased restriction and hypersalinity in the basin, was punctuated by episodes of terrigenous clastic influx (Red Cave, Tubb, and Glorietta Formations) probably triggered by relative sea level drop.

Most carbonates in the sequence are dolostone and, with the exception of the Wichita Group, most were deposited in normal marine conditions. Bein and Land (1982) determined that San Andres dolostones are the product of the reflux of hypersaline brines following southward migration of evaporite-precipitating environments over previously-deposited carbonates (limestone).

Paleozoic Basin History

Early Paleozoic

Early Paleozoic deposition in the Palo Duro Basin area probably began in the Late Cambrian associated with downwarping of the southern midcontinent following rifting along the Southern Oklahoma Aulocogen (Hoffman and others, 1974). Thin basal Paleozoic arkose and quartzose sandstone constitute initial transgressive deposits that accumulated as much of the region subsided. With continued subsidence, the overlying Ellenburger Group carbonates accumulated in a shallow-water, epeiric sea that spread across much of central North America during the early Ordovician (Ross, 1976). Although some have suggested that the Texas Arch was emergent through the southern Panhandle area during the Ellenburger deposition (Adams, 1954), there is no direct evidence to support this contention. Instead, it appears likely that Ellenburger carbonates were deposited across the entire Texas Panhandle and that the present distribution of these carbonates is a function of subsequent erosion along the Texas Arch.

Middle Paleozoic Tectonism

The middle Paleozoic history of the southern Texas Panhandle is poorly understood due to the nonpreservation of rocks of this age. It is uncertain whether the absence of Silurian and Devonian rocks from the Palo Duro Basin is due to nondeposition or erosion or both. Truncation of the Lower Devonian Hunton Group in the Anadarko Basin, where the middle Paleozoic section is more completely preserved, however, indicates that the region experienced a major period of tectonism in the Middle Devonian (Maxwell, 1959; Ham and Wilson, 1967). Uplift and erosion of Ellenburger and younger Ordovician, Silurian, and Devonian rocks along the Texas Arch probably took place at this time.

Following middle Devonian tectonism the Texas Panhandle began to once again subside and receive sediment. Shales of the Late Devonian Woodford Formation were deposited in low energy, anoxic basin in the northeastern Panhandle and in the Midland Basin. The Texas Arch remained emergent throughout the remainder of the Devonian Period (fig. 5).

Early Carboniferous Deposition

Inundation of the Texas Panhandle continued in the Early Mississippian. Although most of the Palo Duro Basin area remained emergent, coarse clastics derived from granitic terranes along the Texas Arch accumulated during the Kinderhookian as proximal, high-energy deposits that were equivalent to Woodford muds accumulating in the northeastern Panhandle (fig. 26a).

The Precambrian core of the Texas Arch may have remained emergent into the Osagean Epoch (fig. 26b). There is no solid evidence of Osagean deposition in the

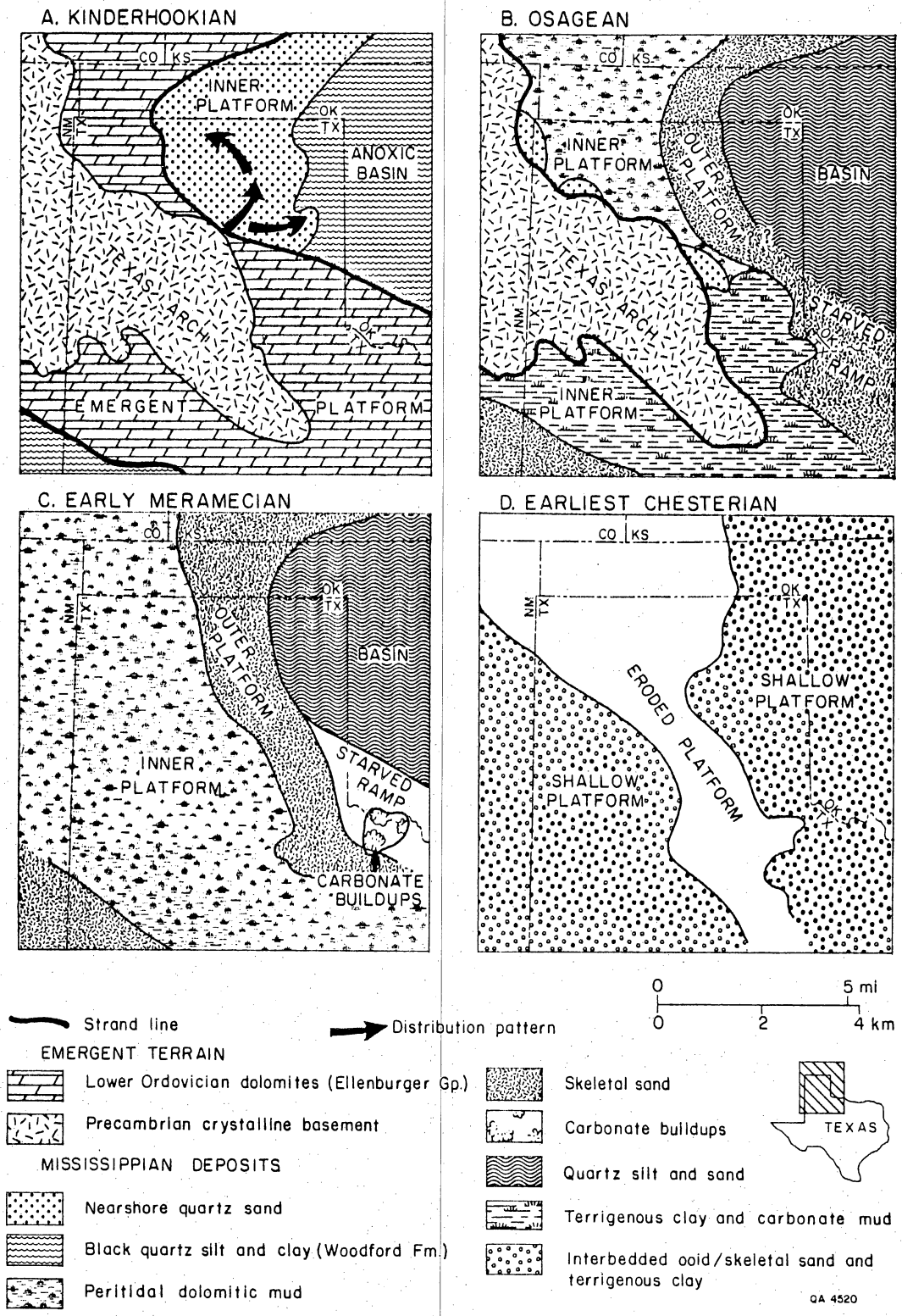


Figure 26. Mississippian paleogeography. A - Kinderhookian. B - Osagean. C - Early Meramecian. D - Earliest Chesterian. After Ruppel (1988).

southern Texas Panhandle except on the northeastern margin of the Palo Duro Basin area (Ruppel and Lemmer, 1986). Late Osagean rocks recognized there are shallow water carbonates and fine grained sandstones that were probably deposited along the transgressive margin of the proto-Anadarko Basin which had begun to form in western Oklahoma. By the early Meramecian, the entire region was submergent and accumulating sediment. The southern Texas Panhandle was characterized by the development of a shallow-water platform that extended across most of the area (fig. 26c). Deposition in the Palo Duro Basin area was characterized by carbonate muds. Isolated carbonate buildups developed coevally along the platform margin in the eastern Palo Duro Basin/Hardeman Basin area (fig. 26c). These buildups were surrounded by aprons of skeletal debris derived from the buildups and spiculitic, carbonate muds deposited in quiet water between buildups.

During the late Meramecian carbonate sand shoals developed across most of the Palo Duro Basin area. By the end of the Meramecian, these sand shoal deposits of ooids and skeletal debris had prograded eastward beyond the platform margin burying the buildups and establishing a shallow-water, high-energy platform across much of the region. At the end of the Meramecian, relative sea level fall resulted in erosion in the central Palo Duro Basin area along the axis of the old Texas Arch (fig. 26d); shallow water carbonate deposition was continuous off the flanks of the Texas Arch (for example in the Hardeman Basin area). This regression was primarily the result of glacio-eustatic sea level fall (Veevers and Powell, 1987).

Following a relative rise in sea level early in the early Chesterian, shallow-water platform carbonates, consisting of ooid and skeletal sands and muds, were once again deposited across the entire Texas Panhandle. Accumulation of these carbonates was periodically interrupted by the influx of terrigenous clastics derived from the northwest (Ruppel, 1985). The influx of these siliciclastics is the first evidence of the beginning of tectonic activity in the region that ultimately produced the structural and depositional patterns characteristic of the Palo Duro Basin.

Late Carboniferous/Early Permian Tectonics and Deposition

Three major episodes of deformation have been recognized in the southern Texas Panhandle (Ham and Wilson, 1967; Budnik, 1988). During the earliest Pennsylvanian (Morrowan), the northern margin of the present Palo Duro Basin began to be uplifted. Mississippian and older strata were eroded from this area while lower Pennsylvanian sediments accumulated on the relatively undisturbed shallow water platform to the south (Budnik and Smith, 1982). Earliest Pennsylvanian deposition in the Palo Duro Basin was characterized by the episodic influx of quartzose siliciclastics and carbonate conglomerates derived from the newly forming highlands on the northern margin of the Palo Duro Basin. Shallow water carbonates (ooid and skeletal sands and muds) accumulated in areas not flooded by these siliciclastics.

Major deformation of the Palo Duro Basin area occurred during the middle Pennsylvanian (Desmoinesian) Ancestral Rocky Mountain Orogeny (Kluth and Coney, 1981; Goldstein, 1982; Budnik, 1988). At this time, the Amarillo Uplift and related structures along the northern margin of the Palo Duro Basin attained significant relief and began to shed large volumes of arkosic siliciclastics, forming fan deltas in the low-lying areas that rimmed the highlands (Dutton, 1980a). Most of the sediment derived from the Amarillo Uplift was trapped in a series of rapidly subsiding troughs that fronted the Amarillo Uplift on the south (Budnik, 1988). Throughout the remainder of the Palo Duro Basin area, shallow-water carbonate deposition was periodically interrupted by the influx of siliciclastics supplied by migrating delta channels (Dutton, 1980a). Fine-grained siliciclastic muds accumulated in the extreme southern part of the Texas Panhandle in a deeper water basin (fig. 27a). According to Budnik (1988), formation of the Amarillo Uplift and related structures was the result of major left-lateral strike-slip movement associated with the approach of the North America and South America/Africa supercontinent.

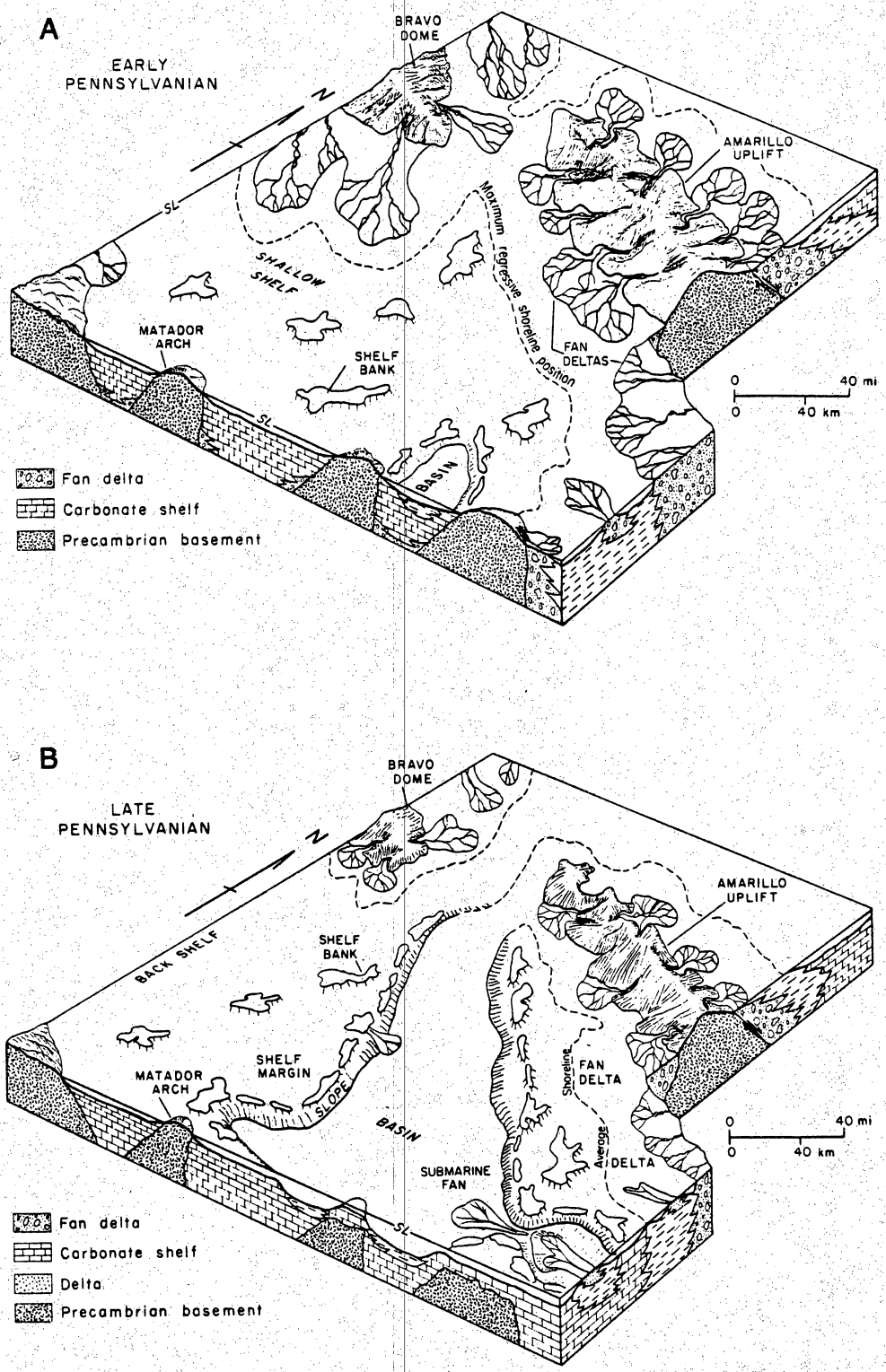


Figure 27. Pennsylvanian paleogeography. A - Early Pennsylvanian. B - Late Pennsylvanian. After Dutton (1980a).

During Late Pennsylvanian deformation, the Palo Duro Basin area was differentiated into well defined basin, platform, and highland areas (Dutton, 1980a). Siliciclastic muds accumulated in the deeper water basin which extended north/northwest through the center of the southern Texas Panhandle (fig. 27b). Carbonates, accumulated preferentially along the margins of the subsiding basin away from the negative influence of siliciclastic sediments, which continued to be carried into the area by fan deltas and delta channel systems along the marginal uplifts (Dutton, 1980a, 1982). Carbonate buildups also developed on local structural highs, principally along the margins of the basin (Budnik, 1988).

Filling of the deep-water basin with silicic muds probably began in the late Pennsylvanian (Virgilian). Depositional conditions during early Wolfcampian, however, continued much the same as they were during the late Pennsylvanian (Dutton, 1980a; Handford, 1980a). The geographic extent of the basin actually reached its greatest limits during the early Wolfcampian (fig. 28a). Continued basin infilling was associated with the lowering of the Amarillo Uplift and a decreasing supply of siliciclastic sediment (fig. 28b). By the end of the Wolfcampian Epoch the Amarillo Uplift had been nearly completely eroded and was submerged. Most of the Texas Panhandle was characterized by deposition of normal marine carbonates, including ooid and skeletal sands, on a relatively flat, shallow-water platform (fig. 28c).

Middle to Late Permian Cyclic Sedimentation

From the late Wolfcampian to the end of the Permian the southern Texas Panhandle area was characterized by shallow-water deposition on a stable platform. This contrasts with the Pennsylvanian and early Permian when the area exhibited wide variations in topography and water depth caused by differential subsidence. During the Middle and Late Permian, structural controls on deposition were limited to small areas

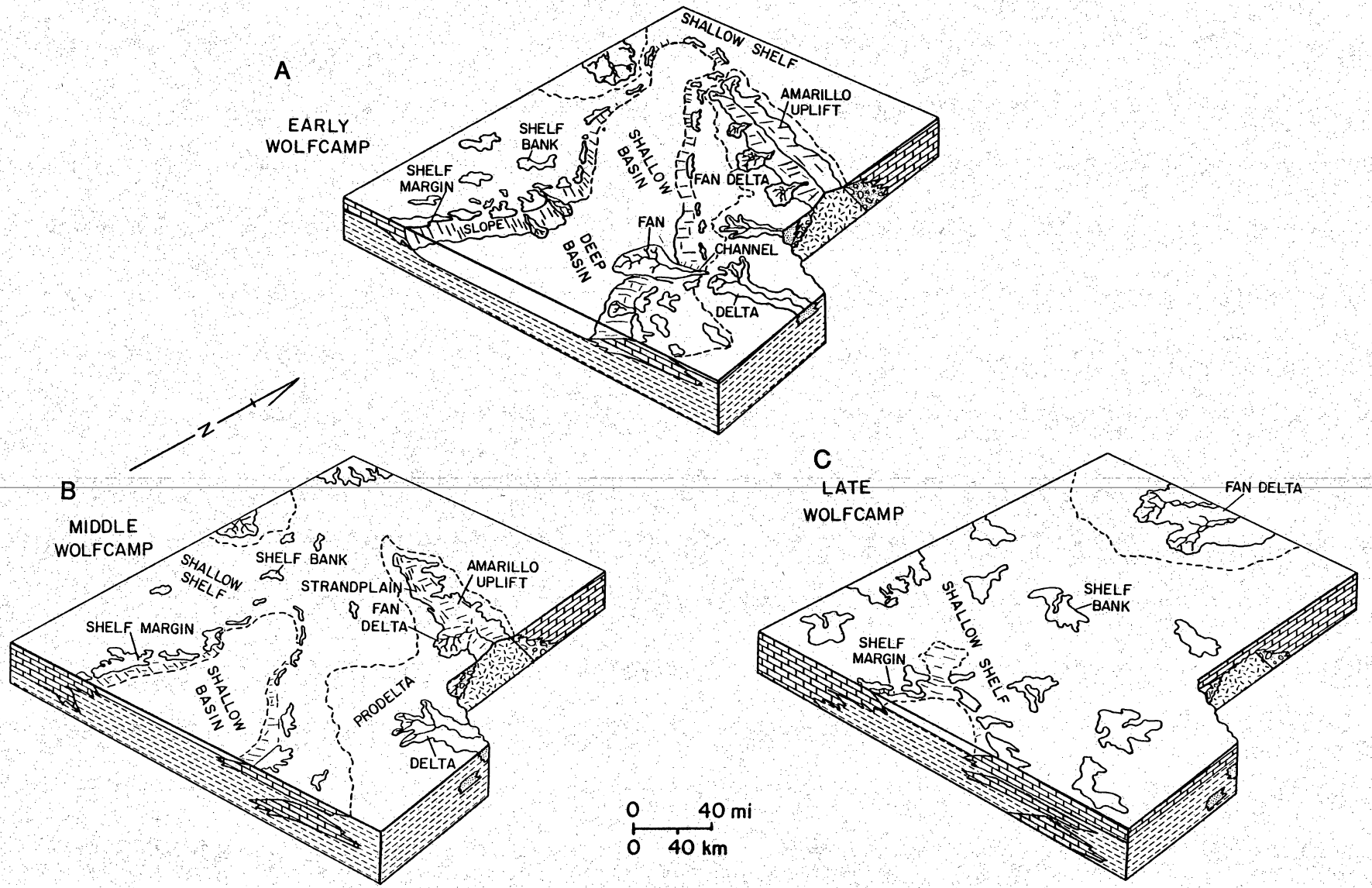


Figure 28. Wolfcampian paleogeography. A - Early Wolfcamp. B - Middle Wolfcamp. C - Late Wolfcamp. After Handford (1980a).

on the southern and northern margins of the Palo Duro Basin (Budnik, 1988). For most of the area the most important control of depositional style was the rise and fall of relative sea level.

Deposition during the Middle to Late Permian took place in three environmental settings: (1) shallow-water, normal marine, (2) hypersaline brine pools, and (3) coastal terrestrial (fluvial and eolian). At the beginning of the Leonardian Epoch, shallow-water carbonates and evaporites (Wichita Group) accumulated throughout much of the Palo Duro Basin area. At this same time, fluvial and eolian terrigenous clastics (Red Cave Formation) were deposited updip in the northern Panhandle (Handford and Fredericks, 1980). Terrestrial environments subsequently migrated southward into the Palo Duro Basin and deposition of carbonates and evaporites was restricted to the northern shelf of the Midland Basin during a fall in relative sea level (regression). A subsequent rise in relative sea level restored carbonate deposition to much of the southern Palo Duro Basin and displaced terrigenous environments northward.

At least seven such cyclic falls and rises in relative sea level occurred during the Middle and Late Permian in the Texas Panhandle. Each fall or regression represents aggradation and southward progradation of paleoenvironments (figs. 19 and 29). Subsequent transgressions document major rises in relative sea level that were in most cases quite rapid. One of these rises, that at the beginning of the Guadalupian, may have been especially especially pronounced.

Leonardian regressive-transgressive episodes indicate a progressive shallowing or offlap presumably caused by progradation of the platform margin southward into the Midland Basin. During this time, for example, carbonate environments moved progressively southward (fig. 23).

Major flooding of the southern Texas Panhandle platform occurred at the beginning of the Guadalupian as evidenced by the return to normal carbonate deposition at the beginning of San Andres time (fig. 23). This event may have been caused by renewed

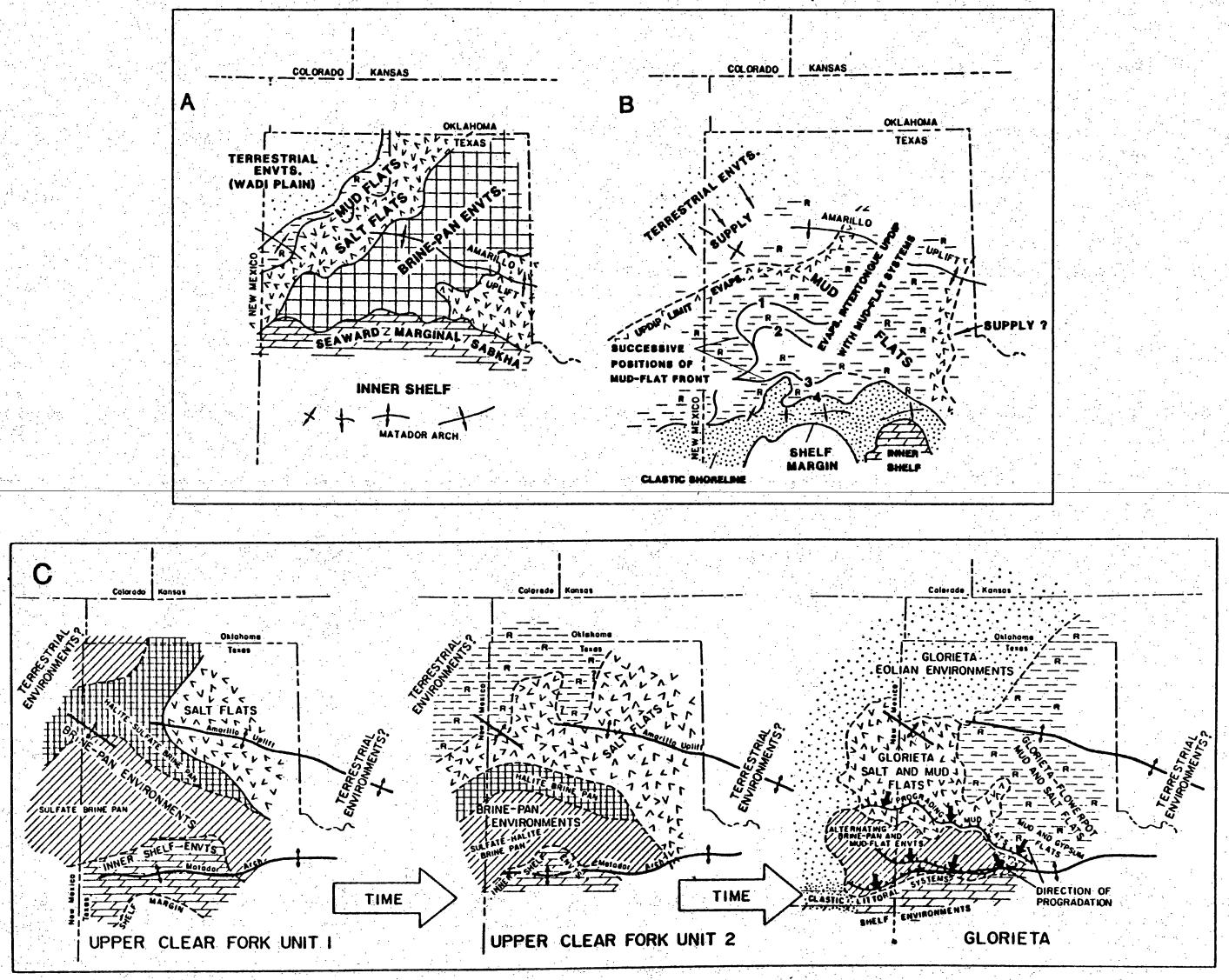


Figure 29. Leonardian paleogeography. (A) Lower Clear Fork, (B) Tubb, (C) Upper Clear Fork through Glorieta. After Presley (1987).

subsidence along the platform margin producing a drowning or backstepping of the margin to the north. The great extent and continuity of San Andres deposits suggests in addition, however, that the Palo Duro Basin platform had become even more flat and broad. Subsequent depositional events appear to have been relatively synchronous across wide areas of the Palo Duro Basin. Late Guadalupian and Ochoan events record progressive restriction and southward migration of environments similar to that seen during the Leonardian (fig. 23).

By the end of the Permian the area was characterized primarily by terrestrial and nearshore terrigenous clastics deposition (Dewey Lake deposits), similar in many respects to the fluvial and lacustrine setting that typified the Triassic (see Dubar, this volume).

The causes for the observed cyclicity in Leonardian to Ochoan rocks in the Palo Duro Basin area are arguable. Pre-Guadalupian cyclicity is well documented throughout the world (Ross and Ross, 1985) and has been ascribed to the combined effects of fluctuations in Permian glacial ice cover and orbital eccentricities (Veevers and Powell, 1987). Guadalupian and Ochoan cyclicity, on the other hand, postdates Late Paleozoic glaciation (Crowell, 1978; Veevers and Powell, 1987). Anderson (1984) has related cyclic deposition in the early Ochoan Castile Group to orbital precession and eccentricity cycles. The observed cyclicity in the Guadalupian and Ochoan of the Palo Duro Basin may also be a response to these Milankovitch cycles. Unfortunately, the lack of chronostratigraphic control in the Palo Duro Basin makes it impossible to assess this possibility.

Further Areas of Study

Despite 10 years of intensive study, several problems regarding the stratigraphy of the Palo Duro Basin remain. Major remaining issues fall into three broad and

overlapping categories: (1) depositional processes, (2) tectonic processes, and (3) diagenetic processes.

General depositional models have been worked out for most of the major lithologic units. Exceptions include the Wichita Group and the Seven Rivers Formation. Knowledge of several other units could also be substantially increased by further examination of these deposits in light of the abundant cores and other data that became available late in the project which, in most cases, were not integrated into previous work. This is true for most of the Permian section, with the exception of the San Andres, which has been studied in great detail because of its potential as a waste repository, and the Queen/Grayburg. Although the San Andres probably serves as good model for other Leonardian to Ochoan deposits, careful examination of these other units would help improve understanding of the processes and controls of cyclic evaporite sedimentation on stable platforms. The Wichita Group is particularly poorly known. Considering its importance as a potential aquitard between the overlying evaporites, which might host a waste repository, and the underlying permeable carbonates (Wolfcamp Series), which could act as a conduit for ground-water flow to the biosphere (see Dutton, this volume), a detailed study of this sequence should be especially important.

Knowledge of the Wolfcamp Series carbonates, the major deep basin aquifer, is still rather rudimentary. Basic models for the depositional setting of these complex and diverse deposits have been presented but no detailed studies of paleoenvironments, lithologic and mineralogic variability, and their effect on the distribution of porosity and permeability have been completed.

Studies of the effects of diagenesis on the Paleozoic sequence have similarly been neglected. Evaporites in the San Andres have been studied in some detail. These studies, which have documented a very complex depositional and diagenetic history for these deposits, are the exception, however. Analysis of the diagenetic alteration of Permian carbonate (for example, the Wolfcamp and Wichita) is especially needed to

help develop predictive models of ground-water flow based on relationship between porosity and permeability.

There is also a great need for studies directed toward establishing a chronostratigraphic framework for the Paleozoic of the Palo Duro Basin. Many unresolved issues, such as (1) timing of tectonic, depositional, and diagenetic events and (2) periodicity and controls of cyclicity, could be approached more rigorously with chronostratigraphic control. Preliminary work with Rb/Sr dating of diagenetic clays, Sr^{87}/Sr^{86} isotopes for dating carbonates and sulfates, and conodont faunas have shown promise in this area.

POST-PERMIAN STRATA AND SURFACE PROCESSES

Jules R. DuBar

General Stratigraphy

Triassic Dockum Group

Introduction

Sediment character, lithofacies architecture, and stratigraphy of Dockum strata within the Palo Duro Basin (table 1) are of special interest because these deposits are superjacent to Permian units that include salt layers designated as a potential host to high-level nuclear waste. Objectives of stratigraphic and sedimentologic studies of the Dockum Group were to: (1) identify the principal depositional episodes and events and characterize corresponding depositional complexes and facies sequences, and (2) detect changes in sandstone trends and thicknesses that might indicate changes in source areas, sediment transport directions, and syntectonic activity in the basin.

Cummins (1890a) named the "Dockum beds" from Triassic deposits exposed near Dockum, in western Dickens County, Texas (see fig. 30 for location). Gould (1906, 1907) named lower Dockum beds the Tecovas Formation and the upper beds the Trujillo Formation. He described the Tecovas as largely shale, the lower part more or less sandy. He stated the thickness of the formation ranges from 90 to 225 ft (27 to 69 m) in the type area, and unconformably overlies the Permian Quartermaster Formation. Cummins (1980a) described the Trujillo Formation as a massive, more or less cross-bedded sandstone and conglomerate with interbedded red and gray shale. He gave the thickness of the type deposits as 25 to 250 ft (8 to 76 m) and stated that the unit was unconformably overlain by Neogene deposits. The Dockum Group of West Texas was classified as Middle Late Triassic (Late Carnian to Early Norian) by Reeside (1957).

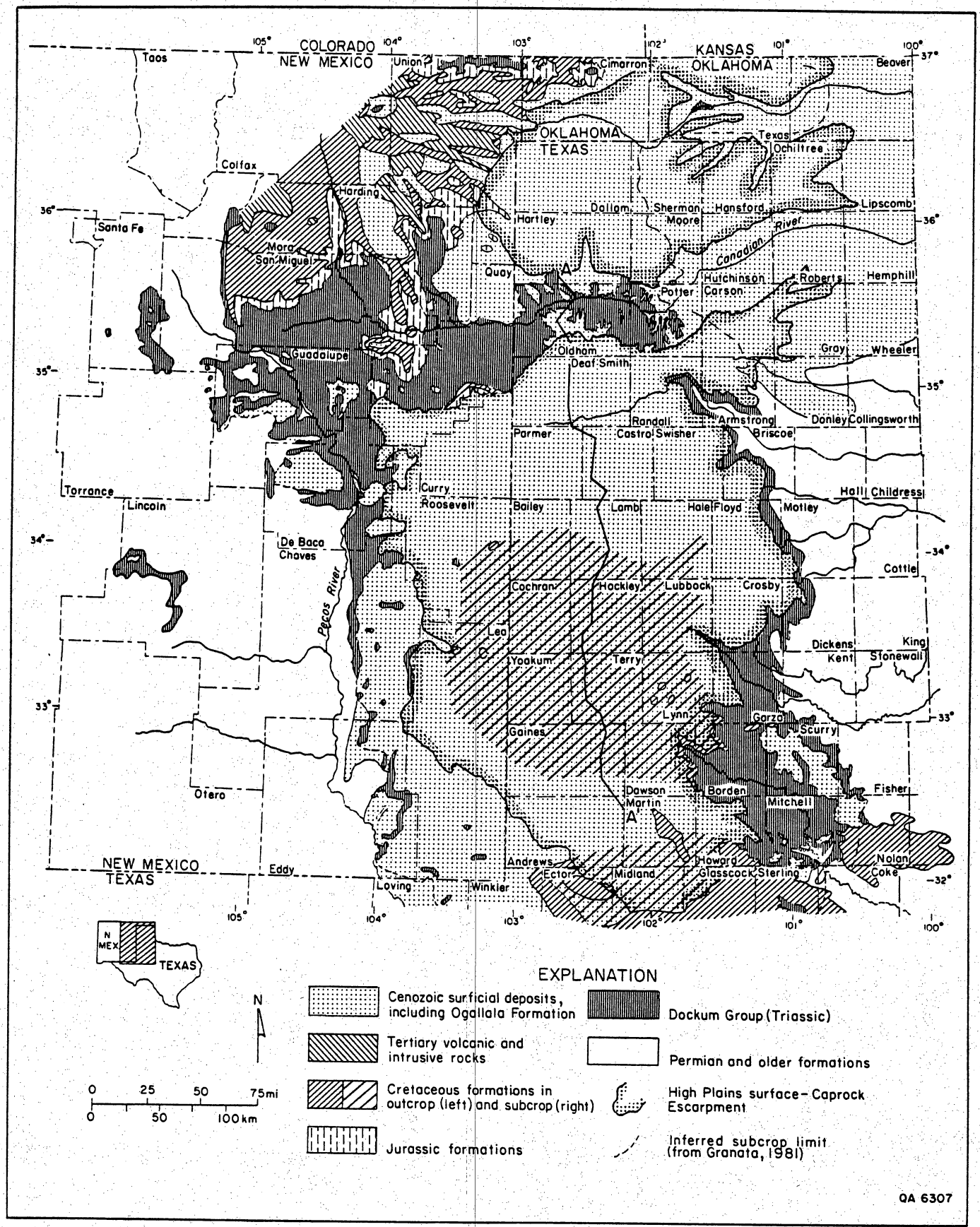


Figure 30. Geologic map of the study area, including surficial geology and subcrops of the Edwards/Trinity Groups. From Dutton and Simpkins (1986).

Since the work of Cummins and of Gould, numerous papers have been published on the Triassic of the Texas Panhandle (for example, Adams, 1929; Green, 1954; Kiatta, 1960; Cramer, 1973; Asquith and Cramer, 1975; Finch, 1975; Dutton and Simpkins, 1986) and several unpublished theses and dissertations have been completed (Cazeau, 1962; Seni, 1978; Boone, 1979; Granata, 1981). Recent reports by McGowen and others (1977, 1979) and by Johns (1985) together provide detailed documentation of the depositional framework of the Dockum Group in the Texas Panhandle and Eastern New Mexico. The work of McGowen and others (1977, 1979) focuses on outcrop studies and on the subsurface in the Midland Basin and that of Johns (1985) covers the subsurface of the Palo Duro Basin (fig. 1). In these papers Triassic sedimentation and stratigraphy are analyzed in terms of the genetic facies that compose the depositional systems.

Most of the following discussion of the Dockum Group is derived from the papers by McGowen and others (1979) and by Johns (1985).

Geologic Setting

The Dockum Group is an intricate complex of lacustrine, fluvial and deltaic clastic facies deposited in a continental basin underlying western and northwestern Texas, eastern New Mexico, and parts of Oklahoma, Kansas and Colorado (fig. 30). Sediment thickness is more than 2,000 ft (610 m) in the Midland Basin (Granata, 1981; McKee and others, 1959) (figs. 31 and 32). In the Palo Duro Basin, an area of about 12,000 mi^2 (19,300 km^2), Dockum deposits range from less than 100 ft (30 m) in thickness in the north to 1,800 ft (549 m) in the southwest (Johns, 1985) (fig. 33).

The Dockum Basin was peripherally filled from the east, south, west and north (fig. 34). Highest gradient streams were located to the north and to the south of the basin. On the east and on the west, the basin was bordered by lowlands over which meandering streams flowed (McGowen and others, 1979).

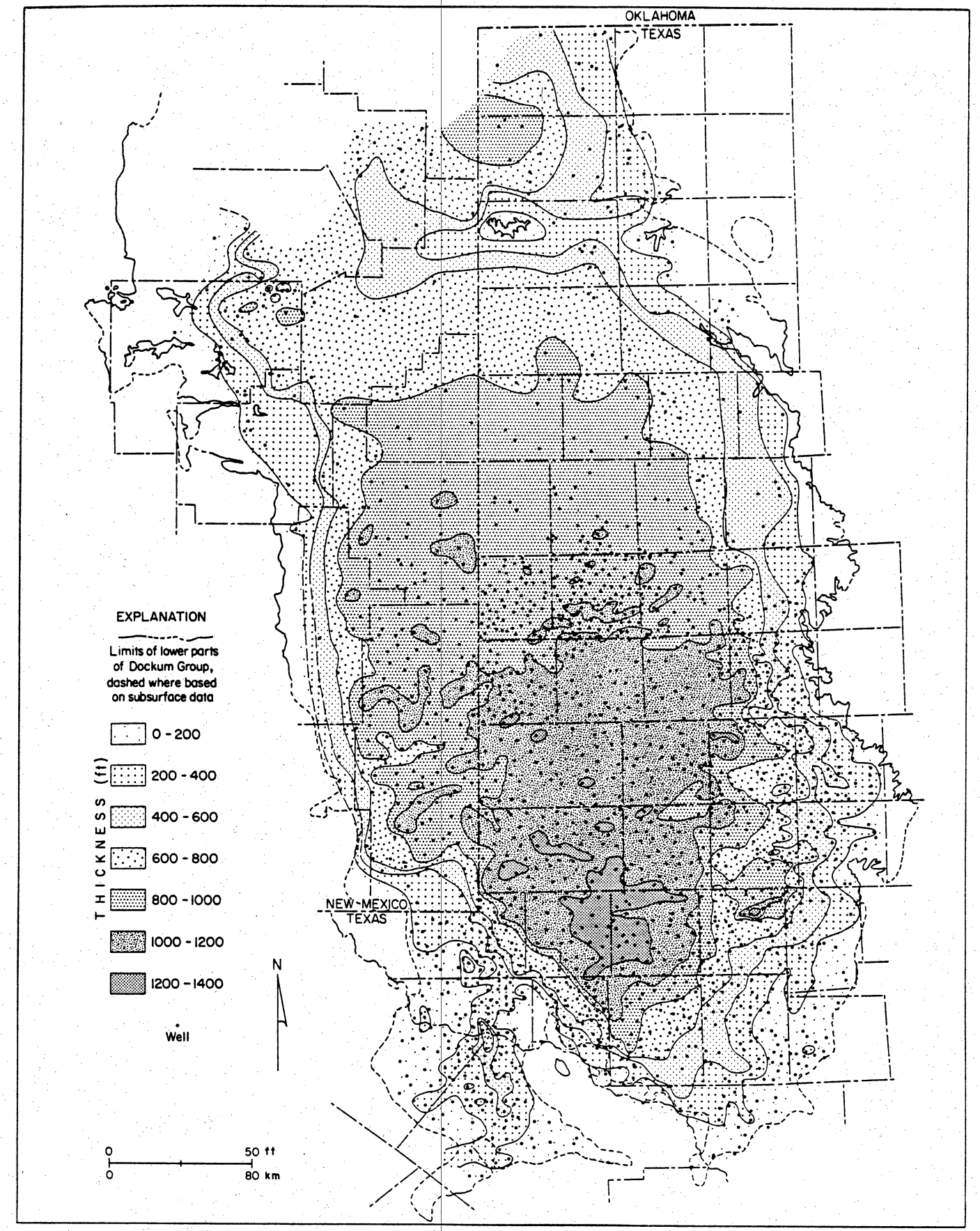


Figure 31. Isopach map of the Lower Dockum Group. From Dutton and Simpkins (1986).

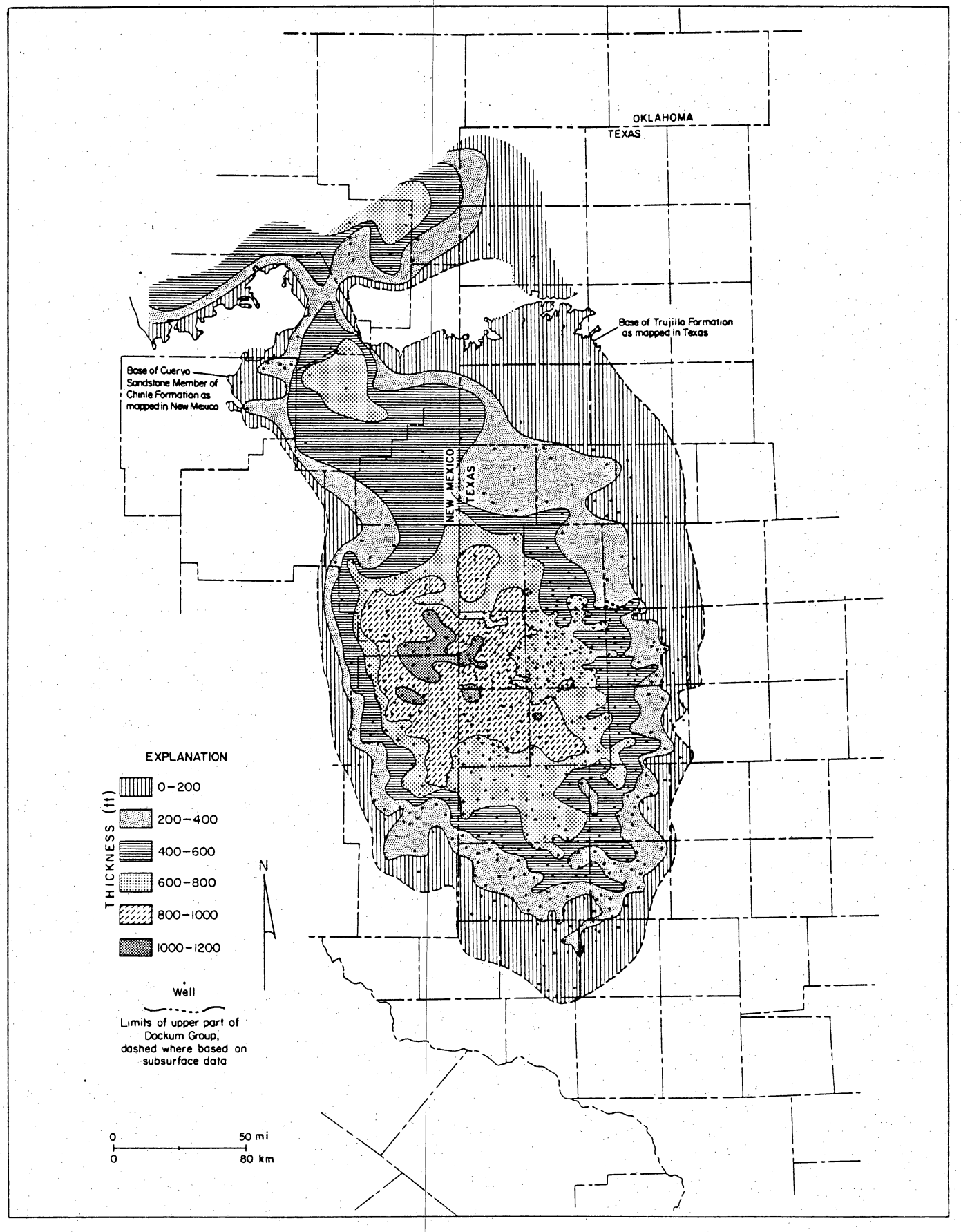


Figure 32. Isopach map of the Upper Dockum Group. From Dutton and Simpkins (1986).

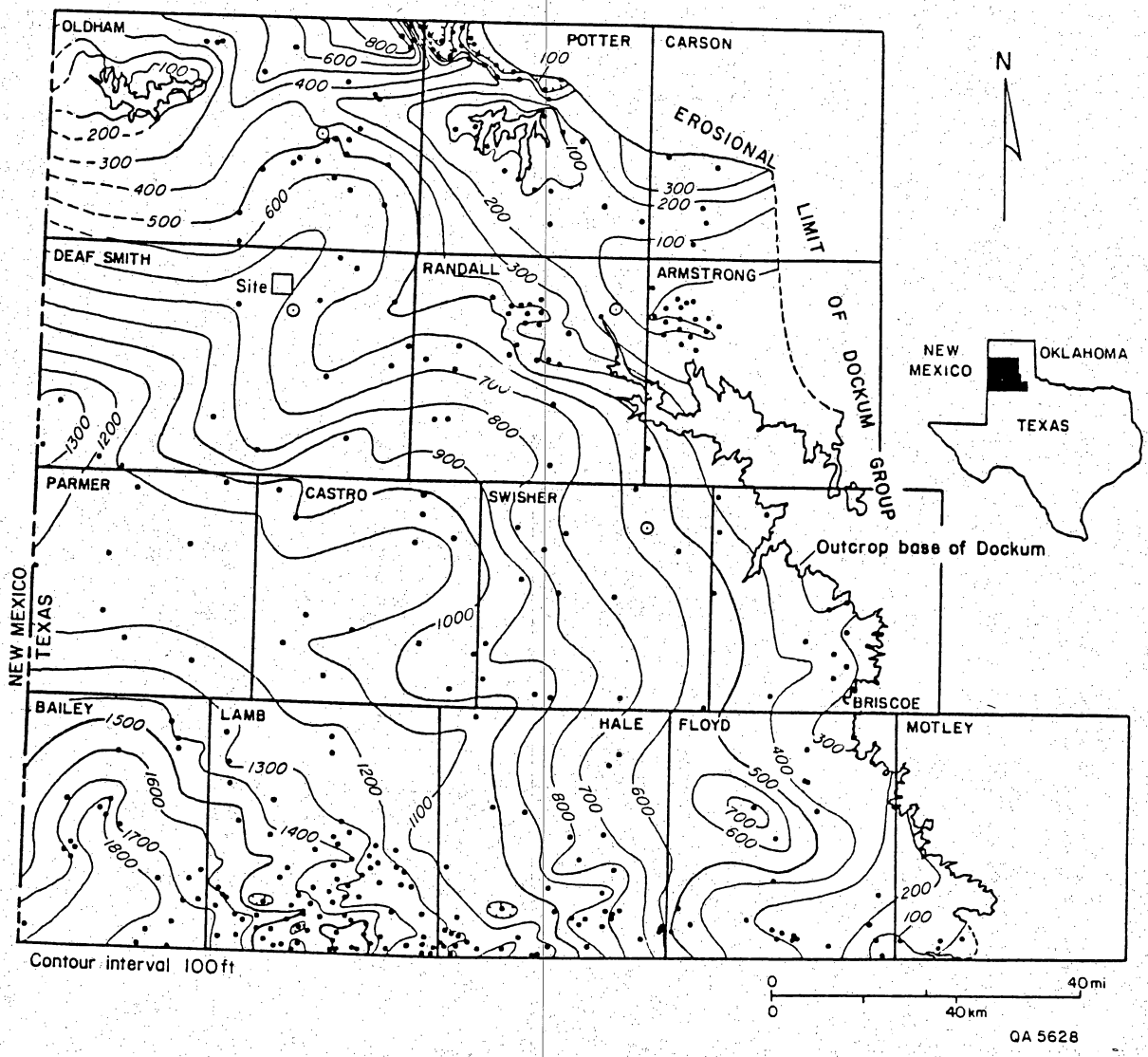


Figure 33. Isopach of the Dockum Group. The unit thickens to the southwest. Isolated thick areas in Carson, Potter, and Oldham Counties are in fault bounded structural lows. Faults are not shown. After Johns (1985).

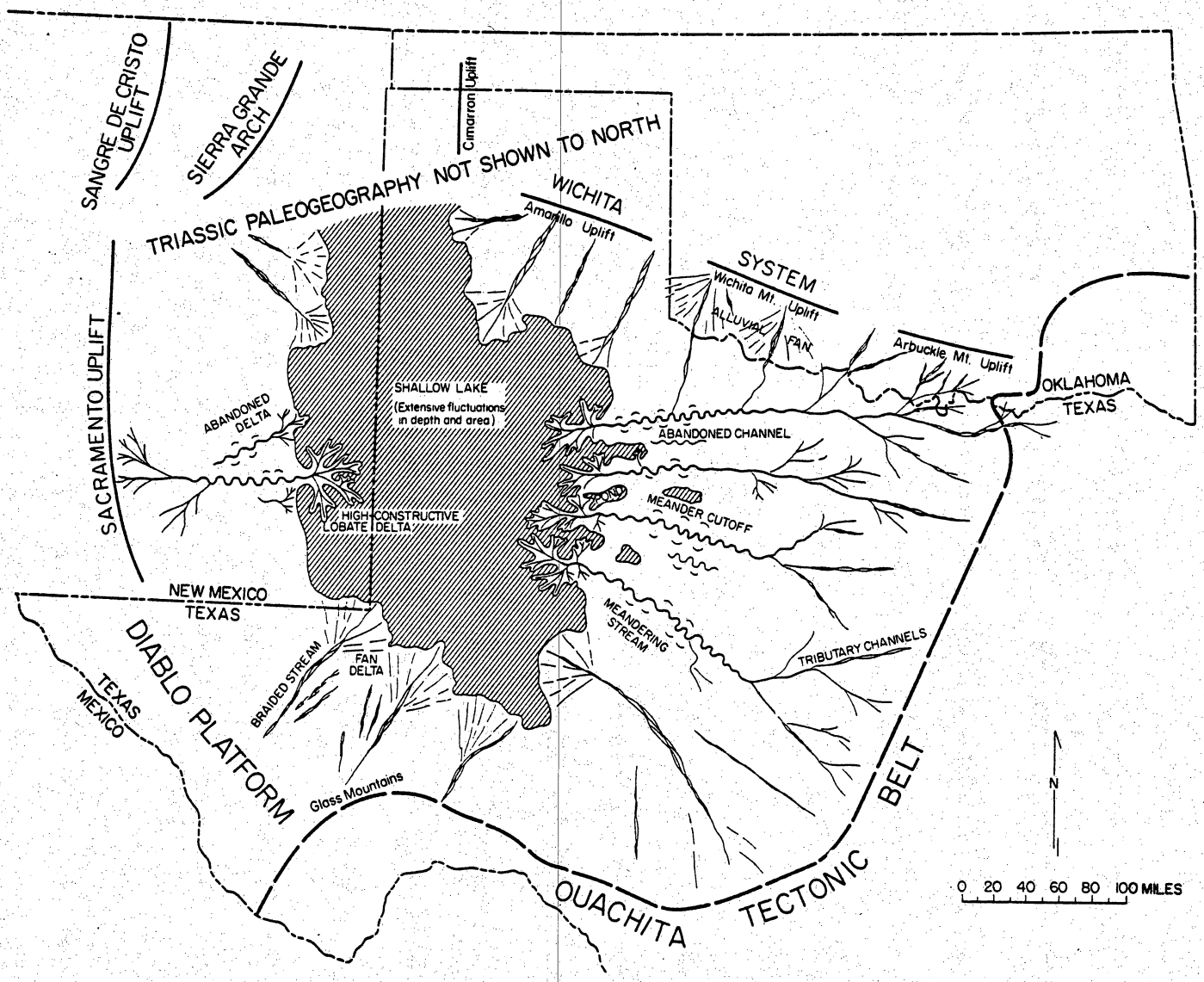


Figure 34. Inferred paleogeography during the initial stage of Dockum sedimentation in the area south of Amarillo Uplift - Bravo Dome. Depositional elements are braided streams, alluvial fans, fan deltas, meandering streams, distributary deltas, and shallow lakes. From McGowen and others (1979).

McGowen and others (1979) observed that an unconformity between the Permian and Triassic is obvious in the northern part of the basin, but physical evidence of an unconformity in the central basin area is not evident. Dockum outcrops in the Texas Panhandle are generally considered to be of Late Triassic age and separated from the Permian by a distinct unconformity. Adkins (1932) cited the evidence for an unconformity between the Dockum and underlying Permian: (1) The Dockum is Late Triassic, and Early and Middle Triassic sediments apparently were not deposited in the basin; (2) there is a dip discordance between Permian and Triassic strata; and (3) unconformities are visible in many places.

Adkins (1932) further noted that there are no fossils near the contact and that the beds on both sides are generally similar in appearance. Some writers have stressed the gradational appearance of the contact (Case, 1922). Reeside (1957) postulated a basin-wide unconformity because of the absence of Early and Middle Triassic fossils in the Dockum. Granata (1981) recognized both gradational and unconformable subsurface contacts between the Permian and Triassic with the former most common in the basin center.

Page and Adams (1940) reported that the first Triassic sediments (Tecovas Formation) deposited in the Midland Basin of Texas are composed primarily of reworked Upper Permian sands and, therefore, closely resemble the underlying, undisturbed Permian beds. Page and Adams note that the Dockum deposits are distinguished from the Permian by a new and rich heavy-mineral suite in the Triassic sediments and because they are not as indurated as the Permian beds. They conclude that because no fossils have been found in these lower Triassic strata they should be considered of Late Triassic age until evidence to the contrary is found.

McGowen and others (1979) suggested that Dockum sedimentation was initiated by opening of the Gulf of Mexico at the beginning of the Triassic with associated uplift in part of the Ouachita Tectonic Belt and subsidence of the land surface across

the ancient Midland and Palo Duro Basins. The uplifting closed the Permian seaway and led to creation of a closed, fresh-water-dominated continental basin. Structures along the Matador Arch apparently had little effect on sedimentation, but local structures influenced lithology distribution and thickness of sediment (Johns, 1985). Basement structural elements affected orientation of depositional axes, sandstone distribution and thickness, and total volume of sediment accumulation.

Midland Basin

McGowen and others (1979) recognized several depositional cycles, each comprising facies deposited during one high stand - low stand couplet of lake levels. During a high stand, thin progradational (delta) and meanderbelt systems were deposited. Typical delta sequences include lacustrine and prodelta mudstones and siltstones, delta-front siltstones and sandstones, and meanderbelt sandstones and conglomerates. Splay deposits are poorly sorted, intrabasinal sandstone and conglomerate incorporated with interdistributary and floodplain deposits. Delta sequences generally were partly reworked by superimposed meandering streams that crossed the area.

Lowering of base level, as the climate became progressively more arid, was accompanied by erosion of subjacent Dockum deposits. Sediments of low-stand facies range from reddish brown mudstone to conglomerate, and are characterized by sharp vertical and lateral changes. McGowen and others (1979) recognized five types of lower Dockum deposits: (1) lacustrine mudstone (2) prodelta mudstone and siltstone, (3) delta front (delta forset) siltstone to conglomerate, (4) delta platform sandstone and conglomerate, and (5) interdeltaic mudstone exhibiting desiccation features, rare gypsum, salt hoppers, and chert.

McGowen and others (1979) thought that the Permian arid conditions were slowly replaced by a more humid climate in the Triassic. Early Dockum sediments interpreted as recording more humid conditions were deposited in: (1) braided and meandering streams, (2) alluvial fans and fan deltas, (3) high constructive lobate deltas, and (4) lakes. Later change from humid to arid conditions produced: (1) lowered base level, (2) erosion of older Dockum sediments, (3) replacement of meandering fluvial systems by lakeward eroding valleys and braided streams, and (4) development of small fan deltas.

Palo Duro Basin

In the Palo Duro Basin structural elevation on the base of the Dockum shows a south to southwest dip of less than one degree. Dips are reversed to the north in Oldham, Potter, and Carson Counties due to structural modification along the trend of the Amarillo Uplift, possibly accentuated by salt dissolution.

Johns (1985) informally divided the Dockum into upper and lower units. The boundary between these units was placed near the middle of the section at the base of a regional progradation event marked by increased sandstone content. The upper Dockum is present only in the central and western part of the Palo Duro Basin (fig. 34). Erosion has removed the unit in the east and there are no outcrops of the unit along the Eastern Escarpment, although its apparent equivalent, the Chinle Formation, is exposed in eastern New Mexico. Formation thickness (fig. 33) reflects pre-Ogallala erosion and thickens westward from the eastern erosional limit.

The Lower Dockum constitutes most of the Triassic section in the Palo Duro Basin. Its greatest known thickness is 1,200 ft (366 m) in Bailey County, and it generally thins to the north and east.

The Dockum Group in the Palo Duro Basin is characterized by repetitive transgressive and progradational sequences of interbedded sandstone, siltstone and mudstone deposited by fluvial, deltaic and lacustrine systems. Changes in precipitation amounts caused significant fluctuations in lake levels resulting in rapid and abrupt vertical and lateral facies and altering of the types of stream systems discharging into the basin.

Johns (1985) identified several depositional cycles in Dockum sediments of Palo Duro Basin. Cycles consist of fine- to coarse-grained progradational fluvial and deltaic clastics overlain by transgressive lacustrine clays and silts. More-or-less laterally continuous sequences of sandstone beds characteristic of progradational hemicycles were referred to by Johns as "sandstone packages." In areas where bounding lutites are thin or absent, boundaries of sandstone sequences were correlated using sandstone character as recorded on gamma-ray logs. Where sandstones are absent, the interval was considered to be absent. Five sandstone progradational hemicycles (sandstone packages) were identified in the lower Dockum by Johns, a study that represents a more detailed subdivision of the Dockum than that done by McGowen and others (1977).

Johns (1985) recognized five progradational events recorded by deposition of coarse-grained fluvial and deltaic clastics. The lower two sequences (hemicycles) represent initial Triassic sedimentation in the Palo Duro Basin. Net sandstone maps indicate that sediment sources were from the west, north and east. Core and outcrop equivalents of the sequence indicate that deposition was by alluvial-fan and fan-delta systems on alluvial plains, and in shallow (<60 ft [<18 m]) lakes. Widespread siltstones and mudstones (transgressive hemicycle or "non-sandstones" of Johns, 1985) deposited between the first and second progradational sequences apparently represent significant deepening and westward expansion of the lake system (Johns, 1985). During deposition of the second progradational sequence the lake level fell and feeder

streams cut deep valleys in updip areas. Reworked sediment was deposited downdip and large areas of lacustrine mudstones were eroded and subaerially weathered. During subsequent rise of lake levels, streams filled the valleys with aggradational bedload fluvial and transgressive deltaic sands.

The upper three cycles record progradational deltaic sedimentation. Progradational sequence 3, a basin-wide deposit, was fed from both western and eastern sources. Core and outcrop equivalents are delta-front, delta and fluvial progradational sequences. Johns (1985) postulated that lake water was deeper during this cycle than during previous cycles. Progradational sequence 3 is characterized by numerous, relatively thin sandstones and interbedded lutites. Sediment source appears to have been primarily from the east, supplemented by a significant western source. Johns (1985) interpreted these thin sandstones to be deltaic sediments deposited over fluvial sandstone beds as the delta environment retreated from the basin center in response to expansion of the lacustrine environment.

Sandstones of cycle 4 differ markedly from those of lower cycles. The volume of sandstone is reduced and localized in the deeper parts of the basin. Net sandstone contour patterns reflect a dominant northwest source with lesser contributions from the east (Johns, 1985). Sandstones are typically thin and separated by relatively thick lutite deposits. Johns (1985) speculated that lake levels were high during this time and that thin sandstone units are proximal delta and delta-front facies, whereas rarer, thick sandstones are channel facies.

Cycle 5 includes all strata between cycle 4 and the base of the upper Dockum. Thickest sandstones are in the southwest part of the basin, however, thin sandstones are scattered throughout the remainder of the basin. Study of core from the cycle 5 section led Johns (1985) to conclude that coarser clastics represent delta-front and distributary facies. Johns also noted that cycle 5 deposits are unconformably overlain by distributary channel-fill conglomeratic sandstone. Johns (1985) concluded that lake levels probably were high during this cycle.

Johns (1985) differentiated 10 lithofacies in Dockum core. These include three lacustrine lithofacies: claystone, mudstone, and siltstone. Dominant sedimentary structures in lacustrine facies are burrowing, fine laminations, and ripple cross-stratification. Johns noted that most lacustrine intervals exhibit evidence of subaerial exposure and weathering ranging from mudstone desiccation cracks to pervasive brecciation, calcification, and reworked rip-up clasts.

Four deltaic lithofacies were recognized by Johns. These are: mudflat facies of finely interbedded mudstone, siltstone and sandstone; marsh or swamp facies; distal and proximal delta-front facies represented by ripple and ripple-drift, cross-stratified siltstone and soft-sediment-deformed, interbedded siltstone and fine- to coarse-grained sandstone; and fluvial lithofacies of thick, interbedded, fine- to coarse-grained sandstone and conglomerate with fining upward grain-size trends. A single, localized graded mudclast conglomerate suggested to Johns (1985) deposition by bed-load streams on an exposed mudflat during lowstand of a lake. One occurrence of an eolian facies located at the distal end of an alluvial fan deposit was recorded by Johns.

Triassic deposits exposed along the periphery of the Palo Duro and Midland Basins appear to be entirely of Late Triassic age. Toward the center of the basin, where the Triassic section thickens and where physical evidence for an unconformity is not clear, Early Triassic sediments may be present. Therefore, it is possible, as McGowen and others (1979) hypothesized, that the Permian-Triassic boundary in the basin center is gradational. Further study of the basin center Permian-Triassic boundary is required to establish its character. If Early Triassic deposits are not present within the Dockum Basin, the model proposed by McGowen and others (1979) to explain the depositional history of the Dockum deposits is flawed and requires re-evaluation. Studies are also needed to evaluate permeability and primary and secondary porosity of the Dockum deposits. Such analyses are needed for determining water resource potential of the Dockum Group.

Cretaceous

Introduction

The Lower Cretaceous stratigraphy of the Southern High Plains of Texas has received little attention. The most complete summary to date is that of Brand (1953), who studied all known Cretaceous exposures and a limited number of drillers logs. The subsurface stratigraphy of the Lower Cretaceous rocks of the Southern High Plains of New Mexico was discussed by Fallin (1988). Objectives of these and other studies were (1) to determine the temporal and spatial distribution of Cretaceous deposits in the Southern High Plains, (2) to delineate and characterize important facies and their depositional environments, (3) to identify, describe and determine the significance of diagenetic aspects of each stratal unit, (4) to outline a conceptual model of aquifers in Cretaceous rocks beneath the Southern High Plains and to evaluate their role within the regional hydrologic system.

Stratigraphy

Brand (1953) stated that exposures of the Lower Cretaceous on the Southern High Plains of Texas lie between longitude 101° and 103° W. and latitude 31° 45' and 34° 10' N. (fig. 35). He examined sections in Ector, Andrews, Gaines, Dawson, Borden, Scurry, Terry, Lynn, Garza, Lubbock, Hockley, Bailey, Lamb, and Floyd Counties. These exposures are all along the southern and southeastern margins of the High Plains and along the western margins of the lake basins of the area. Other small outcrops occur in Lea and Curry Counties, New Mexico (Theis, 1932; Nicholson and Clebsch, 1961). According to Nativ and Gutierrez (1987), who studied the hydrology and hydrochemistry of the Cretaceous strata, there are two main areas of Cretaceous subcrop (fig. 35). They report that the Cretaceous strata in these areas

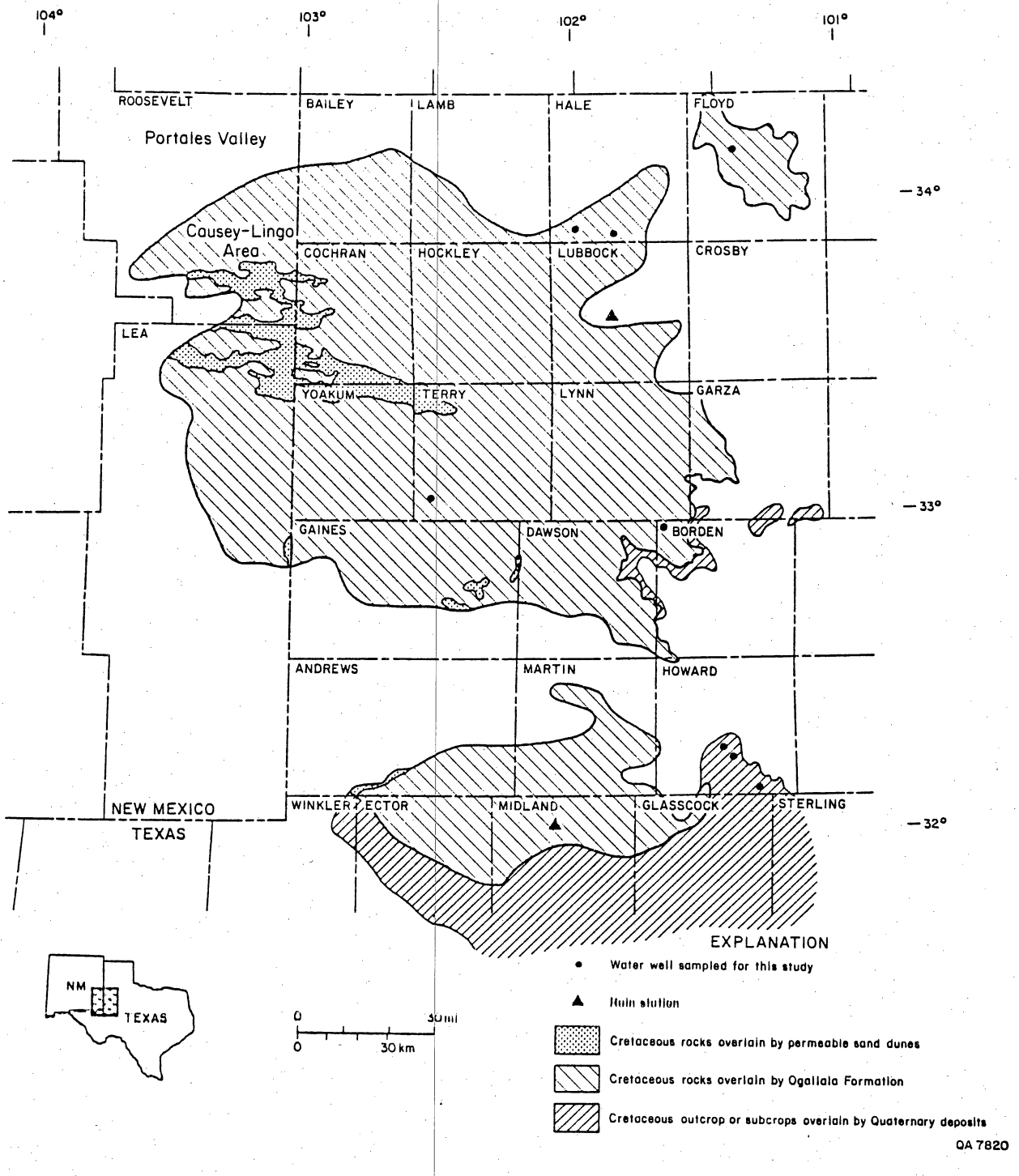


Figure 35. Map of Cretaceous subcrops underlying the Southern High Plains, including location of sampled wells and rain sampling stations. From Nativ and Gutierrez (1987).

range in thickness from a featheredge to at least 250 ft (76 m). Regional dip is southeastward at about 7 or 8 ft/mi (1.3 to 1.5 m/km). Cretaceous sediments were deposited throughout the Texas Panhandle and eastern New Mexico in a shallow sea of unknown extent (Moore, 1958). Subsequently they were partly or in places wholly removed by erosion prior to deposition of the Ogallala Formation. The erosional remnants of the Cretaceous deposits lie unconformably on the Triassic (Dockum Group) and are unconformably overlain by the Ogallala Formation or younger deposits.

Brand (1953) summarized the depositional sequence for the Early Cretaceous of the Texas High Plains as follows: (1) The sites of deposition were within the epineritic and littoral environments of a stable shelf. (2) Sand and gravels of the Paluxy Formation (table 2) are transgressive. The coarse sands probably were deposited at the leading edge of the littoral zone in an advancing sea. Gravel lenses indicate action of inflowing streams, or of rip currents. (3) The Fredricksburg Group records a transgression during Walnut time and further marine expansion during deposition of the Kiamichi (table 1). Limestone precipitation occurred in a warm, shallow sea during Comanche Peak time and rudist biostromes were widespread, both of these facts reflecting a very stable environment over most of Texas during Edwards time. (4) Kiamichi deposition began with an influx of mud. Mud entering the sea at this time probably reflected a more humid climate than during the preceding interval. (5) Shallowing of the Kiamichi sea is indicated by the profusion of massive, thick-shelled pelecypods (*Gryphaea* and *Exogyra*), ripple-marked and cross-bedded sandstone and conglomeratic limestone. (6) Paucity of fossils, thinly-laminated, dark gray and black shales of the upper Kiamichi indicate low sedimentation in a reducing environment. (7) Washita time began with marine transgression and increased water circulation. Mineralogically, Duck Creek shales are similar to those of the upper Kiamichi, with colors differing with state of oxidation.

Nativ and Gutierrez (1987) identified three aquifer elements: the Trinity Sandstone (fig. 36), the Edwards Limestone (fig. 37), and the Kiamichi-Duck Creek sandstones and limestones. Fallin (1988) studied the hydrology of the lower Cretaceous rocks of eastern New Mexico. He noted that the Cretaceous deposits there and in extreme western Texas form buried mesas with more than 200 ft (61 m) of relief. A typical lower Cretaceous section in New Mexico includes relatively thin basal sands and sandstones overlain by marls, clays and limestones. Regional subsurface profiles show that the basal sand correlates with the Antlers Formation (=Paluxy Formation of the Trinity Group in Texas [table 2]). These arenaceous deposits are white to light blue, unconsolidated to moderately well cemented, fine to coarse grained, and quartz-rich; quartz grains bear characteristics of dunal and nearshore marine origin. The lower sands and sandstones form a blanket which pinches and swells as it thins regionally to the northwest. Unit thickness ranges from a featheredge to at least 60 ft (18 m) with maximum thickness over erosional scour channels and other topographic lows on the Dockum surface.

Light blue clay and argillaceous shallow marine limestones overlie the basal sands and sandstones to the south. Limestone distribution is spotty, and rarely exceeds 53 ft (16 m) in thickness. The limestones and underlying clays correlate lithologically with the Walnut and Comanche Peak Formations of the Fredericksburg Group in Texas (table 2).

Dark blue-gray shales capped with yellow-brown clay overlie all older Cretaceous deposits of the area. Thickness of these sediments ranges from a featheredge to more than 160 ft (49 m). The fact that the upper yellow-brown clay covers the entire subcrop area suggested to Fallin (1988) that it may represent an oxidized weathering profile developed following uplift and erosion during Laramide time. Fallin (1988) correlated the middle fine-grained sequence with the Kiamichi Formation. He thought that upper parts of the section could include a part of the Duck Creek Formation.

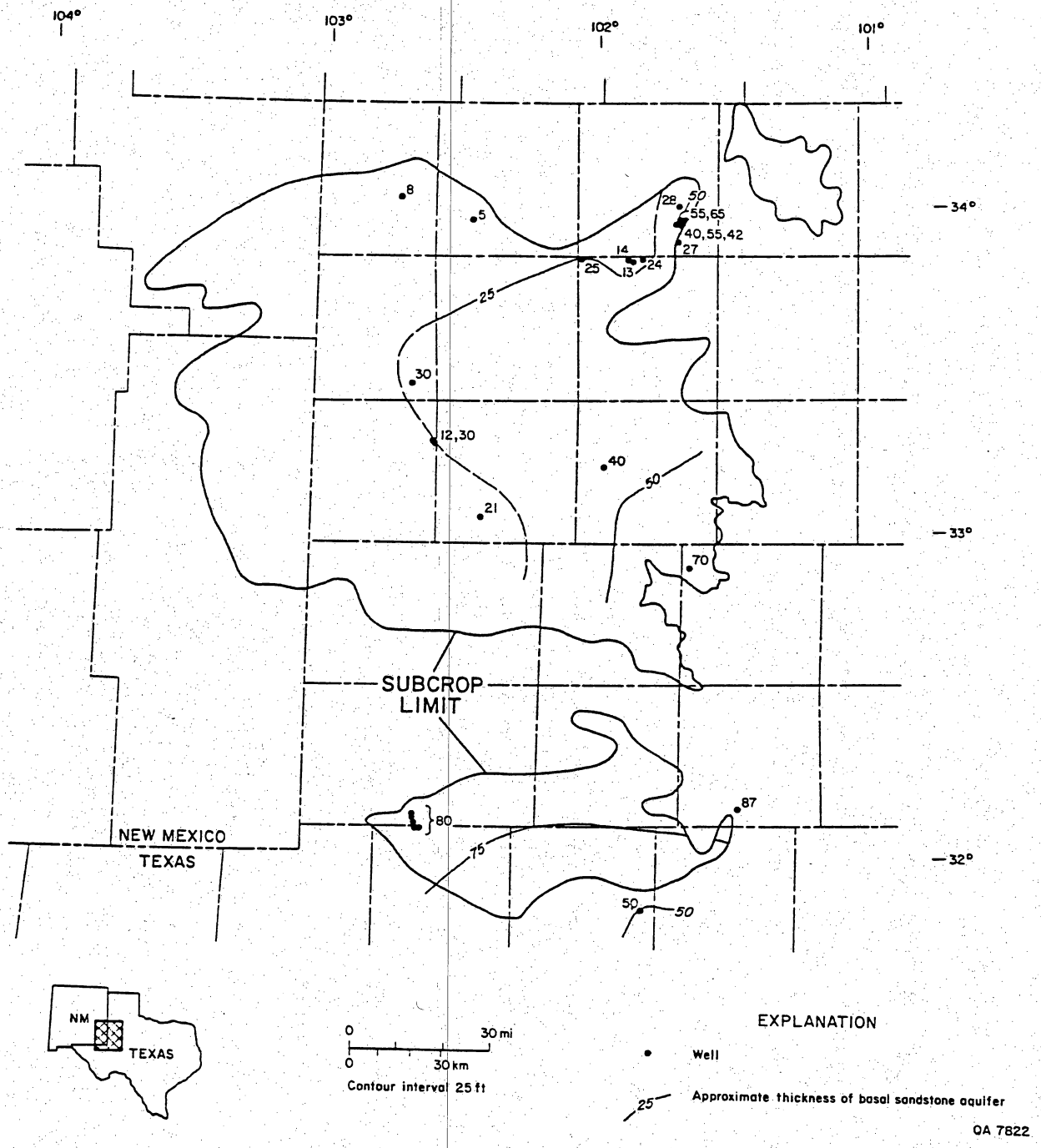


Figure 36. Thickness of basal Cretaceous sandstone aquifer. From Nativ and Gutierrez (1987).

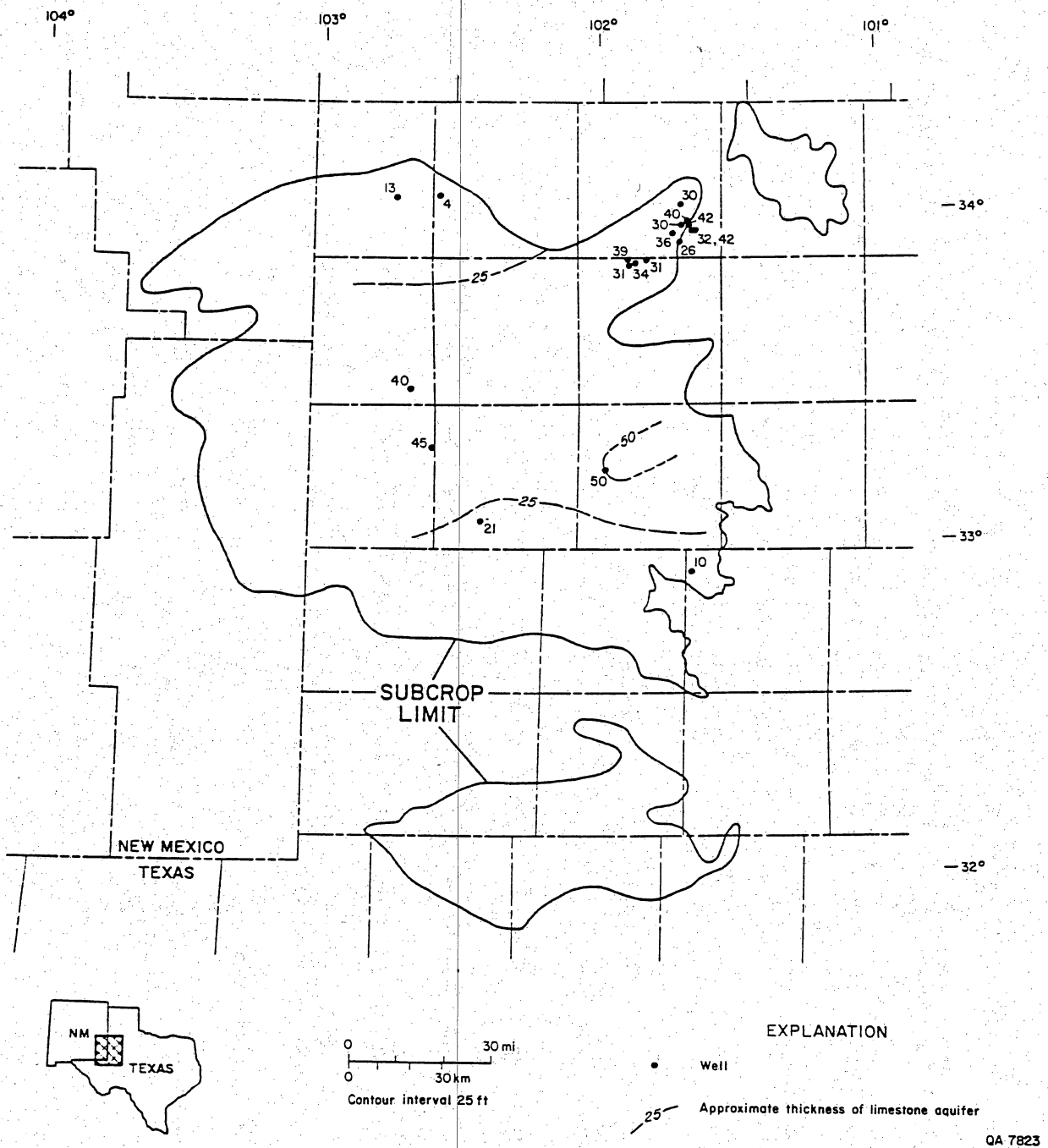


Figure 37. Thickness of the Cretaceous limestone aquifer. From Nativ and Gutierrez (1987).

Table 2. Cretaceous Comanchean section, Southern High Plains, Texas (modified from Brand, 1953).

Group	Formation	Description
Washita	Duck Creek	Thickness: 0 to 40 ft (0 to 12 m). Shale, yellow-brown to dark gray, and thin gray to brown limestone and sandstone; *Desmoceras brazoense* at base. Unconformity at base.
Fredricksburg	Kiamichi	Thickness: 108 ft (33 m). Shale, dark gray to yellow-brown and thin light gray to yellow-brown limestone and sandstone. Lower part contains numerous specimens of *Gryphaea navia*; upper part *Gryphaea tucumcari*, *Oxytropidoceras* spp. throughout.
	Edwards	Thickness: 20 to 30 ft (6 to 9 m). Limestone, rudistid, white to yellow-brown. Forms escarpment in southeast.
	Comanche Peak	Thickness: 50 to 80 ft (15 to 24 m). Limestone, argillaceous, thick to massive, light gray with light gray marly interbeds.
	Walnut	Thickness: 8 to 23 ft (2 to 7 m). Shale and sandy shale, light gray and light gray argillaceous limestone. *Gryphaea mucronata* and *Ostrea crenulimargo* abundant. Unconformity at base.
Trinity	Paluxy	Thickness: 10 to 25 ft (3 to 8 m). Sand and lenticular quartz gravel beds, light gray to purple. Triassic shale pebbles in basal part. Unconformity at base over the Triassic System.

Essentially none of the research objectives for study of the subsurface and surface Cretaceous strata have been fully accomplished. Further study of these deposits seems warranted for at least two reasons: (1) to establish the interrelationship of Cretaceous-rock aquifers to the High Plains aquifer for better management of water resources and control of potential ground-water contamination, and (2) to determine the role of Cretaceous deposits in the initiation and development of so-called ground-water lakes, described later in this section. Some of the ground-water lake basins are inset into Cretaceous strata and are bounded on the west margins by Cretaceous exposures.

Ogallala Formation

Introduction

Stratigraphic studies of the Ogallala Formation have had various objectives, including (1) to determine the spatial and temporal distribution of Ogallala deposits, (2) to delineate and characterize the petrography and paleobiology of each significant facies, (3) to interpret the depositional environments of each facies and to determine temporal and spatial facies relationships throughout the Southern High Plains, (4) to identify, describe and evaluate the significance of diagenetic and pedogenetic aspects of the deposits, and (5) to attempt to establish a chronostratigraphic framework for the Ogallala based on paleontological evidence, pedogenetic characteristics, magnetopolarity, radiometric dating and tephrochronology.

The Ogallala Formation of the Southern High Plains is an erosional remnant of a larger sheet of continental sediments derived from the Rocky Mountains and laid down on a nearly mature erosion surface of Permian, Triassic and lower Cretaceous strata (fig. 38). Deposition of the Ogallala occurred during the Late Miocene (13 to 12 m.y.) and earliest Pliocene (5.0 to 4.5 m.y.). Thickness of the Ogallala ranges from a featheredge to more than 800 ft (250 m) (fig. 39). According to Seni (1980), thickness variations are related to irregularities of the pre-Ogallala surface (fig. 40). He stated that Ogallala thickness was influenced by (1) subsurface geology, (2) structure, (3) facies associations, and (4) post-depositional erosion. Thickest Ogallala deposits are north and east of Amarillo, Texas, where the Ogallala-Permian contact is characterized by collapse basins caused by solution of Permian evaporites, a process that Seni (1980) speculates was active during Ogallala deposition (see Gustavson, this volume). These collapse features range in area from 200 mi^2 (500 km^2) for a basin in Carson County filled with 800 ft (250 m) of Ogallala sediments, to chimneys 10 ft (3 m) in diameter. He also noted that salt dissolution

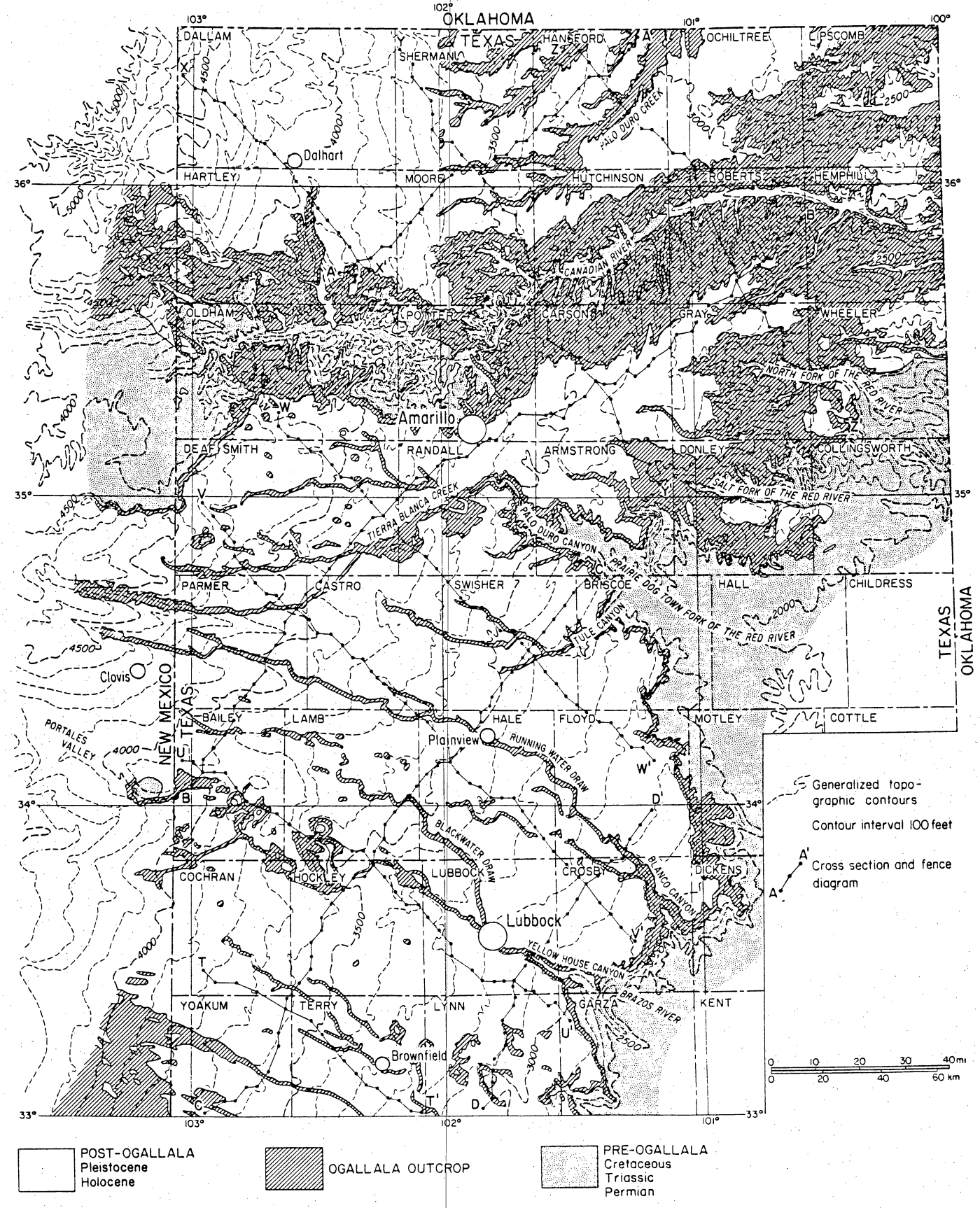


Figure 38. Geologic map of study area with locations of cross sections, creeks, rivers, canyons, and towns cited in text. After Seni (1980).

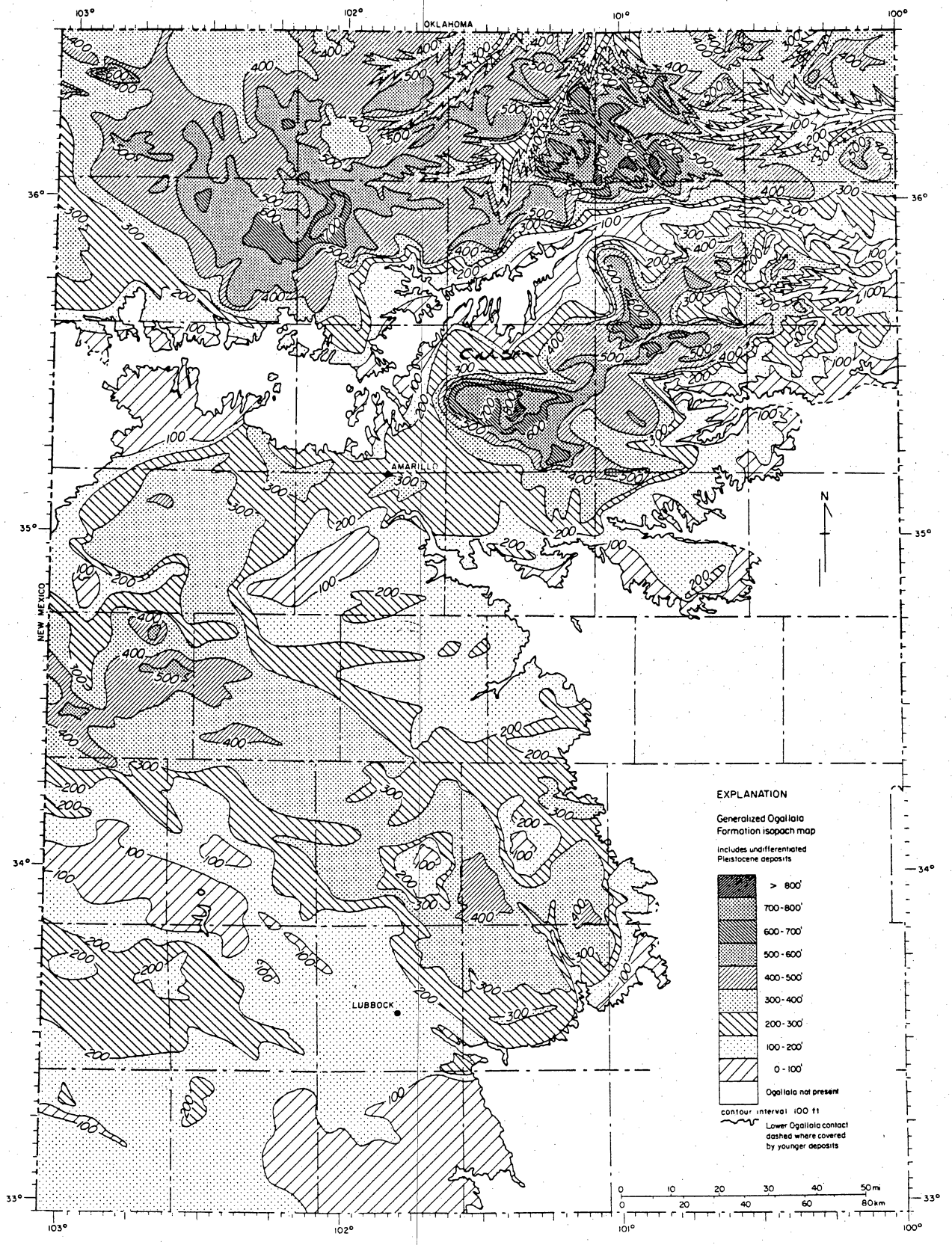


Figure 39. Isopach map, Ogallala Formation. Ogallala thickness is defined as net thickness of all deposits above base of the Ogallala Formation. Undifferentiated Pleistocene deposits, normally less than 20 ft (7 m) thick, may be included in this total. After Seni (1980).

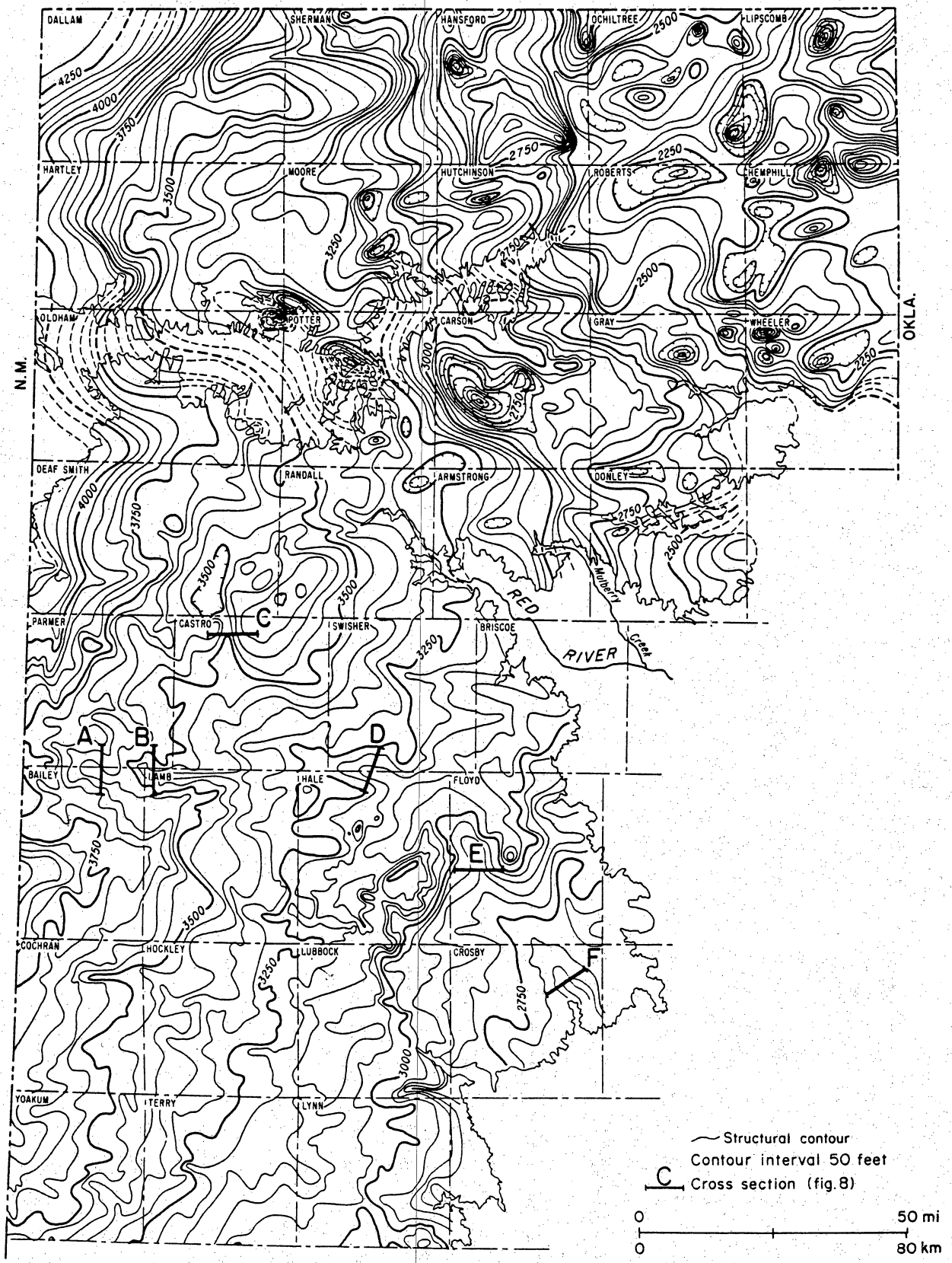


Figure 40. Structure contour map, base of Ogallala Formation. South of northern boundary of Deaf Smith, Randall, Armstrong, and Donley Counties. After Seni (1980).

had little effect on Ogallala thickness in the central part of the Southern High Plains where 200 to 2,000 ft (60 to 600 m) of relatively impermeable Triassic strata lie between the Ogallala and Permian evaporites.

The High Plains of Texas and New Mexico has risen to its present elevation within the past 10 million years, and was probably much lower in elevation during most of Ogallala time (Hawley, 1984). Pleistocene sediments overlying the Ogallala have a much higher dip than those of the Ogallala (Winkler, 1985).

An extensive erosional surface underlies the late Neogene Tertiary Ogallala and cuts across Permian through Cretaceous strata. Throughout much of the Southern High Plains the Ogallala is overlain by the Blackwater Draw Formation. The nature of this contact, whether predominantly unconformable or transitional, is unresolved.

Stratigraphic Classification

An excellent summary of the stratigraphic nomenclatural history for deposits now assigned to the Ogallala Formation in the Southern High Plains is presented by Winkler (1985).

Darton (1899, p. 734) used the name Ogallala Formation for, ". . . a calcareous formation of late Tertiary age exposed in western Nebraska and extending into Kansas and Colorado." A type locality was designated by Elias (*in* Stirton, 1936, p. 178) as exposures in the canyon two miles east and 0.5 mile north of the town of Ogallala, on the Feltz Ranch in Keith County, Nebraska. Elias applied the name Ogallala Formation to all late Tertiary beds, primarily arenaceous, that overlie the Arikaree and that range from late Miocene to possibly late Pliocene age (Schultz, 1977). The formation extends 900 miles (1,448 km) from South Dakota to northwestern Texas, and has a maximum width of 300 miles (483 km) (Frye, 1971). The Ogallala underlies most of the High Plains, an area of approximately 29,000 mi^2 (~75,000 km^2) (Gustavson and Winkler, 1988).

Elias (1942) noted that fossil seeds of grass and other plants form stratigraphically significant assemblages. Frye and others (1956) refined the technique of Elias and subsequently utilized fossil seed zones for internal stratigraphic subdivision of the Ogallala exposed along the High Plains escarpment in Texas. These authors also used fossil plants, physical stratigraphy, and evidence from scattered vertebrate and molluscan assemblages to correlate the floral zones of Texas with the floral zones and members of the Ogallala in Kansas.

Ogallala deposits of Nebraska were classified as a group by Lugn (1938, 1939). The four included formations, from the bottom upward (as originally perceived) are the Valentine (Johnson, 1935), Ash Hollow (Engleman, 1876; Lugn, 1939), Sidney (Lugn, 1938), and Kimball (Lugn, 1938, 1939). In Kansas these units were classified as members or facies of the Ogallala Formation (Moore and others, 1951). Frye and Leonard (1957b, 1959) could not distinguish the Valentine, Ash Hollow, and Kimball as mappable lithologic units in Texas so they applied these names to floral zones that they thought were roughly equivalent to the members and formations in Kansas and Nebraska.

A detailed biostratigraphic analysis of the Ogallala floral zones in Kansas as established by Elias (1942) and Frye and Leonard (1959) led Thomasson (1979) to conclude that past interpretations of the "seeds" in Ash Hollow and Kimball floral zones were not tenable and subsequently he has defined biostratigraphic units for the Ogallala based on assemblage zones. Previously Galusha (1975) had questioned the validity of the Valentine floral zone as conventionally applied in Nebraska. On the basis of mammalian evidence, Skinner and Johnson (1984) concluded that the Valentine Formation of Nebraska is of Late Barstovian age (~13 m.y.).

Some geologists have differentiated members within the Ogallala of the Southern High Plains and others have treated the Ogallala as a group comprising more than one formation. Such classification schemes generally have not found application outside their type areas.

In Crosby, Floyd and Lubbock Counties, Texas, the Ogallala consists of two distinct and mappable units (Evans, 1949, 1956; Winkler, 1985, 1987). Evans (1949) informally named the lower of these units the Couch Formation and the upper, the Bridwell Formation. Subsequently, Evans (1956) raised the Ogallala to the rank of a group in that part of the High Plains where the Couch and Bridwell formations could be distinguished.

Evans (1949) provided the following definition of the Couch Formation:

> The Couch formation, to be named for the Couch ranch in Blanco Canyon, Crosby County, is composed of compact, well-sorted calcareous sand and gravel. The formation rests unconformably on Triassic and Cretaceous rocks and is separated by an unconformity from the overlying Bridwell formation. At the type locality, about 7 miles east of Crosbyton, the formation is 125 ft thick, but its thickness varies markedly with the relief on the underlying bedrock surface. The formation has its maximum development in the deeper pre-Cenozoic valleys, and thins and wedges out against higher ridges and divides on the bedrock surface.

Evans (1956) described, but did not name, two members of the Couch Formation. A basal member, recognizable in most places, is a cross-bedded deposit of sand and gravel occupying very broad, relatively shallow channels, and an upper member is composed of well-sorted, semi-consolidated calcareous and clayey sands. He speculated that sediments of the upper member were derived from stream deposits, sorted, and redeposited by the wind. On the basis of vertebrate fossil evidence Evans assigned both members to the Early Pliocene.

The name Bridwell Formation was proposed by Evans (1949) for deposits seven miles east of Crosbyton, Texas, in Blanco Canyon on the D. R. Couch Ranch, but the formation is named for Bridwell Ranch in northern Blanco Canyon. Evans (1949, p. 6) described the formation as follows:

> . . . bedded, unconsolidated sand and clay with thick channel gravel occurring both at the base and at higher levels in the formation. Except for a few gray, highly calcareous members, including the caliche caprock, the Bridwell sediments are characteristically reddish-brown and contrast conspicuously with the much lighter underlying Couch formation. The basal gravel rests unconformably, occasionally in shallow channels, on the Couch. The upper surface, approximately the caliche caprock horizon, suffered local erosion at different times during the Quaternary and perhaps in the late- pre-Quaternary.

According to Evans (1956) the Bridwell averages about 155 ft (47 m) thick in Crosby County, but thins to the south and southwest and may thicken in some of the deeper valleys. He stated that where the Bridwell overlaps the Couch it rests directly on Cretaceous or Triassic strata. Generally it is unconformably overlain by Quaternary deposits but in places it forms the High Plains surface. Evans (1949) concluded that the Bridwell originated as channel and floodplain facies of broad, aggrading valleys. After cessation of fluvial deposition, caliche caprock developed as a subsoil lime accumulating on the alluvial plain. On the basis of its contained vertebrate fossils, Evans assigned the Bridwell to the Middle Pliocene.

Winkler (1985) made a detailed study of the stratigraphy, vertebrate paleontology, and depositional history of the Ogallala deposits exposed in Blanco and Yellowhouse Canyons, Crosby, Lubbock and Garza Counties, Texas. Winkler was able to distinguish and map the Couch and Bridwell Formations of Evans (1949, 1956) and he agreed that in the area of study the Ogallala should be assigned the status of a group. He cautioned, however, that these units are probably not distinguishable throughout the entire Southern High Plains. Winkler (1985) also recognized the two members of the Couch Formation delineated previously by Evans (1949, 1956) and informally named the upper unit the Crosbyton member for exposures in Blanco Canyon

east of Crosbyton, Texas. He did not name the lower unit because of its limited outcrops and because in places it is difficult to differentiate its coarse-grained facies from those of the Bridwell Formation.

In Potter County, Texas (see fig. 30 for location), Patton (1923) divided the Ogallala beds into two formations. He called the lower unit the Potter Formation and the upper, the Coetas Formation. The exact nature of the Potter-Coetas contact was not observed by Patton but he noted that the Potter unconformably overlies both Permian and Triassic strata. Deposits stratigraphically above the Coetas were termed "Undifferentiated Cenozoic" by Patton.

The name Potter Formation was proposed by Patton (1923, p. 78) for:

> . . . a group of sediments characteristically developed along the Canadian River, near the central part of the county (Potter) and which bear a definite relation to the sediments to which the name Coetas formation is given. The term Potter formation is applied to those beds of coarsely stratified and partly consolidated sand and gravel which can be shown to occupy a stratigraphic position, just below the Coetas formation, or which show a definite relationship to beds which do occupy such a position. These beds have been classed by other writers as "mortar beds" on account of their resemblance to the artificial product, as they are somewhat firmly cemented by calcium carbonate.

Patton emphasized that he did not intend to classify all mortar beds as Potter, but only those directly overlain by the Coetas Formation. He further characterized the formation as being somewhat reddish and darker than the overlying beds. It contains a large number of water-worn, reworked specimens of *Gryphaea*. The gravel ranges from pieces several inches in diameter to a fraction of an inch. About 80 percent of the pebbles consist of quartz and quartzite and lesser numbers of sedimentary and igneous clasts. The formation is strongly crossbedded in many localities. Because the

unit is resistant to erosion it is a cliff former in the area. No indigenous fossils were noted by Patton. Because the Coetas had been dated as Pliocene on its contained vertebrate assemblage, Patton concluded that the Potter was either of Pliocene or pre-Pliocene age.

Patton (1923) used the name Coetas Formation for slightly consolidated sands and fossiliferous, flaggy limestones that stratigraphically overlie the Potter Formation. The name was derived from Coetas Creek in the eastern part of Potter County. From a composite section of the unit exposed in the type area he determined its thickness to be approximately 230 ft (70 m). Patton apparently was the first to report on the presence, in these beds, of vertebrate fossils which he said were poorly preserved and scattered through both the sand and limestone facies. On the basis of the paleontological evidence he assigned a Pliocene age to the formation and concluded that it was a correlative of the Blanco beds. Post-Coetas sand, gravel and clay deposits of the area which in places lie directly upon the Coetas Formation were referred to by Patton as "the undifferentiated Cenozoic formation." He pointed out that a very prominent feature of the undifferentiated Cenozoic sediments is the caliche or caprock of the formation.

Wilson (in press) studied outcrops of lower Ogallala (Potter and Coetas formations of Patton, 1923) sediments exposed along the Canadian River valley north of Amarillo. He concluded that these sediments were deposited on a pre-Ogallala valley system that was influenced by surface structure of the John Ray Dome, by the scarp developed at the erosional limit of the Triassic strata, and by subsidence induced by dissolution of Permian salt beds. The Ogallala is represented in the area by the basal Potter member, a 0- to 150-ft- (0- to 46-m-) thick braided stream deposit; the middle LX member, (0- to 100-ft [0- to 30-m] thick), eolian silty sand sheet; and the Coetas member, a 0- to 60-ft- (0- to 18-m-) thick upper lacustrine deposit. The highly eroded Coetas member is unconformably overlain by undifferentiated Ogallala sediments.

Deposition of the three lower members was restricted to a narrow paleoerosion valley along the southwest flank of the John Ray Dome. East of the dome the paleovalley widens where the Potter stream flowed into the Carson County collapse basin. In this basin thickness of the lower three members is 500 ft (152 m) and the entire Ogallala section is 800 ft (244 m) thick. The lower Ogallala members are present only in the paleovalleys and collapse basins. Above the Triassic paleo-scarp to the south, upper (undifferentiated) Ogallala sediments, 0 to 200 ft (0 to 61 m) thick, blanket the region and lie unconformably on the Triassic strata.

Wilson (in press) noted that post-depositional collapse basins along the Canadian River contain Ogallala strata that dip 20 degrees into the depressions. Pre-Ogallala collapse features are evident but major collapse structures on the south side of the John Ray Dome occurred after deposition of the Coetas member. Some of the basins contain undisturbed, thickened sections of undifferentiated upper Ogallala sediments, whereas in other basins these upper deposits dip into the basin indicating post-depositional subsidence.

Wilson observed that major collapse basins are underlain by structural lows and parallel large subsidence faults. Budnik (1987) suggested that at least some of the Ogallala deformation near the John Ray Dome was caused by Miocene tectonic activity involving upward movement of the Amarillo Uplift in response to Basin and Range extension and opening of the Rio Grande rift in central New Mexico and Colorado.

Mammal Ages

Ogallala deposits of the Texas High Plains include the type localities of two North American Mammal Ages, the Clarendonian and the Hemphillian. The Clarendonian Mammal age was defined by Wood and others (1941) based on the Clarendon local fauna located near Clarendon, Donley County, Texas (see fig. 30 for location). The Clarendon type fauna is used as a unit in correlation although it has

not been biostratigraphically characterized and is one of the least understood of Clarendonian faunas.

According to Schultz (in press):

> The Clarendon fauna is representative of what has been termed the Clarendonian chronofauna (Webb, 1969, 1977, 1983; Tedford, 1970), a coherent association of species lineages dominated by ungulates that emerged in North America during the late Barstovian, about 15 million years ago, reached its peak in the Clarendonian and declined by the end of the Hemphillian, around 5 million years ago. The rise and fall of this late Miocene chronofauna were apparently controlled by a late Cenozoic trend toward cooler and drier climates at temperate latitudes (Webb, 1983). As forest biomes gave way to parkland savanna, the abundance and diversity of ungulates and the ratio of grazers to browsers increased. Later, however, the trend toward increasing aridity and the spread of steppe conditions led first to the extinction of virtually all browsers, secondly to the decimation of grazers, and finally to the wholesale destruction of the chronofauna by the end of Hemphillian time. Hemphillian faunas, especially later ones, show a marked decrease in diversity compared with Clarendonian faunas. There is a remarkably close resemblance between the late Miocene ungulate fauna of North America and the Recent ungulate fauna of the African savanna despite their entirely independent origins (Webb, 1983).

The Clarendon fauna is thought to correlate with the Ash Hollow faunas of north-central Nebraska and south-central South Dakota and to be Late Clarendonian in age (Schultz, 1977, in press).

One volcanic ash bed is known in the Clarendon region. It is located in the bed of Petrified Creek, 0.5 mi (0.8 km) north of Gidley's three-toed horse quarry, but is not in superpositional sequence with the fossil-bearing strata. According to Schultz

(1977, in press), Boellstorff (personal communication to Schultz) obtained a glass-fission date of 8.1 ± 0.5 m.y. for this ash. If this date is correct the ash lies close to the Clarendonian- Hemphillian boundary, and apparently would be of early Hemphillian age. Tedford and others (1987) gave an age range for the Clarendonian of 11.5 to 9 m.y., and place the division between Early and Late Clarendonian at about 10 m.y.

Reed and Longnecker (1932) first applied the name "Hemphill beds" to upper Ogallala deposits in Hemphill County, Texas. The Hemphillian Mammal age was later defined by Wood and others (1941) based on the local fauna at Coffee Ranch Quarry in Hemphill County. A section at the Coffee Ranch, measured and described by Dalquest (1969), is recorded in table 3.

Dalquest noted that fossils occur in all the beds but were most abundant in unit 6. He thought that sediments at this site were deposited in a lake or a bog later buried beneath 30 ft (9 m) of eolian sands.

Wood and others (1941) considered the Hemphillian to be of middle Pliocene age. Berggren and Van Couvering (1974) gave the age range of the Hemphillian as about 8.5 to 9 m.y. to about 4.5 m.y.; thus the Hemphillian ranges from Late Miocene to Early Pliocene. However, the youngest dated Hemphillian deposits in North America are about 5.0 to 5.2 m.y. (Lundelius and others, 1987).

Dalquest (1969) considered the Coffee Ranch Local Fauna to be a unit fauna, which represents vertebrates that lived at the site for perhaps only a few centuries, where a closed basin formed and then filled with sediment.

According to Schultz (in press), Hemphillian Local Fauna correlates with the Optimal (=Guyman) Local Fauna of Oklahoma (Hesse, 1936; Savage, 1941), the Rhino Hill and Edson faunas of Kansas (Harrison, 1983), the ZX Bar Local Fauna in the upper part of the Snake Creek sequence of northwest Nebraska (Skinner and others, 1977), the Camel Canyon, Redington, Wikieup and White Cone Local Faunas of

Table 3. Upper Ogallala section, Coffee Ranch quarry, Hemphill County, Texas (after Dalquest, 1969).

Unit	Description	Thickness (ft)
8	Overburden of buff-colored sandy clay and soil	25
7	Volcanic ash	9
6	Compact bentonitic clay	2
5	Greenish gray sand with some clay	2
4	Reddish brown, sandy clay, variable in thickness in the deposit and with sharp but contorted contact with beds above and beneath	1
3	Greenish sand and sandy clay with some pebble bands and thin calcareous sandstone layers	5.5
2	Slick, hard reddish brown clay with calcareous crusts and nodules	0.2
1	Buff-colored eolian sandy sediments of the Ogallala Formation; bottom not exposed in the area.	

Arizona and Chamita fauna of New Mexico (Lindsay and others, 1984), and the Upper Bone Valley fauna of Florida.

Izett (1975) dated the ash which overlies the Coffee Ranch Quarry using the fission track method on zircon and glass shards. Glass-mantled zircon microphenocrysts yielded a date of 6.6 $\pm$ 0.8 m.y. and glass shards a date of 4.7 $\pm$ 0.8 m.y. Boellstorff (1976) obtained a glass fission track date of 5.3 $\pm$ 0.4 m.y. for the same ash. Lindsay and others (1975) determined that the fossil-bearing sands and the ash bed are normally magnetized and concluded that they should be included in magnetic epoch 5 because the ash date indicates it is older than the Gause Epoch (2.4 to 3.3 m.y.). Schultz (1977) concluded that the faunal, numerical dates and evidence of magnetic polarity studies place the Coffee Ranch Local Fauna in the Late Hemphillian (fig. 41).

Origin of Ogallala Deposits

Most authors, including Johnson (1901), Sellards and others (1932), Smith (1940), Bretz and Horberg (1949a, 1949b), Frye and Leonard (1964), Frye (1971), Seni (1980), Hawley (1984), and Reeves (1984) have considered the Ogallala of Texas and New Mexico to be composed primarily of fluvial sediments that originated as a series of deposits from ephemeral streams or as a great alluvial plain or bahada. It was commonly recognized that an erosional surface with deep, wide valleys was invaded and buried during the late Tertiary and that the present High Plains surface is a reflection of the depositional slope of the Ogallala. In general the importance of eolian sedimentation had not been recognized until pointed out by Gustavson and Holliday (1985). However, Evans and Meade (1945), Reeves (1972) and Winkler (1985) have described substantial eolian sections in the Ogallala.

Frye and Leonard (1959) disagreed with the commonly proposed hypothesis that the Ogallala plain was developed by a series of adjacent gigantic fans that spread out

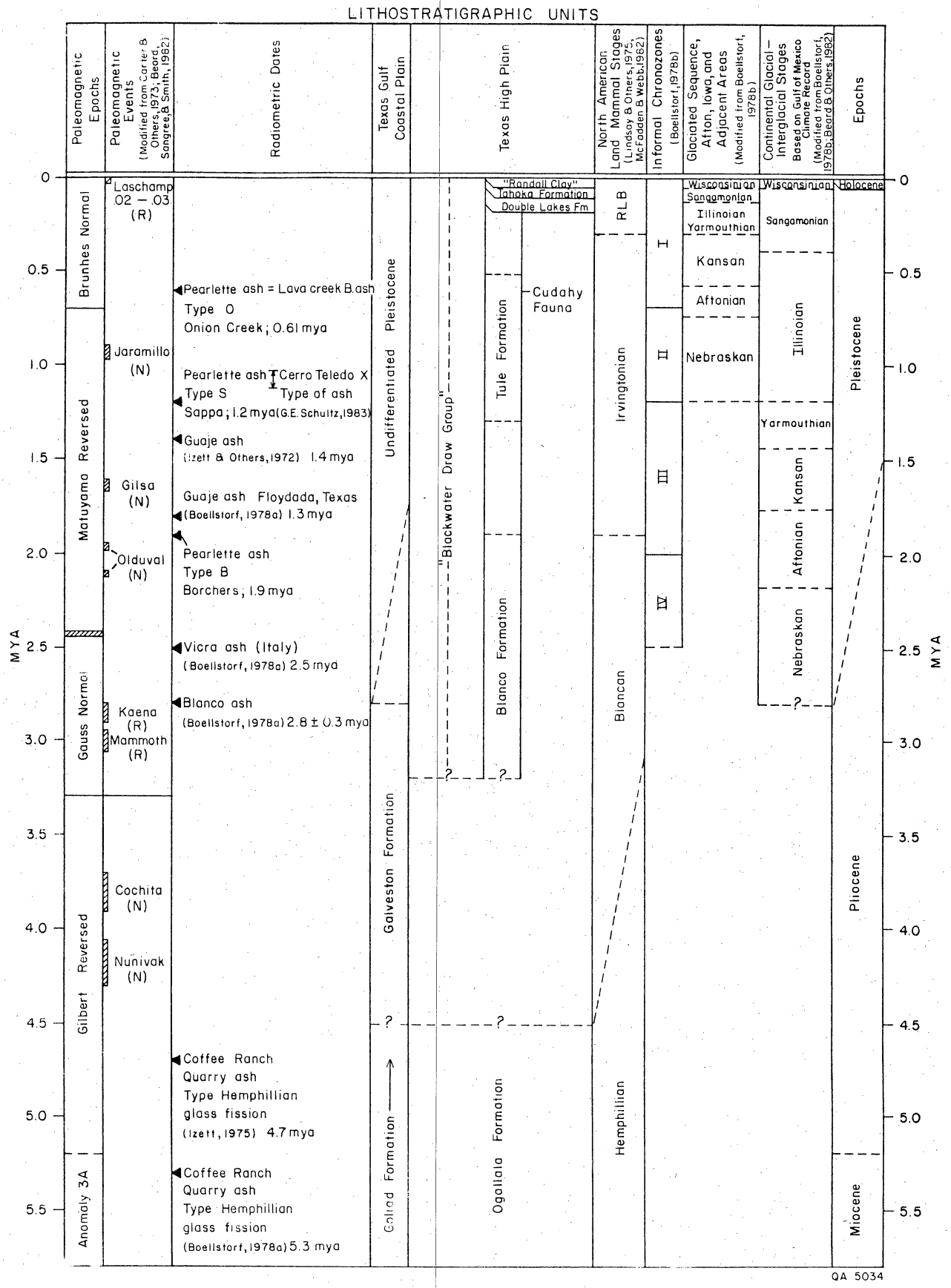


Figure 41. Late Cenozoic stratigraphy of the Southern High Plains.

over an essentially planate surface. They considered it apparent that, if such were true, the oldest deposits in the alluvial complex should exist in the area of initial deposition, the occurrence of these oldest deposits should be roughly aligned in a north-south direction parallel to the source area, and there should be progressive overlap of younger beds toward the east. Frye and others (1956) traced zones of several ages for considerable distances west to east and to show that the younger units rather than progressively overlapping older deposits to the east displayed a north and south overlapping relationship on the gently sloping sides of broad and relatively shallow pre-Ogallala valleys. These data led to the conclusion that the mechanism of Ogallala deposition was not that of alluvial fan building, but rather was one of valley alluviation (Frye and Leonard, 1959, p. 7).

Seni (1980) concluded that the Ogallala Formation of the Southern High Plains is a wet fan system that fringed the eastern flanks of the Rocky Mountains. He considered the system analogous to the modern Kobe River alluvial fans in Nepal and India.

Using data from more than 15,000 drillers' logs supplemented by a reconnaissance investigation of outcrops along the Ogallala escarpment, Seni (1980) constructed cross-sections, an Ogallala isopach map, and a sand and gravel percentage map. From this data base Seni recognized three overlapping Ogallala fan lobes developed over pre-Ogallala valley systems. He determined that the lobes are composed of sediment eroded from the Rocky Mountains and transported east, south and southeast. From north to south the three lobes are the Dalhart-Amarillo, developed in a low on the Permian and Triassic strata; the Clovis-Plainview, overlying Triassic beds; and the Brownfield-Lubbock, deposited on Cretaceous strata (fig. 42).

Seni (1980) concluded that the three overlapping lobes were deposited and abandoned in sequence from north to south. He reasoned that the successively southward shift of lobe formation was governed primarily by the step-like, southward increase in elevation of the subjacent pre-Ogallala strata.

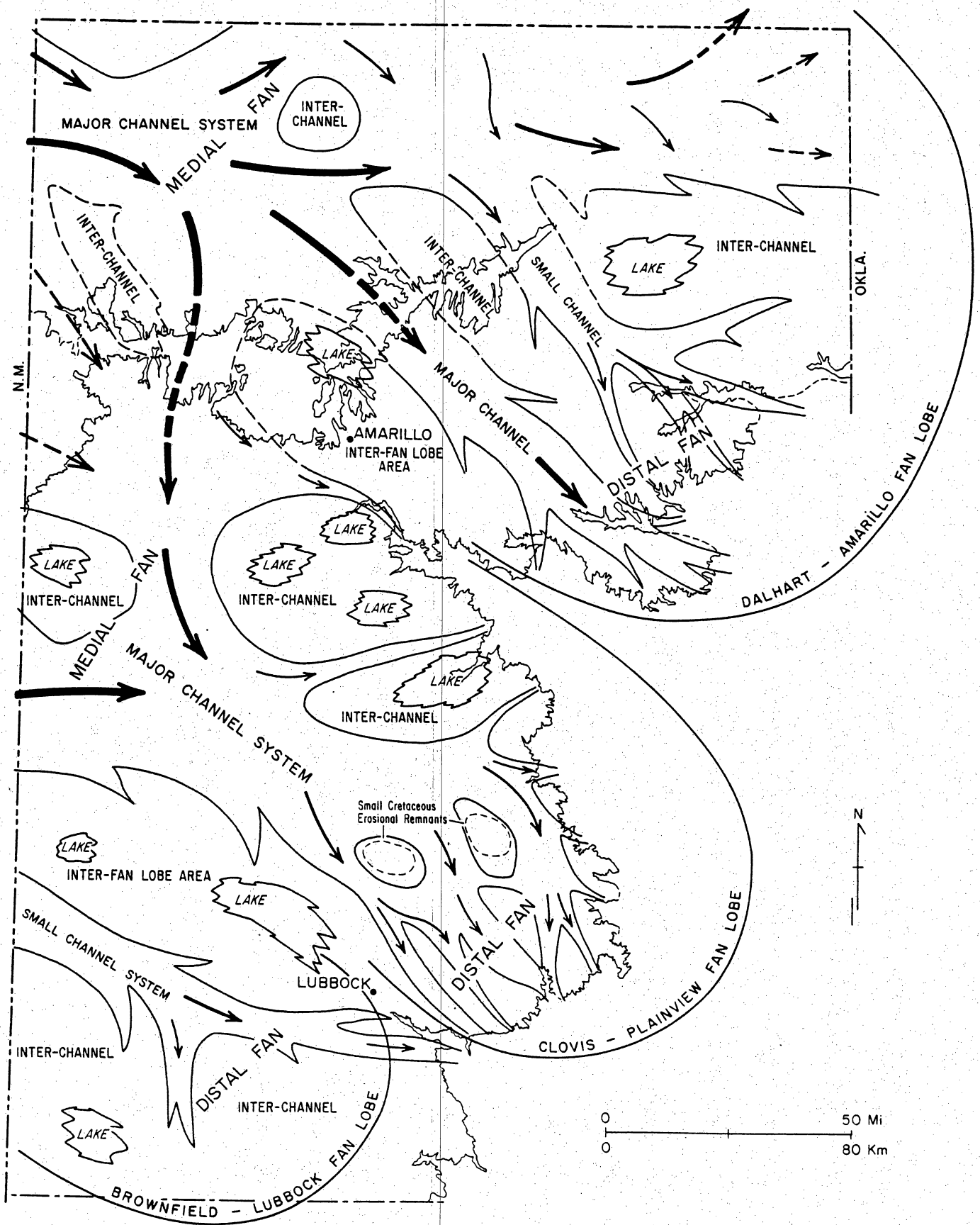


Figure 42. Schematic illustration of Ogallala depositional facies and sediment dispersal systems. Width and length of arrows indicate relative intensity of fluvial processes. Facies are not time-equivalent. After Seni (1980).

Seni (1980) used fossil flora data reported by Frye and Leonard (1957b, 1959) to assign relative ages to the three fan lobes. He stated that the oldest Ogallala floral zone (Valentine) is thickest in the Dalhart-Amarillo lobe, and that the youngest floral zone (Kimball) is characteristic of sediments of the Brownfield-Lubbock lobe. Although not stated by Seni, one may surmise that he considered the Clovis-Plainview lobe to be primarily of an intermediate age (Ash Hollow Floral Zone?). It is noteworthy that Frye and Leonard (1964) reported the occurrence of the Valentine floral zone as far south as Floyd County within the area assigned by Seni to the Clovis-Plainview lobe. Also, these authors concluded that southernmost Ogallala floral assemblages all indicate deposition during middle and late Ash Hollow time.

Reeves (1984) disagreed with the Seni (1980) depositional model which depicts Ogallala sediments as primarily the product of a series of overlapping fan lobes of different ages. From a synthesis of available evidence including data derived from clay mineralogy, invertebrate and vertebrate fossils, floral zones, petrology of Ogallala gravels, water well logs and study of outcrops in New Mexico he concluded that: (1) Ogallala fans of the Southern High Plains south of the Canadian River were constructed by three separate northwest to southeast flowing fluvial systems, (2) a fourth, newly recognized, fan exists in the southernmost part of the southern fluvial system, (3) Ogallala fans are coalescent and are thus of the same age, (4) the Ogallala Formation of the Southern High Plains consists of broad areas of eolian sands, and (5) climate of the Southern High Plains during Ogallala time was semiarid and not wetter than today.

In their study of the depositional facies of the Ogallala Formation in northwestern Texas and eastern New Mexico, Gustavson and Winkler (1988) examined 36 exposures located along the Caprock Escarpment and in stream cuts and quarries, and the core from the DOE-Gruy Federal Grabbe No. 1 stratigraphic test well. Lithologic

descriptions, facies interpretations and age data for these and other sections are included in Gustavson and Holliday (1985), Gustavson and Winkler (1988) and Winkler (1985).

Analyses of data derived from these sections led Gustavson and Winkler (1988) to conclude:

> . . . that the Ogallala of Texas and eastern New Mexico consists of alluvial sediments that partly fill paleovalleys, and widespread thick eolian sediments capping paleo uplands and mostly fluvial sections. These deposits were apparently laid down under arid to subhumid climatic conditions and do not constitute coalescing or overlapping wet alluvial fans.

Gustavson and Winkler (1988) found that paleo-uplands and most fluvial facies in paleovalleys are overlain by thick sequences of silty, very fine sand or loamy sand. Silt content of these sediments ranges from 4 to 40 percent; clay content ranges from 0 to 15 percent. Where measured these very fine sands are 100 and 124 ft (30 and 38 m) thick over paleo-uplands near Silverton, Texas and at Buffalo Lake National Wildlife Refuge southwest of Canyon, Texas. The very fine sands of paleo-uplands have yielded no datable materials; however, Gustavson and Winkler found that thick sequences of very fine sand composed upper parts of the Ogallala in most of the 37 exposures and core (fig. 43) and that they are capped by the Caprock, the top of the formation.

In exposures of fine sandstone, primary sedimentary structures are generally absent except for intervals of wave-ripple laminae and eolian dune cross-bedding. Coarse sand grains are commonly frosted and well rounded to subrounded. In many areas siliceous and calcareous tubules (rhizoconcretions) that mark the former presence of rootlets are preserved. Numerous pedogenic calcretes (buried calcic soil horizons) are preserved within these sands and are characterized by $CaCO_3$ glaebules up to 1 in (2.5 cm) in diameter, coalesced glaebules, rhizoconcretions, and massive carbonate

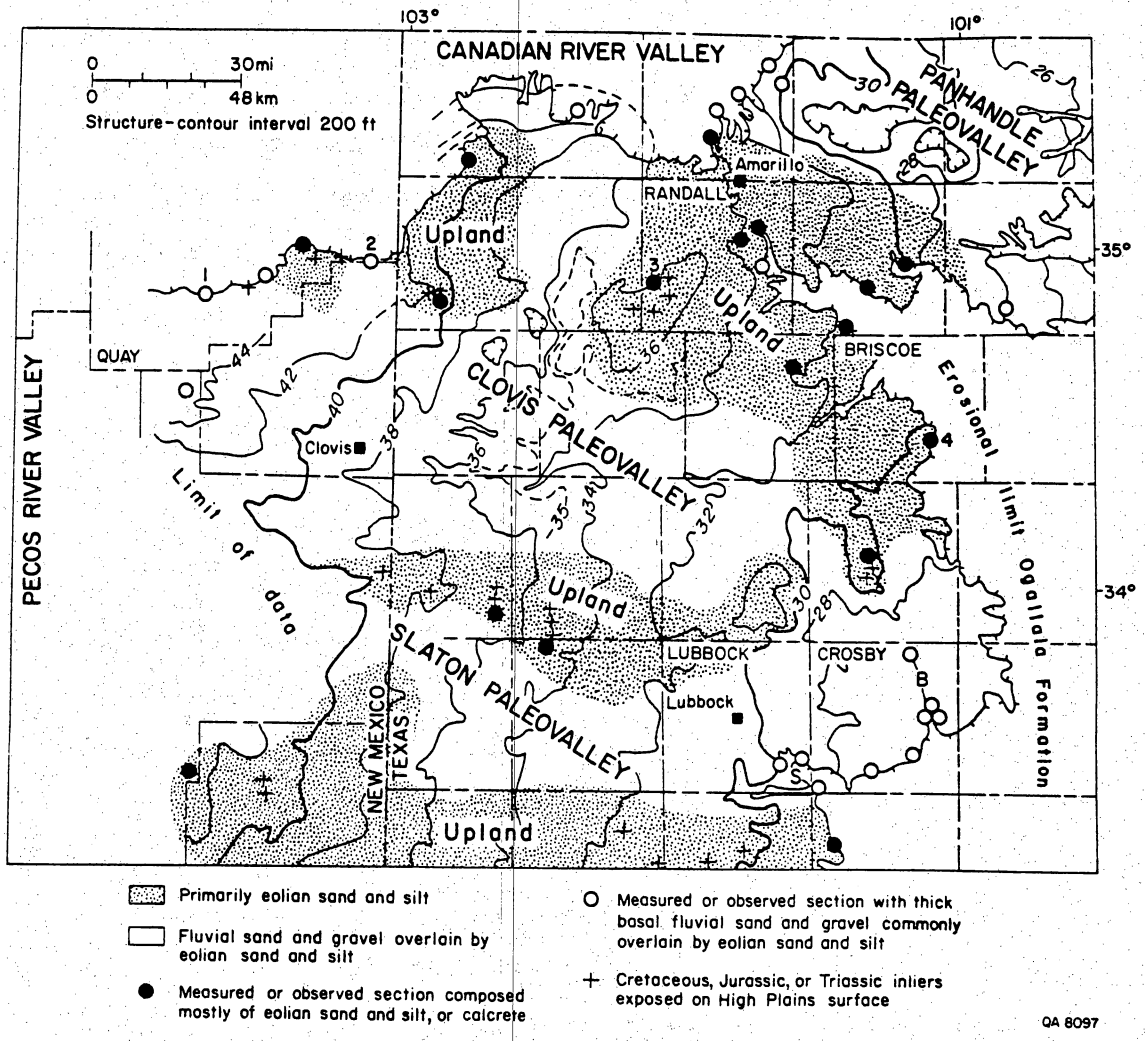


Figure 43. Paleotopographic map of pre-Ogallala erosional surface. From Gustavson and Winkler (1988).

zones that locally are laminated, brecciated, or pisolitic. Less commonly, buried Bt horizons characterized by accumulations of illuvial clay are preserved. Paleosols show no significant thickness changes over distances of as much as 0.5 mi (0.75 km).

Gustavson and Winkler (1988) concluded that deposition by eolian processes is suggested by grain size, presence of rounded and frosted quartz grains, and rare eolian sedimentary structures. However, Ogallala eolian sediments have more silt- and clay-size material than typical sand sheet deposits (Fryberger and others, 1979; Kocurek and Nielson, 1986). Conversely, these sediments have much more coarse material than a typical loess (Miller and others, 1984). Gustavson and Winkler (1988) thought that bioturbation and pedogenesis were probably responsible for the obliteration of sedimentary structures and that these processes homogenized sediments originally deposited as sand sheets and loess units.

Gustavson and Winkler (1988) reasoned that initial source areas for the eolian sediments were floodplains of streams that deposited Ogallala fluvial facies. Later, when these streams had been diverted to form the Pecos and Canadian Rivers, the valleys and floodplains of these streams became the sources of eolian sediments. They concluded that the character of wind erosion and deposition during the late Tertiary was similar to that which existed throughout much of the Quaternary.

Gustavson and Winkler (1988) found that fluvial facies consist mainly of laterally extensive, vertically stacked, channel-fill sequences that range in texture from coarse clast-supported gravel to sand. Horizontal bedding predominates, but both planar and tangential cross-beds are present. Fining upward sequences in gravel- and sand-size sediments are common, and in many places bar and channel surfaces are capped by mud drapes with desiccation cracks. Armored mudballs are in most fluvial sections. Collectively, primary structures preserved in these sections indicate deposition by high-energy braided stream systems. In detail, parts of these sections are comparable to modern braided streams described by McKee and others (1967), Williams and Rust (1969), and Cant and Walker (1978).

The presence of mud drapes, desiccation cracks, and armored mudballs suggests that these fluvial sequences were deposited by intermittent or perennial streams that had wide fluctuations in water and sediment discharges. Clarendonian and Hemphillian faunal assemblages in these units (Winkler, 1985, 1987) and elsewhere in the region (Schultz, 1977) are dominated by animals adapted to subhumid or semiarid environments. Most of the ungulates in these faunas were grazers, indicating that grasslands were widespread.

Gustavson and Winkler (1988) found no record of significant fluvial deposits overlying paleo-uplands. They did note dispersed, well-rounded, siliceous gravel-size clasts in the lowermost 1.5 to 3 ft (0.5 to 1.0 m) of sediment overlying uplands. They judged these clasts to be a residuum left behind as strata older than the Ogallala were being eroded, and that they probably are the oldest fluvially transported sediments above the middle Tertiary erosional surface. Winkler (1985) thought that increase in climatic aridity suggested by increased calcification in late Ogallala time is probably related to an increase in the height of the ranges along the Rio Grande Rift, which produced a growing rain shadow.

According to Gustavson and Winkler (1988), termination of fluvial sedimentation contributing to deposition of the Ogallala resulted from diversion of the streams that drained east and southeast across Texas and New Mexico. It was hypothesized that regional subsidence resulting from dissolution of as much as 656 ft (200 m) of Permian bedded salt along the northern and western margins of the Southern High Plains led to development of the Pecos and Canadian valleys (Kelley, 1972; Gustavson and Finley, 1985; Gustavson, 1986a). Diversion of southeast-flowing streams to form the early Canadian and Pecos valleys probably ended most deposition of Ogallala fluvial facies in western Texas and eastern New Mexico. Cessation of fluvial deposition has not been precisely dated in most areas, but in the Slaton and Blanco sections in Lubbock and Crosby Counties respectively, and in the Axel locality (Randall County),

fluvial deposition ended by Hemphillian time (Schultz, 1977; Winkler, 1985). However, Gustavson and Winkler (1988) observed that the ancient Portales River, which included the upper reaches of the Pecos River above Ft. Sumner, New Mexico, and which flowed across the High Plains to the modern Brazos River, may not have been diverted until the Quaternary.

Paleoclimatology

No systematic, in-depth paleoclimatological analyses of the Ogallala deposits of the Southern High Plains have been attempted. Numerous authors, however, have commented on the general paleoclimatological conditions that possibly existed during all or part of Ogallala time. Study of terrestrial mollusks from the Ogallala of northeastern New Mexico led Leonard and Frye (1978) to conclude that the climate of the area at that time was more stable with, perhaps, slightly greater average annual rainfall and lacking the extreme hot, dry summers now characteristic of the area.

Schultz (in press) summarized paleoecological and paleoclimatological conditions suggested by the Clarendon fauna as follows:

> Sedimentological and faunal evidence suggests that the habitat of the Clarendon fauna was primarily a stream-border environment dominated by medium to large grazing mammals. The streams were occasionally subjected to high energy or flood conditions as evidenced by the coarseness of some channel deposits, the broken waterworn nature of some of the bone fragments in the channels, and the presence of thinly bedded overbank deposits. As a rule, however, quiet water flow prevailed. Adjacent to the streams were broad, grass-covered floodplains and scattered marshes, ponds, and oxbow or floodplain lakes in which fossils accumulated. Deciduous trees grew mainly along stream borders, but the habitat would best be considered a parkland savanna. As a result of salt dissolution in the

subsurface, sinkholes developed locally in the Permian red beds where the overlying sediments were not too thick; these sinkholes served as waterholes and deathtraps for the animals.

The climate of the plains during Hemphillian time apparently was mild and subtropical but with less moisture than during the Clarendonian (Schultz, 1977). Shotwell (1955, 1958) concluded that at the Coffee Ranch Quarry site a proximal Hemphillian fauna lived in a grassland habitat and was dominated by large grazing herbivores. He speculated that a distal community near the site lived in brush or open woodland and included peccaries, mastodons and wolverines.

The absence of widespread lacustrine deposits, browsing vertebrate remains, aquatic mollusks, and fossil seeds in the Ogallala, particularly south of a Lubbock-to-Clovis line, indicates that Ogallala climate was semi-arid to arid with permanent water existing only along drainage channels. Regional clay mineralogy (with general absence of kaolinite clays), lithofacies (oxidized, fine-grained, commonly eolian silts, clays and sands), and soil profiles (multiple caliche horizons) also substantiate a climate that became progressively arid throughout Ogallala time.

That the climates of the world may have been relatively arid during Ogallala depositional time is suggested by the work of Donnelly (1982). Donnelly (1982) determined from study of deep-sea cores that there has been a six-fold increase in the rate of accumulation of Al_2O_3 in north and central Atlantic and Pacific Ocean sediments indicating greatly increased denudation of the continents during the past 15 m.y. He concluded that this increase in detrital sedimentation was caused by a global change from an equable mid-Tertiary climate to a later Tertiary climate increasingly dominated by irregularities in precipitation and an incidence of catastrophic wet and dry periods with coincident decreased capability of plant cover to hold soil.

It is of interest that termination of Ogallala deposition appears to correlate closely with the NH7 (5.2 to 4.7 m.y.) deep-sea hiatus. According to Keller and Barron

(1983), oxygen isotope studies show that generally cool, but oscillating temperatures prevailed during this time interval. They also concluded that the present oceanic circulation system was established by latest Miocene time (7.0 to 5.0 m.y.). It seems likely that a worldwide change in ocean circulation would be accompanied by corresponding adjustments in the world's climatic patterns.

Suggested Studies

The broad aspects of Ogallala stratigraphy, the general nature and character of the principal depositional systems, and the geologic processes responsible for deposition of the Ogallala have been determined. Future studies need to concentrate on refinement of the stratigraphic framework and on detailed delineation and characterization of the depositional systems and lithofacies complexes present throughout the Texas High Plains. Because Ogallala deposits of this region represent a long interval of geologic time and underlie a vast area, the geologic history is undoubtedly much more complex than has been generally perceived. In reference to the Ogallala of Nebraska, Diffendal (1984, p. 21) observed that:

> The overall geologic history of the Ogallala in the southern Panhandle of Nebraska is one of multiple cycles of erosion and deposition. It is not the history of a single major episode of erosion followed by gradual valley filling that has been reported previously from elsewhere along the outcrop belt.

Blanco Formation

Introduction

Studies of the Blanco Formation generally have been designed to (1) determine the spatial and temporal distributions of the Blanco deposits on the Southern High Plains, (2) identify the geological processes responsible for deposition of the Blanco strata, (3) identify the depositional environments in which the Blanco beds originated, (4) reconstruct the paleoclimatic conditions that prevailed on the Southern High Plains during accumulation of the Blanco deposits, and (5) identify, describe, and evaluate the significance of diagenetic and pedogenetic aspects of each facies.

Blanco deposits exposed along the Blanco River (also called White River) in Dickens and Crosby Counties (fig. 30) were first studied by Cummins (1890a, 1891, 1892, 1893). Cummins (1890a) applied the name "Blanco Canyon Beds" to all the post-Cretaceous strata of the Texas High Plains. He considered these deposits to have formed in an inland sea or large lake. In 1892 Cummins changed the name of the deposits to "Blanco Beds" because he thought the double name was undesirable and because the beds also occur far from Blanco Canyon. Cummins (1890b) reported the Blanco Beds to be equivalent in age to the Green River beds when actually, as he pointed out in 1892, he meant White River beds.

Vertebrate fossils collected from the Blanco beds by Cummins in 1889 and 1890 were sent to E. D. Cope for identification. Cope (1893) thought that the fossils indicated an age intermediate between Loup Fork and the "*Equus* beds." Gidley (1903) concluded that Blanco deposits were restricted to a relatively small area in the vicinity of Mount Blanco, and that they are of Pliocene age.

The type Blanco sediments were described by Evans and Meade (1945, p. 491) as,

> . . . mainly well-bedded light gray, calcareous sands and clays with some fresh-water limestones, tufa, diatomite and coarse gravels. The finer-grained materials make up the main body of the deposits but grade marginally to coarse sand and gravel.

They also observed that the Blanco ranges from 56 to 74 ft (17 to 23 m) thick.

The type locality of the Blanco Formation is about 9 mi (14 km) north of the city of Crosbyton in Crosby County, Texas. The best exposures are west of FR 651 between Crawfish Creek and FR 193 in the J.R. Bridwell Ranch. The White River has cut approximately 220 ft (67 m) into the Llano Estacado exposing up to 89 ft (27 m) of Blanco beds. The basal beds, sands, and limestones of the Blanco are separated by an erosional disconformity from the underlying Bridwell Formation of the Ogallala Group. Except on topographic highs the Bridwell Formation is strongly reduced to a depth of 3 ft (~1 m) below the Blanco beds. The uppermost unit of the Blanco is a white to red-brown laminated caliche that is commonly contorted. This caliche is separated by an erosional disconformity from the overlying eolian cover sands.

Evans and Meade (1945) noted that the Blanco sediments are unconformably underlain by reddish-brown sands and clays of the middle Pliocene (now the Ogallala Formation) and are overlain by a widespread sheet of surface sands of eolian origin (now the Blackwater Draw Formation). They measured and described seven sections of the Blanco, of which their Section 1 is reproduced in table 4.

Pierce (1974), in his study of Blanco type deposits, included x-ray diffraction and thin-section analyses of clay, carbonate and silicate mineralogy, mechanical analysis of insoluble residue and study of pollen, diatoms, and invertebrate fossils. According to Pierce (1974), near the center of the basin the Blanco Formation is composed dominantly of very fine sand, dolostone and magnesium clays. Peripherally, limestones and coarser clastics which include caliche gravel and cobble occur in arroyo fans.

Table 4. Type section of Blanco Formation, measured on west wall of Mt. Blanco Canyon 3,500 ft (1,067 m) south of the Blanco-Cone road near the mouth of Crayfish Canyon, Crosby County, Texas (after Evans and Meade, 1945).

Unit	Description	Thickness (ft)
10	Bentonitic clay, greenish gray, sandy in upper part	10
9	Sands, gray to light greenish gray, containing calcareous nodules	4.7
8	Diatomaceous earth, light gray; varies in thickness from 1 foot to more than 5 feet; at this section	1
7	Bentonitic clay, sandy	9.4
6	Caliche, gray, jointed	6.8
5	Bentonitic clay, calcareous and sandy	6.2
4	Fresh-water limestone, thin flaggy beds, reef-like tufa masses locally present	6
3	Clay, gray, calcareous	2
2	Sand, light greenish gray, massive	12.5
1	Clay, light tan to greenish	2
	Total	60.7

Origin and Paleoclimatology

The origin of the type Blanco beds and the climatic conditions under which they were deposited have long been controversial topics. Gidley (1903) decided that the Blanco deposits are of fluvial origin rather than lacustrine as suggested by Cummins (1890a). He pointed out that they are restricted to a relatively narrow valley formed by pre-Blancan erosion. He explained the presence of diatomite as deposits in relatively short-lived ponds located on the valley floor adjacent to the stream that flowed through the valley. Baker (1915) and Mathew (1924) concurred with Gidley that the Blancan deposits were of fluvial origin.

Evans and Meade (1945) and Evans (1948) thought that the type Blanco beds were of lacustrine origin for several reasons. First, boulders and gravels derived from the Ogallala (Bridwell Formation) caprock occur at the base of the Blanco and are confined to the basin margins, whereas finer-grained sediments are in the deeper central part of the basin. This distribution pattern is the same as that of existing playas of the region and reverse that in fluvial deposits. Second, exposed Blancan sediments tend to be well-stratified and some of the strata are traceable across most of the exposed area. Third, bentonitic clays, fresh-water limestones, and localized diatomite deposits are indicative of quiet water deposition. And fourth, no evidence exists along Blanco Canyon or its tributaries of a connecting filled valley segment between the two areas of Blanco deposits or of an extension of such a valley either above or below these areas.

Evans and Meade (1945) and Evans (1948) made three points in their interpretations of the depositional history of the Blanco Formation. First, following development of the alluvial plain in middle or upper middle Pliocene, there was a period of nondeposition during which sufficiently dry climatic conditions led to development of an extensive caliche zone near the plains surface. Later, basins 60 to 70 ft

(18 to 21 m) deep were formed, which eventually contained lakes. A dry climate and an appreciable amount of time would be required to produce basins by deflation. Second, the essentially unbroken sequence of lacustrine strata composing the Blanco Formation indicates deposition in permanent or semi-permanent water bodies during a time interval sufficient to accomplish basin filling. Such permanent lakes most logically were a product of a sustained period of moist climatic conditions. And third, after Blanco basins were filled to a level approaching that of the surrounding plains, a return to relatively arid conditions is indicated by a wind-scoured erosional surface on the Blanco beds, and by deposits of loess-like dust, unaltered volcanic ash, and unstratified surface sands. Blanco beds thus appear to represent a period of relatively high humidity bracketed in time by periods of relatively high aridity.

Frye and Leonard (1957a) presented evidence to support a hypothesis that the Blanco deposits were of fluvial rather than lacustrine origin and were deposited under arid rather than humid conditions. Salient points of the argument are: (1) A fluvial origin is attested to by presence of local channel deposits, including large abraded cobbles of Ogallala caprock. (2) The generally fine texture of the Blanco sediments indicates a low gradient, shifting and anastomosing stream with very unstable bottom conditions. Such an environment, they thought, would account for the lack of aquatic mollusks, fish and amphibians, aquatic turtles, and rodents with muskrat affinities, and for the rarity of beaver remains. (3) That the climate of the area was semi-arid is evident from the lack of terrestrial mollusk, amphibian and microtine rodent remains which, if present on the plains, would have been carried into the Blanco sediments by tributary streams.

Pierce concurred with Evans and Meade (1945) that the type Blanco deposits are largely of lacustrine origin which collected in a closed basin; however, he differed with those authors in concluding that the basin contained only seasonal, or at times,

perennial water, and thus was not a permanent lake. He disagreed with the statement of Evans and Meade (1945) that the Blancan climate was humid, concluding:

> . . . the flora and fauna strongly support a semiarid to arid climate interpretation. The apparent absence of fishes, aquatic turtles and mollusks suggest that water accumulation was largely intermittent and transitory. While the basal sand and cyclic units suggest seasonal water accumulation, only very rare diatoms have been recovered from this unit. The lacustrine sands of the main bone bed zone, below the diatomite, can be interpreted as at least seasonal water on the basis of the ostracode species recovered. Candonid ostracodes are normally spring bloomers. Their occurrence suggests an early spring precipitation maximum. The flora of the diatomite unit also suggests only seasonal water accumulation, although semipermanent water may have accumulated at this time. The absence of an invertebrate fauna in the diatomite is not surprising considering the highly leached character of the unit. Above the diatomite, there is little indication of other than intermittent ponding, with sepiolite-rich immature paleosols comprising the dominant lithology.
>
> The sediment economics and mineralogy of the Blanco beds indicate deposition in a playa environment. Mineralogy and paleontology are indicative of an arid to semiarid climate.

Dalquest (1975) also concluded that the type Blancan deposits were largely of lacustrine origin but differed with Evans and Meade (1945) and Pierce (1974) in postulating that the basin was open rather than closed. He noted that some of the minor lithologic units of the Blanco are discontinuous in strong contrast to the evenly distributed sediments typical of deflation basins of West Texas. He thought that the bentonitic clays and freshwater limestones indicated still water but could have collected in low areas or in small temporary depressions after floods or heavy rains as in

arid areas today. He agreed with Pierce that the diatomite bed was a lacustrine deposit and he pointed out that the sandy mud beneath the diatomite extends for 300 ft (91 m) or more beyond the diatomite facies. He noted that aquatic plant remains are abundant and that some aquatic snails are present. The ponds in which these organisms lived were, according to Dalquest, shallow and limited to areas of a few acres. He observed that the absence of aquatic vertebrate fossils and the varied character of the Blanco sediments do not support a view that Blanco sediments collected in a closed basin occupied by a lake. It appeared to Dalquest that it is more probable that the deposits collected in a shallow drained basin, formed by stream erosion or deflation. It is noteworthy that to date leeward sand dunes have not been found associated with the Blanco basin (Pierce, 1974).

The vertebrate fauna reported by Dalquest (1975) and earlier workers is dominated by grazing herbivores and a few browsers. Dalquest concluded that,

> . . . the evidence suggests that the Blanco local fauna represents one rather uniform habitat: grassy plains with narrow belts of trees fringing water courses and in the shady parts of valleys.

North American Blancan Mammal Age

The age of the Blanco Formation of Texas and of the North American Blancan Mammal Age have long been the subjects of controversy (Dalquest, 1975; Evans and Meade, 1945; Gidley, 1903; Hibbard, 1958; Meade, 1945; Osborn, 1918; Pierce, 1974). McGrew (1948) observed that the Blancan faunas of North America represent the last group of Pliocene faunas, the first group of Pleistocene faunas, or they are transitional and overlap both. He pointed out that the original definition of the Blancan Mammal Age was given by a Geological Society of America Committee headed by Horace Wood II and that it was based on the local fauna at Mount Blanco, Texas.

The Blancan fauna is a mixture of typical Blancan genera and long-ranging genera that survived through most or all of the Pleistocene (Schultz, 1977). McGrew (1948) reported that at least 25 genera became extinct near the end of Ogallala time. Among these genera, McGrew cited the neohipparians and other horses, many antilocarprids, various carnivores and rodents, and the rhinoceroses. Tanner (1984) added to this extinction list the long-jawed mastodont (*Ambeledon fricki*) and he speculated that extinction of the rhinoceroses must have been the result of habitat restriction related to protracted drought that led to the widespread development of the caliches that cap the Ogallala Formation.

Since 1948 additional and refined studies of the Blancan faunas, coupled with evidence derived from numerical dating and petrographic analyses of volcanic ashes and from magnetic stratigraphy have combined to convince most workers that the Blancan Mammal Age is entirely of Pliocene age. It now seems probable that the Blancan began at 4.0 to 4.5 m.y. and ended at 1.7 to 1.9 m.y. (Lundelius, in press; Lundelius and others, 1987). Lundelius and others (1987) reported that the Hemphillian-Blancan boundary is not well dated, although it falls within the Gilbert reversed polarity chron probably between 4.0 and 4.4 m.y. (fig. 41). Youngest numerically dated Hemphillian faunas are the Pinhole Local Fauna (California) which is in a tuff dated at 5.2 ± 0.1 m.y. (Evernden and others, 1964; Sarna-Wojcichi, 1976), and the Santee Local Fauna in northeastern Nebraska overlain by a volcanic ash dated at 5.0 ± 0.2 m.y. by fission tracks on glass shards (Voorhies, 1977).

Despite evidence from recent studies, the Blanco Formation in Texas has not yet been precisely dated, although it is generally conceded that the Blancan Local Fauna is probably of Late Pliocene age.

In the type area of the Blanco Formation near Mount Blanco, a one-foot-thick volcanic ash bed occurs near the base of the Blackwater Draw Formation and just above the Blanco caliche. The ash has been correlated with the Guaje pumice bed of

the Jemez Mountains, New Mexico, by Izett and others (1972, 1981). The Guaje pumice bed has been dated at 1.4 m.y. based on the glass shard fission track method (Schultz, 1977, in press). Using the same technique, Boellstorff (1976) obtained a date for the Guaje pumice of 1.77 ± 0.44 m.y.

Another ash called the "Blanco Ash" lies 30 to 33 ft (9 to 10 m) below the top of the type Blanco deposit. Pierce (1974) reported that this ash, which is restricted to the northwest part of the basin, grades laterally into the prominent diatomite unit that caps some of the main vertebrate quarries in the lower part of the formation. The uranium content of this ash is too low to allow fission track dating by use of zircons (Schultz, 1977, in press), but Boelstorff (1976) used the glass shard fission-track method to obtain a date of 2.8 ± 0.3 m.y. for this ash. Chemical similarities led Izett (1981) to correlate the Blanco Ash with the 111 Ranch Ash in Graham County, Arizona, dated by the fission track zircon method at 2.3 ± 0.15 m.y. by J. J. Dickson of the U.S.G.S. This ash was also dated by the glass shard fission track method at 2.4 ± 0.38 m.y. (Dickson and Izett, 1980).

Lindsay and others (1975) found that the entire Mount Blanco section below the Guaje Ash is reversely magnetized. This led these authors to conclude that "The 1.4 m.y. ash and the absence of a normal magnetic polarity zone beneath the ash in the Mount Blanco section indicate the Mount Blanco fossils occur in the lower matuyama magnetic interval between 1.4 and 2.4 m.y. in age." Galusha and others (1984) reported that the ash at the 111 Ranch is normally polarized and therefore cannot be a correlative of the Mount Blanco Ash. In their judgment, the 111 Ranch Ash correlates with the Gauss Chron although the matuyama/Gauss boundary is placed at 2.47 m.y. (Mankinen and Dalrymple, 1979).

The evidence discussed above indicates that most of the type Blanco Formation is Late Pliocene in age and the possibility exists that the uppermost part of the formation is of Early Pleistocene age. Holliday (1988) studied the lower paleosol of the

Blackwater Draw Formation, which lies below the Guaje Ash and above the top of the type Blanco deposits at Mt. Blanco. He concluded that this soil required at least 200,000 years to form. This age estimate suggested to Holliday that type Blanco deposition ceased about 1.6 m.y. and that the formation is therefore entirely of Pliocene age. Inception of type Blanco deposition appears to closely coincide with the first major North American glaciation 2.7 to 2.4 mya (Carter and others, 1986).

Other Blanco Deposits of the Southern High Plains

The Blancan beds of Cita Canyon are exposed around the heads of small canyons which drain south into North Cita Canyon, a tributary of the Prairie Dog Town Fork of the Red River, Randall County, Texas (Schultz, 1977).

Lindsay and others (1975) determined that the Cita Canyon section contains a lower normal magnetic polarity zone and an upper reversed magnetic zone. In their opinion the Cita Canyon Blancan fauna occurs in the upper part of the Gauss epoch and therefore is somewhat older than the type Blanco deposits.

Rita Blanca beds were named by Evans and Meade (1945) for deposits that occur on the northeast side of Rita Blanca Creek, 6 mi (10 km) west-northwest of Channing, Hartley County, Texas, and 1 to 2 mi (1.6 to 3 km) north of the road from Channing to Romero. Evans and Meade (1945) provided a generalized section for this area (table 5).

According to Johnson and Savage (1955) the fossiliferous beds (Unit 1) are friable sandstones or sands laid down in lakes or playas (table 5). These beds unconformably overlie deposits of the Dockum Group (Triassic). Vertebrate fossils from Unit 1 deposits have established their age equivalence to those of the type Blanco Formation (Kirkland, 1963).

Kirkland (1963) studied a lower Rita Blanca thin (~65 in [165 cm]), laminated, fossiliferous clay bed that he considered had been deposited in a permanent lake. He

Table 5. Description of Rita Blanca beds, Blanco Formation, located on the northeast side of Rita Blanca Creek, 6 mi (10 km) northwest of Channing, Hartley County, Texas (after Evans and Meade, 1945).

Unit	Description	Thickness (ft)
3	Post-Blancan (Pleistocene) sands, clays, and volcanic ash	≤40
2	Blancan sands, bentonitic clays, and thin-bedded calcareous sandstone. This section has yielded a sparse assemblage of Blancan fauna and tracks of camels occur locally in some of the sand beds.	~50
1	Blancan clays, distinctly laminated, containing abundant plant remains and some fossil fish. Some sand beds and thin, uniform layers of fresh-water dolomite are interbedded with the shales. Locally, shales are gypsiferous.	~30

estimated the lake to have been about 6 mi (10 km) long and 3.5 mi (5.6 km) wide. Kirkland recognized two types of cyclic lithology within the clay unit: (1) alternations of carbonate-rich and carbonate-poor clay (rhythmites) several decimeters thick, and (2) alternating light and dark-colored laminae within the rhythmites. The 1,400 light and dark laminae collectively resemble modern non-glacial varves. Kirkland (1963) concluded the couplets were annual deposits based on the microstratigraphic position and cyclic changes in: (1) the main constituents: clay, calcium carbonate, silt and organic matter; (2) carbon isotopes; (3) ostracode valves; (4) leaves, predominantly *Quercus* and *Salix*; (5) adults and pupae of Chirononidae (midges); and (6) percentage composition of *Artemisia* (sagebrush) pollen and moss spores.

Kirkland (1963) concluded that the lake apparently was stratified during formation of the varves, which developed in the deeper part of the lake in a stagnant hypolimion lacking larger animal species. He observed that there were six major climatic pulsations within the 1,400 yr interval of varve formation, and that temperature and precipitation changes apparently were inverse with the climate fluctuating between warm-dry and cool-wet periods.

Blackwater Draw Formation

Blackwater Draw Formation stratigraphic studies have focused on (1) determining the spatial and temporal distributions of the unit, (2) characterizing significant depositional facies, (3) determining paleoclimatic conditions under which the formation was deposited, (4) establishing a chronostratigraphic framework on the basis of paleontological evidence, magnetopolarity, radiometric dating, and tephrochronology, and (5) determining the stratigraphic relationship of the unit to deposits of both extant and ancient playas.

Introduction

The Blackwater Draw Formation lies at or near the surface of most of the Southern High Plains, an area of approximately 31,250 mi^2 (80,000 km^2). The formation is a thin, up to 89-ft (27-m) thick sheet-like body of sediment that texturally grades from sandy in the southwest to argillaceous in the northwest (fig. 44). Samples from a drainage well near Amarillo do not contain sand coarser that 0.2 mm, whereas soil to the southwest contains abundant sand grains from 0.50 to 0.20 mm in diameter (Lotspeich and Coover, 1962). Gustavson (in press) pointed out that it is surprising, considering the extent of the Blackwater Draw Formation and its importance to regional agriculture, that so little is known of it.

Several investigations have contributed to our knowledge of the age and origin of the Blackwater Draw Formation. Judson (1953) discussed an extensive blanket of sediments that cover the Ogallala Caprock in the Southern High Plains, and which are apparently the "cover sands" of Frye and Leonard (1957). From a study of outcrops and well logs, Judson determined that the "cover" was up to 40 ft (12 m) thick, to be almost universally present throughout the Southern High Plains, and to consist of a highly calichified sand with a high percentage of silt- and clay-size particles. He concluded that the deposit was of eolian origin, that it is of post-Ogallala Caprock age, and that it had a long and complex history involving accumulation during periods of aridity and of weathering during moist intervals.

Frye and Leonard (1957b) referred to Blackwater Draw sediments as "cover sands," and included with them the Judkins Formation, an eolian sand identified by Huffington and Albritton (1941), adjacent to the southwest and west part of the Texas High Plains. They reasoned that the cover sands were in part eolian, in part sheetwash, and possibly in part basin fill (Frye and Leonard, 1967). They hypothesized that the cover sands were derived from major valleys such as the Pecos to the southwest and west. A well-developed soil, supposedly restricted to the upper

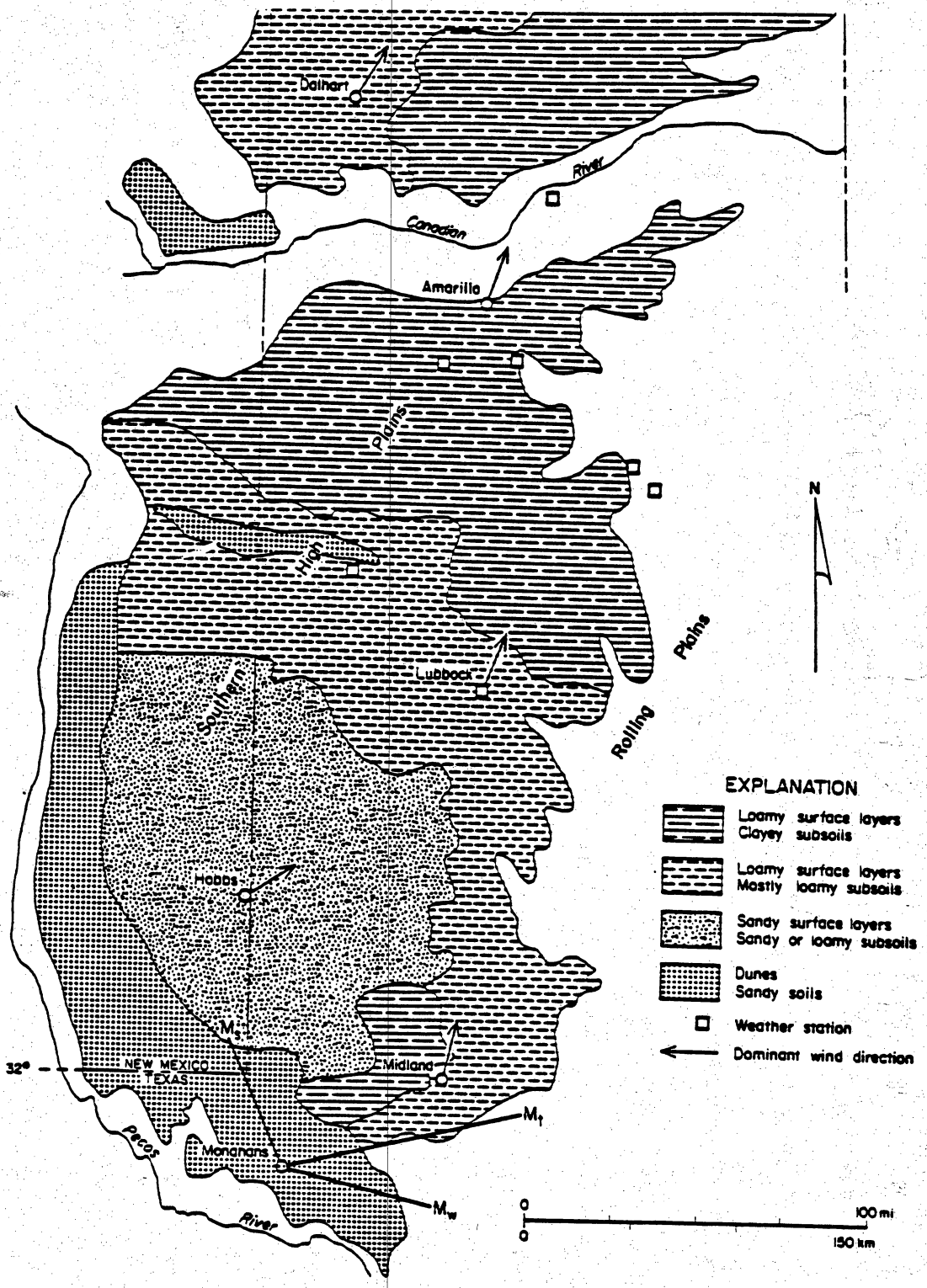


Figure 44. Distribution of the Blackwater Draw Formation showing fining of surface soils to the northeast. Coarsest soils occur east and north of the Pecos and Canadian Rivers suggesting that these may be the source areas for at least the upper part of the Blackwater Draw Formation. Modern dominant wind directions are also shown. Mean summer (Ms), mean winter (Mw), and mean total (Mt) winds are shown for Monahans, Texas. After Gustavson (in press).

part of the cover sands was correlated by Frye and Leonard (1964, 1967) with the "Sangamon Soil" of Illinois. These authors (1957b) concluded that the cover sands were of Illinoisan age because they thought the sands lay stratigraphically above Kansan lacustrine beds (Tule Formation), and below Wisconsinan lacustrine beds (Tahoka Formation?). They also noted that the sparse molluscan assemblages of the cover sands can be distinguished from those of the Kansan by the absence of species restricted to the Kansan but that cover sands assemblages are indistinguishable from those of the Early Wisconsinan.

Reeves (1976) accepted an Illinoisan age for the "cover sands" of Frye and Leonard (1957b) and proposed that they be formally termed the Blackwater Draw Formation. He designated a type locality located on the west side of Blackwater Draw west of New Deal at approximately lat 33°46' N., long 101°52'30" W. and a reference section located north of the town of Ralls on the Floydada road (Texas Highway 207) about 100 yards west of the present highway on the south side of White River Canyon in southern Floyd County (lat 33°53' N., long 101°22'30" W.) (see fig. 30). He stated that the formation is of eolian origin, and is chiefly fine sand to sandy clay loam. He noted that the unit increases in thickness from a featheredge in southwestern New Mexico northward to at least 88 ft (27 m) and in the northern High Plains it is overlain unconformably by the "Peoria Loess." He also observed that the "Sangamon Soil" of the Southern High Plains is the product of multiple periods of pedogenesis and that it therefore is an unreliable stratigraphic marker. Seittheko (1975) studied the textural variation of the surface soil of the Blackwater Draw deposits and concluded that it fined downwind from southwest to northeast, supporting the hypothesis that the sediments originated in the Pecos valley (fig. 44). Allen and Goss (1974) and Hawley and others (1976) recognized that in some places the Blackwater Draw contains up to seven well-developed soils including that at the surface. This indicated to Gustavson and Holliday (1985) that the unit was deposited episodically.

Origin of Blackwater Draw Deposits

Gustavson and Holliday (1985) proposed a model for the Quaternary geologic history of the Southern High Plains that they considered also to be applicable to the Rolling Plains. These authors concluded that the Quaternary geologic history is cyclic and postulated that each cycle began with eolian deposition, followed by non-deposition accompanied by landscape stability and soil formation, and ending with erosion. Although they considered wind to be the dominant geomorphic and sedimentologic agent during the Quaternary, they noted that during any given interval wind erosion and sedimentation probably operated together along with pedogenic processes as is true today. They observed that during a single cycle, different processes would dominate from time to time. Three principal phases of the cycle were postulated. (1) During deposition, material derived from the Pecos valley would be blown onto the High Plains surface and across the Rolling Plains at a rate greater than erosion or pedogenesis. The result would be deposition of a more-or-less continuous sheet of sediment across the area. (2) With subsequent slowing of sedimentation and minimal erosion, landscape stability would be established with soil formation in the eolian sheet. During this phase lacustrine sediments would accumulate in surface depressions and with sufficient time a well-developed soil would form similar to the present surface soil of the region. (3) Erosion by wind deflation would then destroy the lateral continuity of the eolian sheet in many areas, and in some places the more easily eroded A horizon would be removed down to the Bt horizon. Erosion would occur immediately before or possibly simultaneously with early stages of the subsequent depositional event. Deflation in a region could result in deposition in downwind areas.

Gustavson and Holliday (1985) concluded that several such cycles would produce vertically stacked disconformable sedimentary layers varying in number from one section to another but with each unit exhibiting similar pedologic properties. They

noted that at the present time the Southern High Plains is well into the cyclic phase dominated by landscape stability and soil formation following at least six complete cycles.

Although lakes may have been more abundant during phase 2 of the Gustavson-Holliday model, it seems probable that they were present on the Southern High Plains during all phases of the cycle. Eolian and lacustrine facies must have developed at varying rates continuously during the Late Pliocene and Quaternary. Lake basin distribution and morphology, lake water chemistry and biology, and the effectiveness of wind to erode and transport sediments would have vacillated in concert with Late Cenozoic climatic fluctuations. Locally and regionally, relatively thin sheets of eolian clastics and small lacustrine deposits would have been buried beneath accumulating sediment, altered by pedogenic processes, or partially or completely removed by erosion. The resultant post-Ogallala sediment cover would consist of laterally restricted lenses of eolian and lacustrine facies suites bounded vertically by buried paleosols or erosional unconformities. Probably each local stratigraphic section preserves only a part of the post-Ogallala depositional record. A stratigraphic composite constructed from a series of strategically located outcrop sections and drill hole core, preferably including fossiliferous lacustrine deposits, should provide a reasonably complete record of the post-Ogallala history of the Southern High Plains.

Type Section

Holliday (in press) measured and described the type section of the Blackwater Draw Formation selected originally by Reeves (1976). Holliday noted two prominent features of the sediments of the type section: their general homogeneity, and the presence of a number of strongly developed buried soils. He identified at least four buried soils in addition to the modern surface soil (fig. 45). The entire section appears essentially homogeneous because the soils have about the same degree of

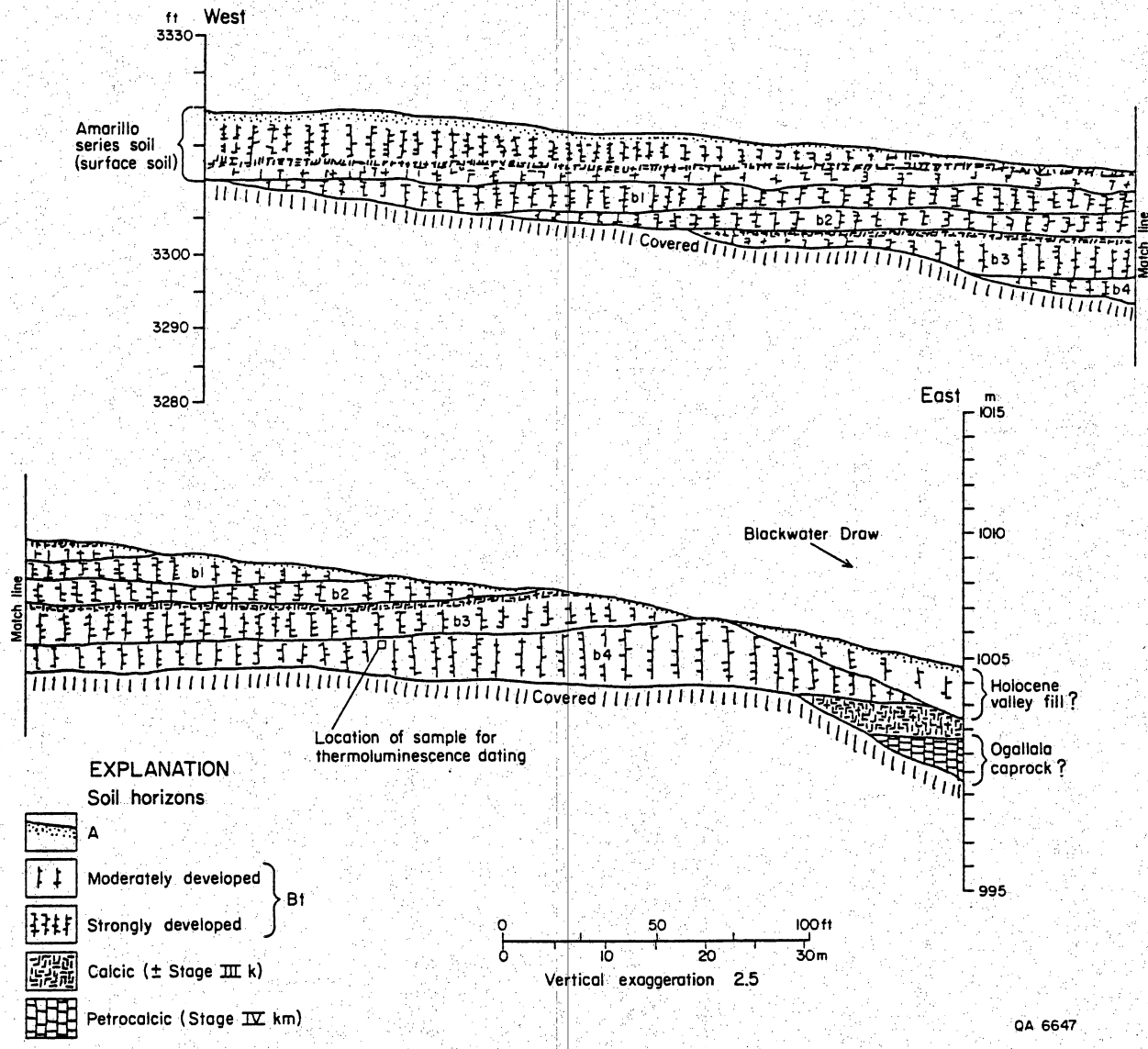


Figure 45. Generalized soil-stratigraphic relationships exposed at the Blackwater Draw Formation type section, Lubbock County. After Gustavson and Holliday (1985).

pedogenetic expression. Holliday described them as reddish-brown (2.5YR - 5YR hues), moderate to strong prismatic structure with common films of illuvial clay in ped surfaces, and stage II to III calcic horizons (after Gile and others, 1966). No A or C horizons, other than those of the surface soil, were recognized by Holliday.

Holliday (in press) described the soils of the type locality for the Blackwater Draw Formation as follows:

> . . . the surface soil has a thick, reddish-brown (5YR hues), argillic horizon (50 to 149 cm) and dominantly coarse prismatic structure. Clay films are apparent on ped faces and very common in thin section. Below the argillic horizon is a Stage III calcic horizon (149 to 290 cm). The boundary between the argillic and calcic horizons is very abrupt and subhorizontal, similar to a depositional contact. Such an abrupt boundary is typical of the regional surface soils, however, and is considered to be pedogenic. All characteristics of this soil indicate that it is very well developed.
>
> Morphologically, the buried soils in the section are more strongly developed than the surface soil. At least some portions of each profile exhibit 2.5YR hues, more continuous and thicker clay films on ped faces, a higher percentage of clay films in thin section, and angular blocky structure. The blocky structure is apparently due to higher illuvial clay content, suggested by the abundant clay films.
>
> The morphologies of the calcic horizons in the buried soils are somewhat similar to one another, but distinctly different from that in the surface soil. The buried calcic horizons are expressed as either patchy films and coats on ped faces, generally the vertical faces, or as vertically oriented, root-like nodules about 1 to 3 cm thick. These nodules appear to follow the joints between prismatic peds and this general morphology is sometimes informally referred to as "ladder structure" (McGrath, 1984, p. 131). Some

consider ladder-matrix carbonate horizons to represent recalcified argillic horizons (for example, McGrath, 1984). There is some evidence that these horizons represent formerly more extensive calcic horizons that underwent dissolution. The upper boundary of the ladder-matrix calcic horizons are commonly horizontal and abrupt. This suggests that the horizons were originally similar to the surface-soil calcic horizon but were subsequently subjected to dissolution with concentration of the carbonate along veins (ped faces?) and as nodules. Illuvial clay is very common in and on peds between carbonate nodules in the ladder-matrix horizons, but such clay is not apparent in the surface soil calcic horizon. This would further suggest that clay illuviation accompanied or followed carbonate dissolution.

Two K horizons are exposed at the bottom of the section. The K1 horizon is a massive, nonindurated calcic horizon, probably representing a buildup of carbonate owing to the impermeable nature of the underlying calcrete. The K2m is a silicified calcrete, probably the "caprock caliche" at the top of the Ogallala Formation.

There is one exception to the strong pedologic expression in the type section of the Blackwater Draw Formation. At the east end of the roadcut on the floor of the Blackwater Draw, the surface soil is weakly developed, exhibiting an A-Bw-Btk profile, and some clay films are apparent in thin section. Although the hues in this soil are 7.5YR to 5 YR, the soil is not considered as well developed as the others in the section because there are no clay films on ped faces and the structure is weak, subangular blocky. Based on stratigraphic position and pedologic expression, this deposit is considered to be late Holocene valley fill commonly found in the draws in the region (Holliday, 1985a, 1985b).

Gustavson and Holliday (1985) suggested that all cycles were of approximately the same duration, and speculated that each cycle lasted on the order of 100,000 years. If this estimate is accurate it is possible that the Blackwater Draw cycles correlate with the Milankovitch 100,000-year eccentricity cycle (Kerr, 1981).

Age

Recognition of the occurrence of multiple buried soils in the Blackwater Draw Formation led Holliday to conclude that deposition of the unit was episodic and that it has a much greater age range than previously thought (Holliday, 1984, 1985b, in press; Machenberg and others, 1985). The buried soils closely resemble the well developed surface soils of the central Southern High Plains. To derive an estimate of the age of the upper part of the Blackwater Draw Formation, Holliday (1984, 1985b, in press) compared pedologic data from the unit with rates of Holocene pedogenesis; some samples were dated using thermo-luminescence (TL) techniques. Pedologic development of surface soil at the type section was compared with late Holocene soils at the Lubbock Lake archeological site (Holliday, in press). Comparisons were made using B-C soil types, $CaCO_3$ content expressed as gm/cm^3, calcic horizon thickness, and characteristics of clay-film coatings determined in thin section. These calculations were plotted along regression lines by Holliday (1982) for the late Holocene soils. The results suggest that the surface soil has been forming for a minimum of 10,000 years.

Five TL dates for the Blackwater Draw Formation are ambiguous. Approximate ages for two samples associated with a dune located 1.2 mi (2 km) north of Lubbock Lake site (Lubbock County) are known. Organic material from the A horizon of the lower buried soil of the dune provided a radiocarbon date of 37,750 ± 3,600 yr B.P.; however, a sample from the underlying Blackwater Draw yielded a date of only 28,600 ± 4,000 yr B.P. Holliday (1985a) concluded from the radiocarbon dates and

pedologic development of the soils in the dune that the Blackwater Draw Formation was buried by the dune about 40,000 to 50,000 yr B.P.

Holliday (in press) studied two sections on the Southern High Plains in addition to the type section. At the type section of the Blanco Formation in Crosby County, Texas, 16 ft (5 m) of the Blackwater Draw Formation rests unconformably on the lacustrine deposits and pedogenic calcrete of the Blanco Formation. The 1.4 m.y. Guaje ash derived from the Jemez Mountain volcanics in New Mexico (Izett and others, 1972) lies within the Blackwater Draw about 13 ft (4 m) below the surface of the formation. The Blackwater Draw at this locality includes three buried soils plus the surface soil. The lowermost soil is beneath the ash which lies about 3 ft (1 m) above the top of the Blanco.

Deposits of the Blackwater Draw and Tule Formations occur interbedded in a section exposed in Swisher County 12 mi (19 km) east-southeast of Tulia in Tule Canyon. Here lowermost Blackwater Draw eolian sediments sandwiched between Tule lacustrine facies include a layer of 620,000 yr Lava Creek (formerly Pearlette O) volcanic ash (Izett and Wilcox, 1982). The Blackwater Draw sediments beneath the ash include a well-developed soil with a Bt horizon, 7.5 YR hues, moderate structural development, and a thin Stage III calcic horizon. Blackwater Draw eolian sediments above the highest bed of Tule deposits also appear to contain a buried soil, or a formerly deeper surface profile (Holliday, in press).

The lowermost five buried soils of the Blackwater Draw Formation indicate that the remanent magnetism, apparently acquired during pedogenesis, is dominantly reversed. This evidence points to formation of the soils during the m.y.tuyama reversed polarity epoch which ended about 0.79 m.y. (Holliday, 1984, 1985a, in press).

Suggested Studies

Future study of the Blackwater Draw should focus on four aspects: (1) origin and character of the Blackwater Draw Formation and its regional correlation, (2) physical

properties and chemical composition of individual soil horizons to establish stratigraphic marker zones useful for regional correlation, (3) chronological calibration of soil-forming processes responsible for development of Blackwater Draw soil profiles, and (4) relationships between Late Pliocene-to-Pleistocene lacustrine facies and the adjacent Blackwater Draw deposits. Where it can be shown that Blackwater Draw sediments interfinger with dated or datable lacustrine deposits the chronostratigraphy of a segment of the Blackwater Draw section can be established. A series of such studies involving lacustrine facies that collectively span Late Pliocene to Pleistocene time, combined with magnetopolarity data for the Blackwater Draw Formation, would provide a chronostratigraphic framework for the entire Blackwater Draw section.

Surface Processes

Extant Playas

Introduction

The term playa is used in southwestern United States for vegetation-free, flat areas at the lower part of an undrained desert basin. They are underlain by silt or sand, and commonly by soluble salts. The term is also applied to the basin containing an expanse of playa, which may be marked by ephemeral lakes (Bates and Jackson, 1980).

Various studies have contributed significant information and insights to characterize playas and playa basins in the Southern High plains, determine their origin, rates of erosion, deposition, and subsidence, and to determine the role of playas in recharging the High Plains aquifer and the potential role of playas in the introduction of contaminants to the aquifer. Various geologic processes and controls have been considered to effect formation of playa basins, including: (1) tectonic processes of downwarping and faulting; (2) differential compaction and subsidence within the Ogallala Formation; (3) deflation; (4) alluvial damming by fan construction; (5) landslides; (6) removal and compaction of sediments by animals; (7) localized leaching of caliche beds; (8) breaching along Pleistocene pluvial streams; (9) original undulations (troughs) in Ogallala caprock; (10) meteorite impacts; (11) incision of caprock and subsequent deflation of leached sediments located in dunefield swales; (12) caliche karst; (13) volcanic craters and calderas; and (14) deep-seated solution followed by collapse of the surface.

The surface of the High Plains of Texas is dotted by more than 30,000 small playas. These basins are generally less than 1.0 mi (1.6 km) in diameter, 25 to 90 ft (8 to 27 m) deep, circular to somewhat elliptical, and intermittently contain ponded

water. These tiny, ephemeral lakes, which may be the principal source of recharge to the High Plains Aquifer, also have been the collection sites during the Holocene and possibly late Wisconsinan of thin (5 to 40 ft [1.5 to 12.0 m]) roughly saucer-shaped lenses of dark gray and brownish-gray clays and loams. The primary sources of clastic sediments are the basin margins and adjacent upland divides; however, some eolian dust to fine-sand size particles and volcanic ash and dust probably have been contributed from distant sources. Soils maps show in each of the larger depressions several soil types commonly arranged in concentric bands on the basin floor and slopes. Some large and elongate or composite basins display a complex mosaic of soil patterns. Some playas occur over or nested within playa fills of older and generally larger depressions indicating genetic cyclicity of playa development.

The central and lowest part of most of these small lake basins contain thin lenses of organic-rich silty clay mapped by soils scientists as the "Randall Clay." The Randall Clay occurs in some playas associated with a highly calcareous, silty, sandy clay referred to as the "Arch Soil." Some geologists have referred informally and incorrectly to these basin deposits as the "Randall Formation." The interchangeable use of the same formal term for both a soil and a lithostratigraphic unit is not only confusing, but violates standard stratigraphic procedure (art. 22, remarks k, p. 856, North American Stratigraphic Code, Commission on Stratigraphic Nomenclature, 1983). Thus the playa deposits have no formal lithostratigraphic name.

Origin and Development of Extant Playas

Theories regarding the origin and development of playas on the Southern High Plains were reviewed by Osterkamp and Wood (1987) and DuBar (1985). Numerous hypotheses have been proposed. The most plausible concepts involve deflation

(Holliday, 1985b), subsidence, or a combination of deflation and subsidence. Subsidence is either deep-seated, resulting from dissolution of Permian evaporites at depth (Gustavson and others, 1980b; Reeves and Temple, 1986), or shallow-seated, resulting from dissolution of Tertiary (Ogallala Formation) or Cretaceous carbonates (Osterkamp and Wood, 1987; Wood and Osterkamp, 1987). Each of these hypotheses has its proponents and there is some field evidence to support each hypothesis. Caran and Baumgardner (in press), on the basis of stratigraphic studies, showed that existing regional depositional and erosional chronologies are poorly constrained and commonly misunderstood. In addition, geohydrologic (recharge and discharge) and geomorphic processes are not adequately explained by existing hypotheses (see Dutton, this volume). Osterkamp and Wood (1987) and Wood and Osterkamp (1987) provide partly quantitative data regarding shallow-seated subsidence, and Gustavson and others (1980a) estimate gross volumetric relations of evaporite dissolution that may be used to crudely estimate subsidence rates. More accurate approximations of both shallow and deep-seated subsidence are needed. Preliminary analyses of deflation (Machenberg and Caran, 1987) and dustfall (Machenberg, 1986) are based on short periods of modern record and where possible should be supplemented by paleoenvironmental interpretations.

Since Gilbert's study of Arkansas playas (1895), deflation has been considered by most geologists concerned with the Southern High Plains to have played a dominant role in development of playa basins (Evans and Meade, 1945; Judson, 1953; Reeves, 1965; and Reeves and Parry, 1969). Initiation of deflation basins generally has been strongly correlated with periods of climatic aridity when vegetative cover was weakened, or locally breached (for example, buffalo wallow; gullying).

Reeves and Parry (1969) contended that the small lake basins were constructed by wind deflation during two periods of lake desiccation. Larger basins with adjacent lee dunes are thought to have formed about 17,000 years ago during Vigo Park time

when prevailing winds supposedly blew S.30°E. to S.80°E. Smaller depressions generally devoid of lee dunes are attributed to deflation during the Middle Altithermal event (about 6,500 to 5,000 years ago) when prevailing winds supposedly were N.20°E. Lotspeich and others (1971) called attention to the apparent role of "country rock" character as an influence on depression size and morphology. According to Lotspeich and others, playas that developed where the mantle is a silty clay loam tend to be larger but less densely spaced than those developed on predominantly sandy soils. Further, Lotspeich and others (1971) noted that lee dunes occur only in association with playas on the sandier soils. Because soils of the Texas High Plains fine toward the northeast, lee dunes occur only rarely north of Randall and Deaf Smith Counties.

The sediments of these dune deposits generally are reworked lacustrine fill containing shells of aquatic fossil snails, rather than eolian sediments of the Blackwater Draw Formation that underlie and surround the playa deposits.

Wood and Osterkamp (1984, 1987) and Osterkamp and Wood (1987) proposed that playa basins of the Southern High Plains developed by a combination of dissolution of caliche and piping of surface material into the unsaturated zone rather than by eolian processes. They assume that playa basins originate wherever surface depressions collect and store water. Osterkamp and Wood (1987) thought that some of the depressions formed by wind deflation, that others formed over inferred bedrock fractures and in existing and abandoned stream channels, and that a few may have been initiated by animal activities or by meteorite impacts.

Osterkamp and Wood (1987) observed that most playas are nearly circular and concluded that this uniformity in shape suggests that playas tend to develop outward from a central position. Five observations support this hypothesis. (1) Playa bottom permeabilities are greater beneath a ring surrounding the basin floor than beneath the basin floor where infiltration is impeded by clay-rich lacustrine sediments (Cronin, 1964; Havens, 1966). (2) Rates of water-level decline decrease following subaerial

exposure of the annular areas (White and others, 1946). (3) Samples from a bore hole about 120 m beyond a playa-floor edge were uniformly dry, whereas samples from a 15 m deep hole near the playa periphery were all nearly saturated (Lotspeich and others, 1971). (4) Thickness of relatively impermeable playa deposits decreases from playa centers, but carbonate contents of underlying beds generally increase toward the periphery of the depressions presumably because of dissolution. (5) Small playas appear to be least mature showing little deposition of organic material and no evidence of geochemical alteration of underlying beds, whereas larger playa floors, up to 0.6 mi (1 km) or more in diameter, may contain 16 to 33 ft (5 to 10 m) of playa deposits and show extensive geochemical changes.

According to Osterkamp and Wood (1987), playa development and enlargement occurs by carbonate dissolution by percolating water in the unsaturated zone, downward piping of water and fine clastic material, and eluviation by ground water of dissolved and particulate matter, which may include plastic movement of silt and expanding clays along microfractures during wetting and drying episodes. These authors agreed with Reeves (1966) and Carlisle and Marrs (1982) that most of the "proto" basins are of eolian origin but differ in concluding that most of the subsequent basin development was hydrologic rather than eolian.

The view that playa basin origin is at least partially a product of solution of Ogallala carbonates by recharging waters has support derived from the studies of Finch and Wright (1970) and Finley and Gustavson (1981). These authors suggested that development of observed physiographic lineaments on the surface of the Southern High Plains is strongly influenced by occurrence of regional joint-sets and minor faulting. Finley and Gustavson (1981) concluded that joint intersections provide potential sites for downward percolation of water, which might enhance playa development as is suggested by dissolution of caliche beneath playas.

Additional evidence of carbonate dissolution directly beneath ephemeral playa deposits of the Southern High Plains is cited by White and others (1946), Havens (1966), Reeves (1971) and Lotspeich and others (1971). The presence of solution openings and pipes in Ogallala caprock exposures have been noted by Nye (1930), Evans and Meade (1945), White and others (1946), and Reeves (1976).

Caran (1987) studied core from three aligned bore holes drilled in the northeastern quadrant of a playa located one mile (1.6 km) northeast of Nazareth, Castro County, Texas. At this site the modern playa is nested in the calcareous, silty sands of sediments apparently of an older, and larger playa. The older lacustrine deposits are possible correlatives of the nearby type Tule Formation. Caran noted that Ogallala Caprock, apparently present at the periphery of the modern playa, is absent beneath its center. He also observed that sediments of the modern playa appear to interfinger with sandy silts of the Blackwater Draw Formation.

Wood and Osterkamp (1987) concluded that their mass-balance studies support the hypothesis, based on hydrologic, geologic and geomorphic data, that the small playas of the Southern High Plains developed by a combination of dissolution of caliche and piping of surface material into the unsaturated zone. Based on their combined studies, Wood and Osterkamp (1987) proposed the following model for playa development. (1) Small depressions, most of which are probably of eolian origin, concentrate surface run off and increase recharge flux through them relative to the area between basins. (2) Clay and colloidal-size organic particulate and organic matter sorped on clays, along with fine-grained inorganic particles are transported into these protobasins and enter the unsaturated zone suspended in recharging water. (3) Organic material introduced with recharging water is oxidized to CO_2 by resident aerobic microbial communities. (4) Some CO_2 escapes to the atmosphere and some forms carbonic acid with the recharging water. (5) Carbonic acid in turn dissolves carbonate and silicate minerals in the unsaturated zone. (6) Instability caused by this

dissolution, the presence of fine smectite clays, an unsaturated zone below the depression, and large, localized ground-water recharge flux, promote particulate transport by micropiping and eluviation of both the lithologic framework and material introduced from the surface. (7) Increases in surficial catchment area are caused by enlargement due to dissolution and particulate transport from the basin, enhance the ground-water flux through the basin floor and induce additional piping. (8) During early stages of basin development, the amount of sediment removed from the basins by dissolution and micropiping is greater than that introduced and the basins are deepened. (9) Deepening permits enlargement of the catchment area as the bounding slopes retreat in response to sediment washed into the developing basin. (10) Some introduced material accumulates on the basin floor, ultimately reducing the hydraulic conductivity and thus the recharge flux. (11) Reduction of recharge in the depression center in turn reduces the amount of dissolution and piping beneath the basin floor. (12) Water unable to recharge in the depression center is recharged at the edges of the basin floor where hydraulic conductivity is greater. (13) Peripheral recharge promotes increased dissolution and piping in these areas and causes horizontal expansion of the basin. Filling and enlargement occur simultaneously.

Suggested Studies

Future playa studies are needed to identify the range of genetic factors (subsidence, deflation, and biogenic transport of sediment) that affect playas and their relative importance, based on geomorphic and stratigraphic data. Completed studies will supplement previous investigations of playas and refine interpretations of rates of processes by which playas formed and are maintained. These interpretations will be applicable to hydrologic and geomorphic models of playa origin and variability in development rates and in recharge rates during the Late Cenozoic.

General techniques for playa investigations are discussed by Bradley (1985) and Berglund (1986). Palynological analysis of samples should be attempted although preservation of pollen and spores in this region appears to be generally poor (Bryant and Holloway, 1985; Hall, 1985). Nevertheless, careful sampling of low-permeability argillaceous lacustrine sediments may yield materials suitable for study. Fossil mollusks (Neck, in press) and ostracodes (Forester, 1985) can provide a very reliable record of paleoenvironmental conditions, and specimens of both groups occur abundantly in many Late Cenozoic deposits of the area. A series of recent papers edited by Porter (1983) discusses a variety of other, well-established indicators of paleoenvironmental conditions relevant to the Southern High Plains and adjacent areas; in particular, papers by Friedman (1983), Smith and Street-Perrott (1983), and Spaulding and others (1983) treat isotopic, paleofaunal, limnologic and paleofloral evidence, respectively.

Ground-Water Lakes

Introduction

The term "ground-water lakes" is applied here to lakes that intersect the water table of near-surface unconfined aquifers. These so-called ground-water lakes are distinguished from playa lakes, which typically are perched at a considerable height above the water table, above an unsaturated zone.

This section reviews the origin and development of Neogene and Quaternary lacustrine systems and provides a history of lake basin subsidence.

The Southern High Plains is a low-relief, mostly internally drained plateau encompassing more than 30,000 extant ephemeral lakes or playas (Schwieson, 1965; Caran, 1987; Caran and DuBar, in press). Playas are not treated in the present discussion. This discussion focuses on saline lakes (sometimes called playas) of the area

and the perennial Bottomless Lakes in the Pecos River valley of New Mexico. Also described are Upper Cenozoic lacustrine deposits, particularly the Pleistocene Tule Formation, which occupy discrete basins throughout the region.

Beneath the Southern High Plains, post-Permian strata overlie more than a kilometer of Permian evaporites. Most of the Mesozoic and Cenozoic cover has been stripped from a zone along the margins of the Southern High Plains, leaving updip evaporites relatively near surface and susceptible to rapid dissolution. In this zone, active dissolution subsidence, including rare catastrophic events, is a significant environmental hazard (Gustavson and others, 1980a, 1982; Baumgardner and others, 1982). Lakes formed in areas of localized subsidence during periods when the water table was high. Caran and Baumgardner (1986; in press) described syndepositional deformation of Pleistocene lake beds at a site in the dissolution zone east of the Plains. However, circumstances governing the onset and progress of subsidence are poorly understood. To date, this process has not been adequately investigated.

There is also indirect evidence of dissolution below interior sections of the Plains, where evaporites are downwarped and covered by thick overburden. Permian salt is abnormally thin beneath the lacustrine Tule Formation in its type area (Gustavson, 1986b), and strata underlying the Tule beds are downwarped locally (Evans and Meade, 1945). Data from Evans (1946), Johnson (1988), and Schultz (in press) indicate karstic subsidence at other Miocene to Middle Pleistocene lacustrine basins in the Texas Panhandle. Possible evidence of subsidence beneath playas on the Plains has been summarized and evaluated by Caran and DuBar (in press).

Some modern lake basins in the area contain commercial deposits of industrial minerals including mirabilite ($Na_2SO_4 \cdot 10H_2O$) (Wood and Jones, in press). Analogous lakes in Mexico (Caran and Winsborough, 1986) and Cenozoic lacustrine deposits at sites in the United States, for example, the Piceance Creek Basin of Colorado (Dyni, 1974), also contain important evaporite minerals and/or hydrocarbons. For these reasons, genesis of ground-water lakes is an important area of research.

Neogene to Pleistocene Lakes

Lacustrine deposits form all or part of the Ogallala, Rita Blanca, Cita, Blanco, and Tule Formations, and several unnamed stratigraphic units (Caran, in press b). The biostratigraphy, magnetostratigraphy, and tephrochronology of the Pleistocene Tule, in particular, are well understood. The Tule Formation of Briscoe and Swisher Counties, Texas, contains diagnostic paleofaunas and is bracketed by Lava Creek B volcanic ash (620,000 yr) near the top of the formation and Cerro Toldeo-X ash (about 1.2 m.y.) at its base (Kumanchan, 1972; Izett and Wilcox, 1982; Schultz, in press) (fig. 46). Deposition of the Tule spanned the paleomagnetic reversal at the Bruhnes-Matuyama chron boundary (about 750,000 yr), which should be recorded in the remanent magnetic profile. Lacustrine sequences generally afford an extremely reliable geomagnetic chronology, and Linsday and others (1975) successfully analyzed the magnetic record of Miocene lacustrine deposits in this region. These chronologic indicators should be used to constrain the origin and development of the Tule and its basin. Hydrologic conditions required to sustain perennial lakes in the region during the Late Cenozoic are unknown and should be investigated by means of sedimentologic and paleoecologic indicators of water depth (and, therefore, paleo-water table).

Modern Lakes

The Bottomless Lakes of Chaves County, New Mexico, are "cenotes," small perennial lakes within deep, steep-walled sinkholes that extend below water table (Caran and others, 1986; Caran, in press a). Underflow of artesian ground water in the San Andres Formation dissolves Permian halite and gypsum less than approximately 30 ft (100 m) beneath the Pecos River valley, causing subsidence and upward

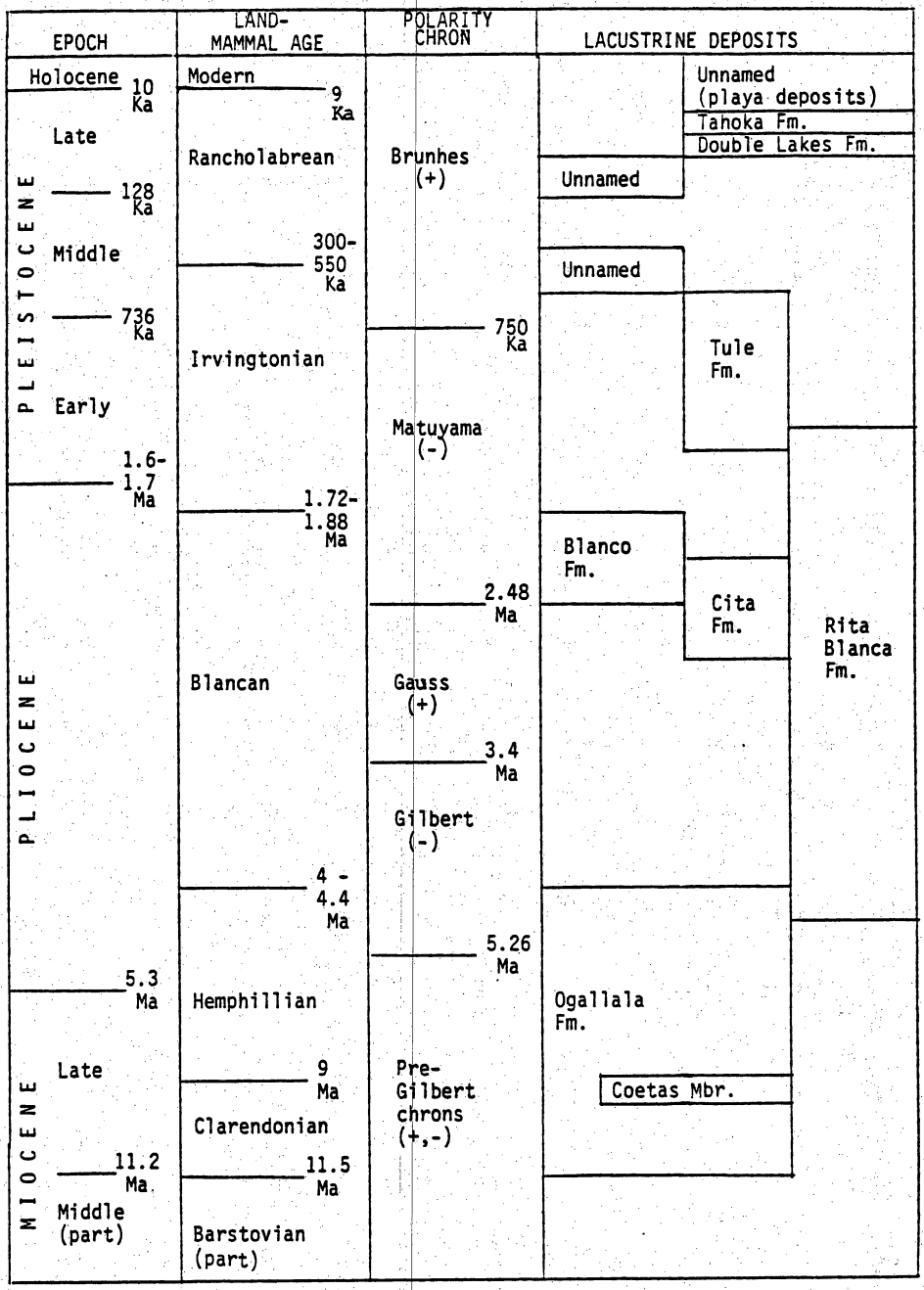


Figure 46. Lacustrine deposits of the Southern High Plains, Texas Panhandle. Modified from Caran (in press b).

propagation of collapse chimneys. Lakes form when these structures reach the surface. Baumgardner and others (1982) described the catastrophic development of "Wink Sink," a cenote formed in 1980 southwest of the Southern High Plains.

The Bottomless Lakes are noteworthy because: (1) their surface-drainage areas are so small that inflows are negligible and (2) calculated mean annual lake evaporation exceeds mean precipitation in this semiarid climate by a factor of seven or more (National Climatic Center, 1977; Farnsworth and others, 1982). Coarse-clastic sediment is comparatively rare and deltas are small or absent. Sediments in Bottomless Lakes are laminated fine-grained clastics and biogenic (mostly algal carbonate) carbonates. Diatomites and evaporites are common in other pre-Late Pleistocene basins and in modern saline lakes. Many of these deposits are fractured and folded by subsidence.

There are approximately 40 saline lakes on the Southern High Plains (Reeves and Temple, 1986). Most are perennial and saline, in contrast to the more numerous playas which are ephemeral and fresh. Dissolution of Permian evaporites at depths of about 152 ft (500 m) produced subsidence basins coincident with several existing saline lakes (Gustavson and others, 1980; Reeves and Temple, 1986; and Reeves, in press). These lakes and most pre-Holocene lacustrine basins lie within the interior of the Southern High Plains. Garcia Lake in Deaf Smith County, Texas, is an attractive candidate for detailed investigations because both the basinal fill and underlying Triassic Dockum Group are well exposed (Eifler and Reeves, 1977).

Suggested Studies

Preliminary studies have shown that modern lakes of the Southern High Plains and adjacent areas are valid analogs of Cenozoic lakes throughout the region. Origin and development of many of the modern basins appear related to dissolution subsidence and emergence of ground water. Investigation of the Pleistocene Tule Formation and modern deposits in Garcia and Bottomless Lakes would greatly improve

understanding of subsidence and origin of ground-water lake systems. Study of these lakes and lacustrine deposits would include detailed geologic mapping and description and sampling of exposed sections and cores to: (1) construct paleoenvironmental interpretations from fossil mollusk, diatom, ostracode, charophyte, pollen, spore, and any recovered vertebrate remains; (2) determine depositional history from texture, fabric, and mineralogy of samples, especially clays, evaporites, and volcanic ash; (3) evaluate strain history from distribution and orientation of fractures in lacustrine deposits and bedrock; and (4) estimate process rates from radiocarbon dating (where appropriate) and magnetostratigraphy (minimum five-step demagnetization). These data are needed from combined investigations of modern and ancient lake deposits to reconstruct the processes and chronology of basin development and infilling.

Paleoclimatology

Several studies have attempted to characterize the range of paleoclimatic conditions that have existed in the Southern High Plains during at least the past 2.0 Ma. Results of these studies serve as a partial basis for long-term assessment of future climatic conditions during the next 10,000 to 100,000 years. However, systematic, detailed, multidisciplinary reconstruction of Late Cenozoic paleoclimates of the Southern High Plains remains to be completed to establish confidence in these predictions. Future study should investigate Late Cenozoic paleoclimatic and paleoenvironmental conditions and their relation to selected geomorphic and geohydrologic processes (including erosion, deposition, ground-water movement, salt dissolution, and subsidence).

Caran and others (1985) and Holliday (1985a) used broadly based stratigraphic studies to characterize paleoclimatic conditions in the Southern High Plains and adjacent areas. These studies showed that existing regional depositional and erosional

chronologies are poorly constrained and often misunderstood. In addition, geohydrologic (recharge and discharge) and geomorphic (erosion and deposition) processes closely related to Late Cenozoic paleoclimate are not adequately covered by existing models. Preliminary estimation of escarpment retreat, stream incision and general denudation (Gustavson and others, 1981; Simpkins and Gustavson, 1987; Baumgardner, 1987) has been hampered by the incomplete and partly qualitative nature of the existing data base. Preliminary analyses of deflation (Machenberg and Caran, 1987) and dust fall (Machenberg, 1986) were based on short periods of modern record; such brief records should be supplemented through paleoenvironmental interpretation.

Evidence from Lake Basins

Tahoka Formation (Late Wisconsinan) lacustrine deposits of the Tahoka Lake Basin, Lynn County, Texas were termed the Tahoka beds by Evans and Meade (1945). Reeves (1976) divided the formation into the upper Brownfield Member and the lower Rich Lake Member distinguished at its top by the thin but widespread Vigo Park Dolomite. Based on radiocarbon dates and fossil evidence presented by Wendorf (1961), Hafsten (1961) and Oldfield and Schoenwetter (1964), Reeves concluded that the Tahoka sediments accumulated between 22,000 and 12,500 years ago and are therefore of Woodfordian age. He further observed that Tahoka sediments crop out around all the large pluvial lake basins and within many of the small basins where they are unconformably overlain by post-Tahoka lake and lee-dune deposits, "Peorian loess," or units of the Blackwater Draw Group. Characteristically, Tahoka sediments include sand, gravel, gypsum, thin dolomite, and black to blue-gray clay (Evans and Meade, 1945). Reeves (1976) reported that sepiolite is the dominant Tahoka clay mineral. According to Evans and Meade (1945) the maximum thickness of the Tahoka type section is at least 23 ft (7 m).

In larger lake basins, Tahoka strata typically overlie early Wisconsinan (Aftonian) lake deposits. The thickest stacks of these older sediments are usually offset from the centers of Tahoka deposition, indicating that the center of deposition moved in the lake basin. These early Wisconsinan sediments were named the Double Lakes Formation by Reeves (1976) for their occurrence in the Double Lakes depression, Lynn County, Texas. The type section (a core) consists of 28 ft (8.5 m) of dark, olive-gray to olive dense clay containing zones with epsomite and gypsum crystals (Reeves, 1976). Thin bands of very fine sand occur at depths of 48 to 55 ft (14.6 to 16.8 m).

Cummins (1893) used the name Tule beds for Pleistocene lacustrine strata exposed in Tule Canyon, Swisher County, Texas. These deposits were referred to as the *Equus* Zone or *Equus* beds by Wilmarth (1957) and were called the Rock Springs beds by Evans and Meade (1945). According to Kumanchan (1972) the Tule Formation is characterized by calcareous silt and clay-size volcanic ash (clay ash), thin-bedded limestone, sand, and gravel lenses. Kumanchan (1972) stated that the Tule beds in the type area unconformably overlie strata of the Ogallala Formation (late Miocene - Pleistocene) and red beds of the Triassic Dockum Group, and that the Tule is overlain by Pleistocene eolian "cover sands" (Blackwater Draw). An isopach of the type Tule deposits shows that maximum thickness of the unit in that area is about 120 ft (36.6 m) (Kumanchan, 1972). Frye and others (1948) correlated the Tule Formation with the Kansan Glacial Stage; this correlation was followed by Reeves (1976). Schultz (1986), on the basis of vertebrate fossils, molluscan assemblages (Cudahy), and radiometric dates for volcanic ash beds that vertically bracket the type Tule section, concluded that the Tule Formation was deposited from about 1.3 mya to about 0.6 mya during the Irvingtonian Land Mammal Stage.

Suggested Studies

A variety of the paleoclimatic indicators needed to summarize conditions in the Late Cenozoic can be developed using paleomagnetic, tephrochronologic, isotopic, paleoecologic and biostratigraphic techniques. General techniques for paleoclimatic reconstruction are discussed by Bradley (1985) and Berglund (1986). Possible application of isotopic indicators of paleotemperature (Stuvier, 1970; Lemeille and others, 1983) should be investigated. Palynological analysis should be attempted, although preservation of pollen and spores in sediments is generally poor (Bryant and Holloway, 1985; Hall, 1985). It is possible, however, that careful sampling of relatively impermeable lacustrine units will provide pollen and spore materials suitable for study (Oldfield and Schoenwetter, 1975). Fossil mollusks and ostracodes (Forester, 1985) provide a reliable record of paleoenvironmental conditions and they are abundant in many Late Cenozoic deposits of the area. A series of papers edited by Porter (1983) discusses a variety of other well-established indicators of paleoclimate that might have application to the Southern High Plains and adjacent areas. Hutchinson (1975), Friedman (1983), Lundelius and others (1983), Ruhe (1983), Smith and Street-Perrott (1983), and Spaulding and others (1983) treated isotopic, paleofaunal, pedologic, limnologic and paleofloral evidence, respectively. The stratigraphic and chronologic models on which most of these studies were based are poorly constrained and commonly in error. No modern comprehensive synthesis of regional climatology during the Late Cenozoic has yet been published; studies by Caran and others (1985) and Holliday (1985a), however, provide a framework.

HYDROCARBON RESOURCES

S. C. Ruppel

As of 1984, approximately 177 million barrels of oil and 100 billion cubic feet of gas had been produced from reservoirs in the Palo Duro Basin (Rose, 1986b). Most production has come from Pennsylvanian rocks, including siliciclastics and carbonates. Lower Permian (Wolfcampian and Leonardian Clear Fork Group carbonates) and Upper Permian (San Andres carbonates) rocks are also locally productive within the basin (fig. 1). Oil has also been produced from Mississippian carbonates in a single-well reservoir. Most of these oil and gas fields are located along the margins of the basin and are generally associated with bounding structures, including the Bravo Dome/Oldham Nose along the northwestern edge of the basin (fig. 1) and the Matador Arch along the southern margin (fig. 1). Until recently, very little production has come from the basin interior (Dutton, 1982; Rose, 1986b). This is unexpected, considering the size of the basin (approximately 14,000 mi^2; 36,000 km^2) and the thickness of the sedimentary sequence (as much as 13,000 ft; 3,000 m).

Hydrocarbon Potential

Several explanations have been offered for the relative scarcity of major hydrocarbon discoveries within the basin. Hryor (in Fritz, 1985) suggested that past exploratory efforts failed because data needed for efficient exploration, such as high quality seismic surveys, were lacking. Rose (1986a) stated that source rocks in the basin have not reached sufficient levels of thermal maturity to generate significant quantities of hydrocarbons. Dutton (1986) concluded that kerogen quality in Palo Duro Basin source rocks may be highly variable and, in some cases, insufficient to generate oil.

On the other hand, several studies produced agreement that most requirements for hydrocarbon generation and trapping (porosity, permeability, seals, traps, and source rocks) are present within the Palo Duro Basin (Dutton, 1980b; Ruppel, 1985; Rose, 1986a). Porosity sufficient to form suitable reservoirs has been reported, at least locally, throughout the entire Paleozoic section, including both siliciclastic and carbonate rocks. Oil shows have been reported from each of these potential reservoir units (Dutton, 1980b; Ruppel, 1985). Reservoir seals are formed by evaporites in the Permian section and by shales in the Mississippian and Pennsylvanian parts of the sequence (Rose, 1986a). Structural traps produced by both faults and anticlinal folds are common across the basin (Ruppel, 1985; Rose, 1986a). Potential source rocks are abundant in the Pennsylvanian and Permian parts of the section where sequences of shales rich in organic matter are common throughout much of the basin (Dutton, 1980b; Rose, 1986b). Mississippian carbonates may act as minor hydrocarbon sources along the eastern margin of the basin (Ruppel, 1985).

No consensus has been reached regarding the remaining element necessary for hydrocarbon accumulation, that is, thermal maturity. Rose (1986a) stated that the Palo Duro Basin is thermally immature. However, detailed studies of source rock geochemistry (Dutton, 1980b, 1982; Ruppel, 1985), including assessment of kerogen color (thermal alteration index), vitrinite reflectance, pyrolysis, and conodont color (color alteration index), indicate that levels of maturity necessary for the production of hydrocarbons have been reached in the basin. These studies imply that most, if not all, source rocks more than approximately 7,500 ft (2,285 m) deep have reached at least the minimum level of thermal maturity necessary for hydrocarbon generation and expulsion (Ruppel, 1985; Dutton, 1986). This includes nearly all Pennsylvanian and underlying source rocks. Although some basal Permian (Wolfcampian) rocks also appear to be thermally mature, most Permian and younger rocks are apparently immature (Dutton, 1986).

Levels of thermal maturity determined in these studies agree well with calculations based on present geothermal gradient and theoretical considerations (fig. 47). The discovery and production of oil from Mississippian and Pennsylvanian, as well as Permian, rocks along the southern and northern margins of the basin, and in a few cases, within the basin, further support this interpretation.

Although mature hydrocarbon source rocks are common in the Palo Duro Basin, recent investigations have shown the quality of these source rocks varies considerably. Dutton (1986) suggested that much of the kerogen present in Pennsylvanian source rocks (shales) in the Palo Duro Basin may be of a type that is more conducive to the generation of gas rather than oil.

Exploration Trends

Rose (1986b) grouped hydrocarbon production in the Palo Duro Basin into major trends. Three of these trends are associated with structures developed along the margins of the basins; these include (1) the Bravo Dome/Oldham Nose trend and (2) Whittenburg Trough trend on the northwestern margin of the basin and (3) the Matador Arch trend on the southern margin. In both of these areas, production comes from small reservoirs that are developed along small anticlines and faults. Production in the Bravo Dome/Oldham Nose area comes from multiple zones in the Pennsylvanian including siliciclastic (fan-delta sandstone) and carbonate (shallow water platform) reservoirs. The Whittenburg Trough trend includes reservoirs that produce principally gas from similar Pennsylvanian and Permian siliciclastic and carbonate reservoirs.

Most fields in the Palo Duro Basin are associated with faults along the Matador Arch; this trend also includes the greatest geological diversity, comprising siliciclastic and carbonate reservoirs ranging from Mississippian to Permian in age (fig. 48).

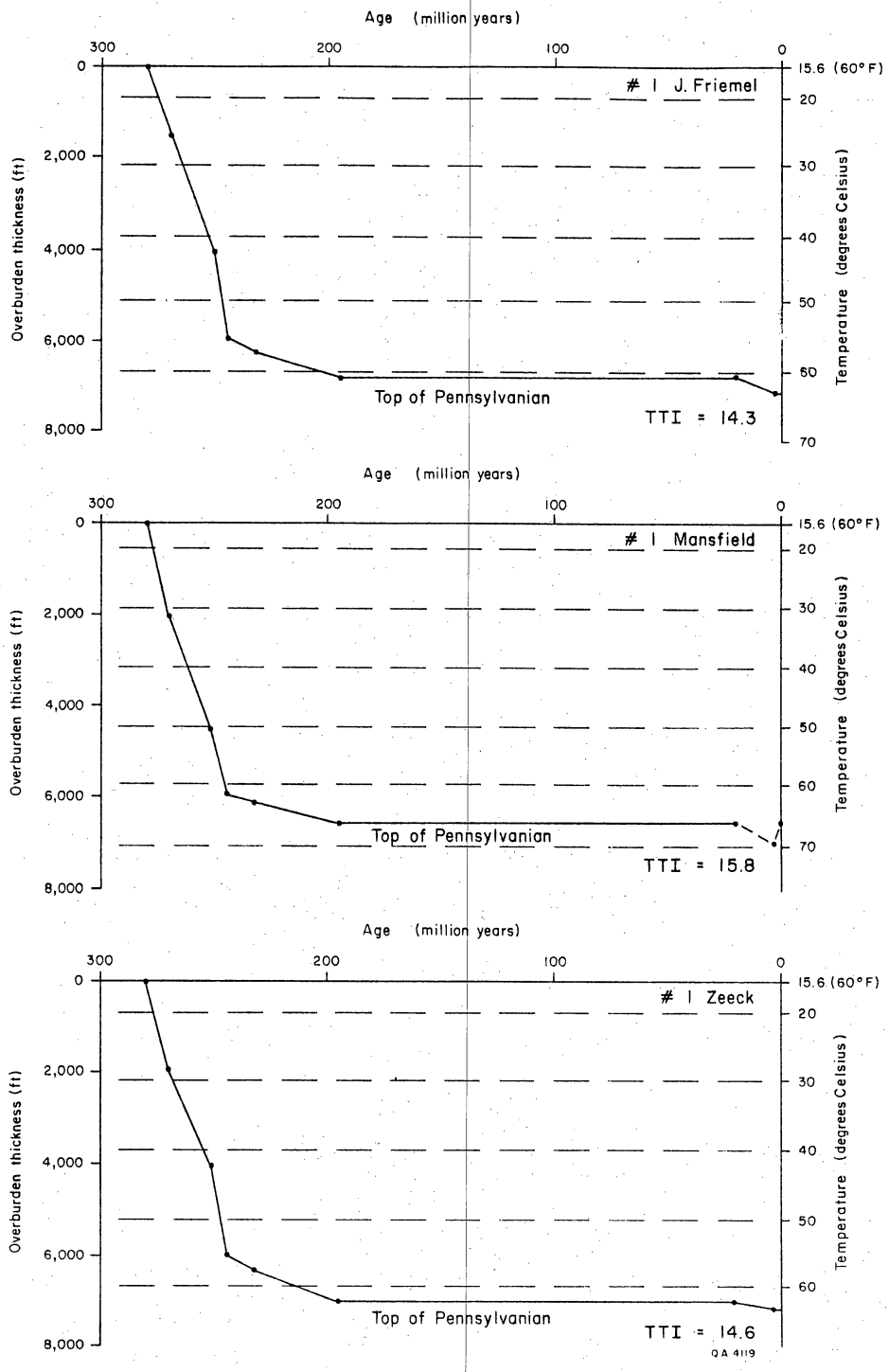


Figure 47. Burial history plots from cored wells in the Palo Duro Basin. Time-temperature index values (Wapples, 1980) are consistent with other data which indicate that rocks below about 7,500 ft (2,290 m) are sufficiently mature to generate hydrocarbons. From Dutton (1986).

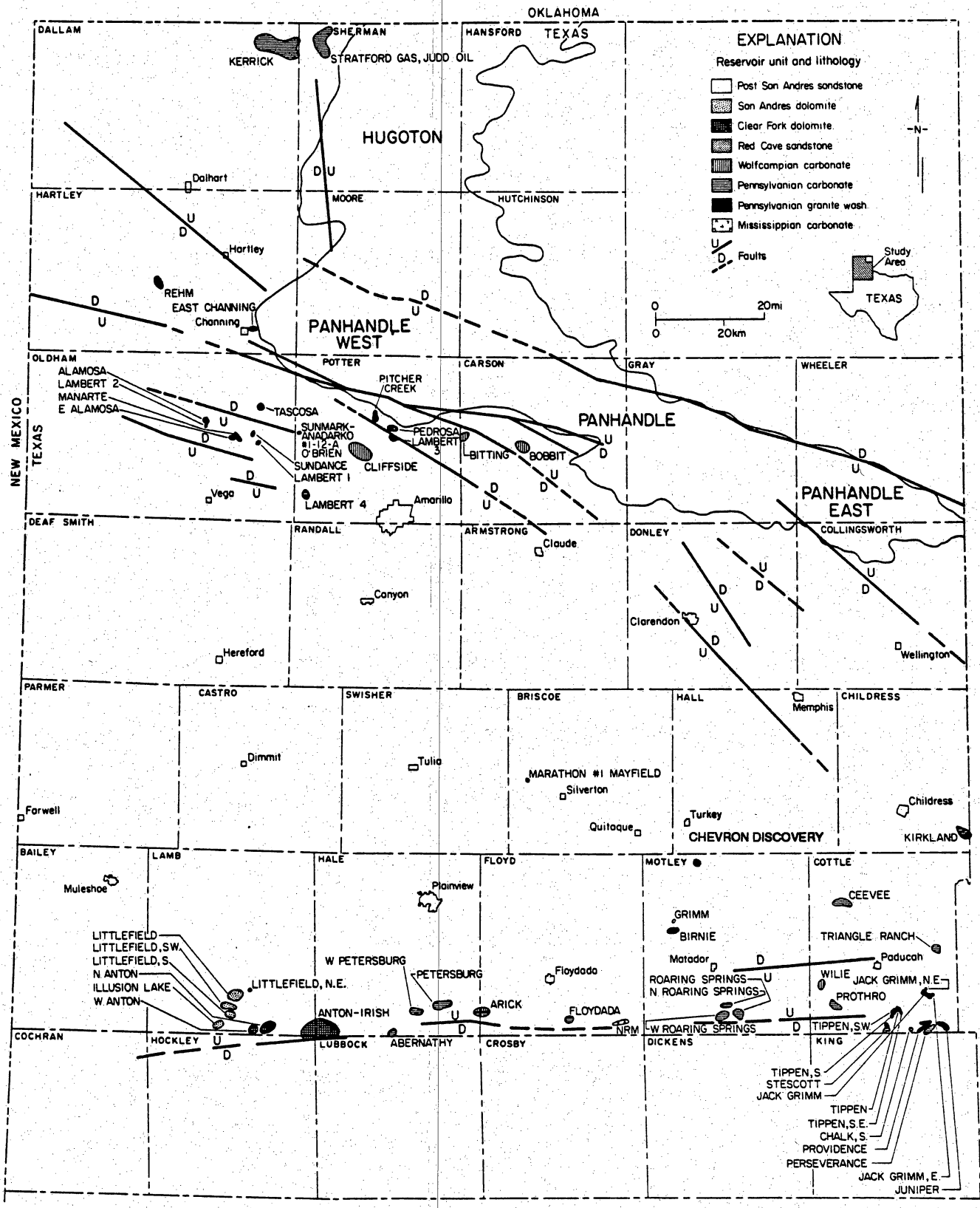


Figure 48. Location of oil and gas fields in the Palo Duro Basin and surrounding areas of the Texas Panhandle. From Dutton (1982).

Only two discoveries have been made in the central part of the basin. A single well produced about 20,000 barrels of oil from lower Pennsylvanian carbonates in 1981. This well, the Marathon Mayfield No. 1, was completed in a carbonate debris flow located immediately west of the eastern Pennsylvanian platform margin (fig. 48).

A second discovery has been recently made further south along this same platform margin. Chevron USA discovered oil in 1987 in northwestern Motley County, approximately 30 miles (50 km) southeast of the Marathon discovery (fig. 48). Early reports regarding this new field indicate that flow rates exceeded 340 barrels per day in the discovery well, the Chevron Mullin No. 1. The field is currently being developed; unconfirmed reports indicate that two additional productive wells have been completed and drilling is continuing on a fourth. The Chevron field is developed in Pennsylvanian limestone on the outer reaches of the Pennsylvanian platform margin with overlying Cisco shales forming the top seal.

On the basis of current productive trends and stratigraphic data, several potential reservoir trends have been identified in the Palo Duro Basin (Dutton, 1980a, 1982; Ruppel, 1985; Rose, 1986a). These include (1) Mississippian platform carbonates, (2) Lower Pennsylvanian (Bend Series) fan delta and fluvio-deltaic, coarse-grained sandstones, (3) Middle Pennsylvanian (Strawn Series) platform carbonates, (4) Middle Pennsylvanian (Canyon Series) platform margin carbonates, and (5) Lower Permian (Wolfcampian) platform and platform-margin carbonates.

Of these five, the first three extend across almost the entire Palo Duro Basin. Mississippian carbonates are absent along the northern margin of the basin (fig. 6); Lower Pennsylvanian coarse-grained siliciclastics thin in the deepest part of the basin. Potential Canyon and Wolfcamp Series carbonate reservoirs, on the other hand, follow the northwest-southeast trending platform margins (fig. 13). Although Rose (1986a) downgraded the potential of these potentially productive trends because of structural orientation and questionable associated traps, these two plays are well-situated with

regard to adjacent basinal source rocks and are sealed above by similar shales. Recent discoveries of substantial amounts of oil in Canyon Series platform-margin carbonates along the eastern platform margin indicate that these trends may contain significant quantities of hydrocarbons.

Lower (Clear Fork) and Upper Permian (Clear Fork) platform carbonates are productive along the southern margin of the basin (Matador Arch) and are also potential targets (Ramondetta, 1982a, 1982b).

Unresolved Issues

Several issues remain unresolved regarding the petroleum potential of the Palo Duro Basin. Foremost among these is the question of whether or not source rocks in the basin are sufficiently mature to have generated significant quantities of hydrocarbons. As outlined above, the data are equivocal and interpretations have varied significantly. Perhaps the most compelling evidence that parts of the basin have become mature enough to generate hydrocarbons is the continued discovery of small oil and gas fields throughout the basin. Although most of these discoveries have been along the margins of the basin, a sizeable oil reservoir was recently discovered in the center of the basin. This discovery will certainly rekindle the interest of even those who doubt the potential of the basin. Further study is needed, however, to determine whether or not this recent discovery is an indication of conditions in the basin as a whole or merely represents a small accumulation due to very specific and localized conditions. Additional studies of the organic geochemistry of oils and source rocks are needed to improve our knowledge of the area.

GROUND-WATER FLOW

A. R. Dutton

Introduction

To develop conceptual models of ground-water flow and to predict flow velocities and travel times in the Palo Duro Basin, hydrogeologic studies between 1979 and 1987 focused on describing and explaining the relationship between hydraulic-head distribution, hydrologic properties, water chemical composition, and lithology (Dutton and others, 1987). This section summarizes study methods and data quality used in the investigations, reviews the conceptual and numerical models of the flow systems, and identifies significant issues and uncertainties that remain in developing and applying the conceptual and numerical hydrologic models.

Because of the scarcity of available field data, most estimates of recharge and discharge rates, ground-water flow velocities, and ground-water residence times have been calculated from ground-water flow model applications. In evaluating the quantitative estimates obtained from modeling, it is important to consider which conceptual framework was used to describe the ground-water flow regime and which data were used for hydraulic properties and boundary conditions. Developing an understanding of a regional flow system for which few data are available is necessarily an iterative process in which initial numerical models need to be refined as different conceptual models are proposed and tested, and as additional data become available.

McNeal (1965) and Handford (1980b) presented preliminary interpretations of the hydrogeologic setting of deep-basin brine-bearing units in the Midland and Palo Duro Basins. Handford (1980b) recognized that flow in the regional, deep-basin system is distinct from flow in the shallow system (fig. 49). Bassett and Bentley (1982, 1983)

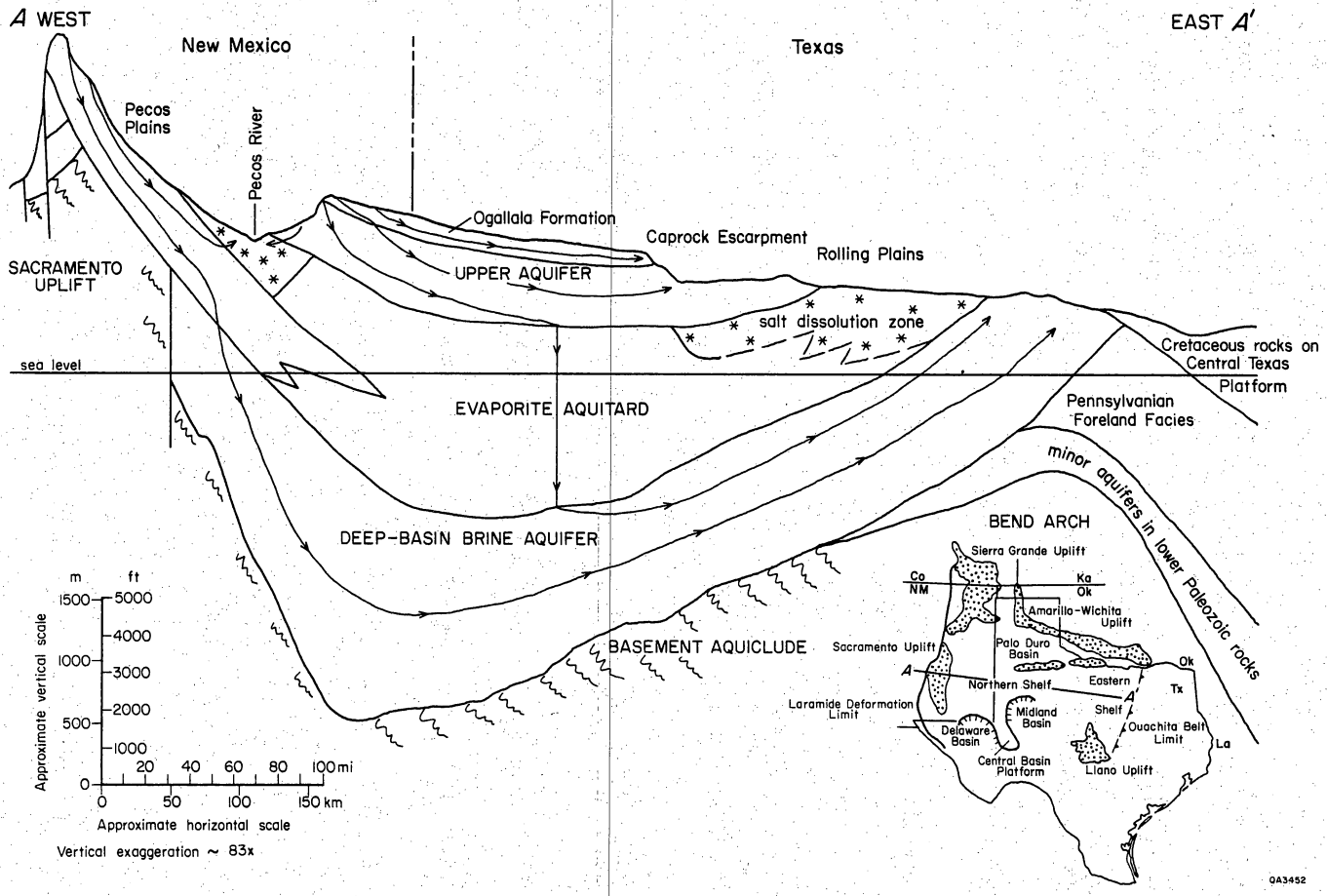


Figure 49. Idealized hydrostratigraphic cross section. Use of the term "evaporite confining system" is preferred to "evaporite aquitard". From Jorgensen and others (in press).

formally divided the stratigraphic section into three major hydrogeologic units that reflect this distinction: (1) an upper fresh-water aquifer system in rocks of the Tertiary Ogallala Formation and Triassic Dockum Group (see DuBar, this volume), (2) a low-permeability evaporite aquitard composed of the Leonardian to Ochoan halite, anhydrite, carbonate rock, and shale in which the proposed repository would be built (U.S. Department of Energy, 1986), and (3) a deep-basin brine aquifer system in pre-Leonardian rocks (see Ruppel, this volume). They further subdivided the upper and deep-basin aquifer units on the basis of hydrogeologic properties (compare tables 1 and 6). Although their properties vary, these hydrogeologic units occur throughout the Southern Great Plains, which lies south of the Canadian River, east of the Basin and Range Province, and west of the Balcones Escarpment-Ouachita Tectonic Belt in central Texas and eastern Oklahoma (Jorgensen and others, in press), and includes the regions adjacent to the Palo Duro Basin, such as the Midland Basin.

Hydrogeologic Data

The amount of hydrogeologic data from wells decreases with increasing depth of ground-water sampling. In the Palo Duro Basin, most hydrogeologic data are available from water-supply wells in the Ogallala Formation and from wells in the unconfined aquifer in the Dockum Group, but data from the more deeply buried, confined aquifer in the Dockum Group beneath the Southern High Plains are sparse. With little hydrocarbon exploration and development and relatively little drilling to depths greater than 1,000 ft, data are sparse in the Palo Duro Basin for deeper units. Most data from the deeper units is from oil and gas wells in the surrounding basins. Wells drilled as part of the U.S. Department of Energy (DOE) testing program provide the only quality-controlled hydrogeologic data for the evaporite section and deep-basin brine aquifer in the Palo Duro Basin. Locations of DOE wells at which hydro-

Table 6. Generalized stratigraphic column, lithology, and hydrogeologic units. Use of the term Evaporite confining system is preferred over the term Evaporite aquitard. Modified from Bassett and Bentley (1983).

System	Series	Group	General lithology and depositional setting	Hydrogeologic element	Hydrogeologic unit
Quaternary			Fluvial and lacustrine clastics	Ogallala aquifer	Upper aquifer
Tertiary					
Cretaceous			Nearshore marine clastics		
Triassic		Dockum	Fluvial deltaic and lacustrine clastics and limestones	Dockum aquifer	
Permian	Ochoa		Salt, anhydrite, red beds and peritidal dolomite	Evaporite aquitard	Evaporite aquitard
	Guadalupe	Artesia			
		Pease River			
	Leonard	Clear Fork			
		Wichita			
	Wolfcamp		Shelf and platform carbonates, basin shale and deltaic sandstones	Wolfcamp carbonate aquifer	Deep-basin brine aquifer
Pennsylvanian				Pennsylvanian carbonate aquifer	
			Basin shale aquitard	Upper Paleozoic granite wash aquifer	
Mississippian			Shelf limestone and chert	Lower Paleozoic carbonate aquifer	Basin shale aquitard
Ordovician		Ellenburger	Shelf dolomite		
Cambrian			Shallow marine (?) sandstone	Lower Paleozoic sandstone aquifer	
Precambrian			Igneous and metamorphic	Basement aquiclude	Basement aquiclude

logic tests were made and from which fluid samples have been collected and analyzed are shown in figure 50; production zones are listed in table 7. Interpretations based on these sparse data have been supported with chemical analyses reported for other wells located around the Palo Duro Basin, including oil and gas exploration and production wells and water supply wells. Data for directly interpreting aspects of the flow system also include chemical analyses of (1) springs located across the Rolling Plains and (2) fluid inclusions in halite obtained from DOE core samples. Core from DOE test wells also was analyzed to obtain porosity and permeability data. The reader is referred to the many articles cited in this report for a listing and detailed discussion of data from the test wells.

Water-level measurements were compiled from computerized records at the Texas Department of Water Resources and published reports of water resources in counties in the Texas Panhandle and eastern New Mexico. Water levels were measured in irrigation, stock, and domestic supply wells during several decades. Some measurements undoubtedly were made during the irrigation season and were affected by drawdown. Dutton and Orr (1986) and Dutton and Simpkins (1986) assumed that historic and seasonal changes in ground-water level at any location are small relative to regional variations in the potentiometric surface. Where several measurements were made at one well, the highest water level was used.

Drill-stem tests (Black, 1956; Bredehoeft, 1965; Hackbarth, 1978) were used to measure permeability and fluid pressure in deep wells drilled for DOE-sponsored tests in the Palo Duro Basin as well as in hydrocarbon-exploration wells in adjacent basins. Given a complete test record, fluid pressure and permeability can be accurately found by methods analogous to those used for recovery tests in water wells (Earlougher, 1977; Kruseman and De Ridder, 1983). Data reduction for test results from DOE program wells followed conventional methods (Matthews and Russell, 1967; Earlougher, 1977). Pressure-buildup data showed varying degrees of wellbore-storage and skin

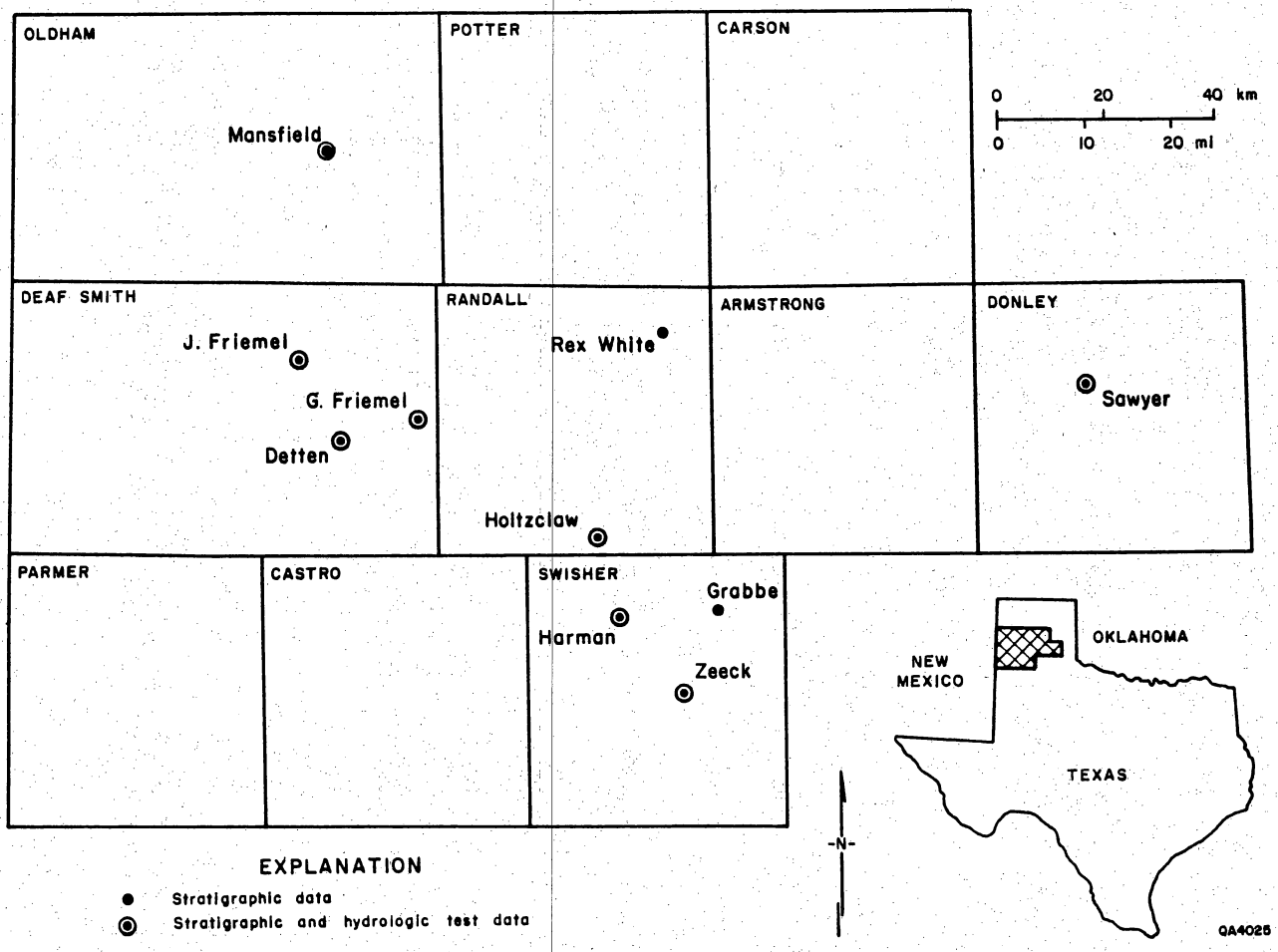


Figure 50. Location of DOE test wells used for collection of core and geophysical logs (stratigraphic data), water samples for geochemical analysis, and hydrologic tests. Modified from Dutton and Orr (1986).

Table 7. Production zones in U.S. Department of Energy test wells in the Palo Duro Basin

Well	Land Elev. (ft)	Zone	Formation	Depth (ft) top	Depth (ft) bottom
Sawyer No. 1	2,580	1	Ellenburger carbonate	4,716	4,746
		2	Ellenburger carbonate	4,604	4,640
		3	Mississippian carbonate	4,500	4,510
		4	Pennsylvanian sandstone	4,258	4,342
		5	Wolfcampian carbonate	3,172	3,189
Sawyer No. 2	2,580	1	San Andres unit-4 carbonate	762	784
Mansfield No. 1	3,668	1	Wolfcampian carbonate	4,818	4,890
		2	Wolfcampian carbonate	4,514	4,638
Mansfield No. 2	3,668	1	Seven Rivers sandstone	723	778
Zeeck No. 1	3,411	1	Pennsylvanian carbonate	7,140	7,230
		2	Wolfcampian carbonate	5,603	5,640
		3	Wolfcampian carbonate	5,470	5,550
		4	San Andres unit-4 carbonate	2,930	2,970
J. Friemel No. 1	4,016	1	Pennsylvanian sandstone	8,168	8,204
		2	Pennsylvanian sandstone	8,122	8,132
		3	Pennsylvanian sandstone	8,040	8,050
		4	Pennsylvanian sandstone	7,840	8,050
		5	Pennsylvanian sandstone	7,707	7,734
		6	Pennsylvanian carbonate	7,300	7,326
		7	Wolfcampian carbonate	5,825	5,926
		8	San Andres unit-4 carbonate	2,754	2,798
		9	Queen & Grayburg sandstone		
Harman No. 1	3,523	1	Salado fractured siltstone	1,124	1,218
Detten No. 2	3,855	1	Seven Rivers sandstone	1,305	1,325

(wellbore damage or stimulation) effects, sometimes limiting data reduction for permeability calculations to type-curve matching (Agarwal and others, 1970). Semilogarithmic plots of fluid pressure against t"/t, the ratio of test time to shut-in time, were used to estimate formation fluid pressure from drill-stem tests. Such plots are known as both Theis recovery plots (ground-water literature) and Horner plots (petroleum engineering literature). The accuracy of pressure extrapolation is variable, depending on incidence of wellbore storage and test duration, but generally is less than ± 25 psi (± 0.17 MPa). Simulation of hydrologic tests using computer models is a useful approach for interpreting permeability and for assessing effects of various boundary conditions, including wellbore-storage and skin effects, but it was not developed until late in the DOE program.

Unfortunately, only pressures observed at the end of each flow and shut-in period are reported in commercial data sources, prohibiting validation of data quality and calculation of reliable permeability estimates. Common errors in reported test results include (1) insufficient shut-in time in cases of low permeability, (2) clerical error in reporting pressures, (3) tool failure or loss of packer seal, (4) poorly calibrated equipment, and (5) supercharging (Bassett and Bentley, 1983; Bair and others, 1985; Bair, 1987b). The shut-in pressure data generally also underestimates true formation-fluid pressure. Drill-stem test data for oil and gas wells were culled from listings purchased from Petroleum Information Corporation, Incorporated. Culling procedures were based on the number and internal consistency of the shut-in pressures (Bair and others, 1985; Smith and others, 1985; Bair, 1987b). Bassett and Bentley (1982, 1983), Bair and others (1985), Smith and others (1985), Dutton and Orr (1986) and Bair (1987b) deleted data values which appeared too high to be considered reasonable; underpressured values were thought due to historic production of oil and water from the deep formations.

Data on chemical composition of ground water in shallow aquifers also was compiled from computerized records at the Texas Department of Water Resources and published reports. Data quality is limited: (1) temperature, pH, and alkalinity values in most reported chemical analyses probably are unreliable estimates of *in situ* values because measurements were not made immediately at well sites, and (2) sodium and potassium concentrations mostly were computed together from the difference between equivalent charge of anions and cations, as is indicated by an exact charge balance.

Data on chemical composition of ground water in deep, saline formations was obtained from commercial data sources such as the University of Oklahoma Petroleum Data Service, later Dwight's Data Service. Samples of brine from oil fields are of two types: those collected during drill-stem tests and those obtained from a wellhead or storage tank at producing wells. Oil-field brine samples may be contaminated by drilling mud or by reaction with well hardware or storage tanks (Bassett and Bentley, 1983). Dissolved gases in samples from separator tanks obviously do not reflect *in situ* conditions. Also, temperature, pH, and alkalinity probably are not measured on site and samples probably are not treated to preserve unstable constituents until laboratory analysis. As in shallow ground-water samples, charge balance is almost always exact, indicating that sodium and potassium concentrations were computed together. Nonetheless, Bassett and Bentley (1982, 1983), Dutton and Orr (1986), and Bein and others (in preparation) assumed that regional trends that are apparent in the geographically distributed data are meaningful and real.

Water samples were collected from DOE test wells (fig. 50, table 7) and from wells in the High Plains aquifer and Dockum Group to supplement the data base and to determine trace constituent and isotopic compositions. Similar methods were followed for collection and treatment of both subsurface brine and shallow ground water. Samples were filtered and treated and temperature, pH, and titrated alkalinity were measured at well sites generally following procedures recommended by Brown and others (1970), Wood (1976), and Lico and others (1982). Specific details of sampling

procedures are given in topical reports.

Data Analysis and Synthesis

Development of valid conceptual models must be based on consistent explanation of both hydrologic and geochemical data of each hydrogeologic unit and of the regional flow system as a whole. Analysis of hydrogeologic data to support interpretation of the flow system generally followed several steps: (1) description of field relations between hydrologic variables, chemistry of water, and stratigraphy and mineralogy; (2) recognition of unique hydrogeologic attributes of the unit; (3) measurement, analysis, and simulation of chemical and hydrologic variables that control the environment; and (4) interpretation of the variables at regional or local scales within a theoretical framework. These steps follow those summarized by Back and Hanshaw (1965) for chemical hydrogeologic studies. Tools applied in the analysis of the Southern Great Plains flow system were diverse, including but not limited to (1) hydrochemical facies mapping, (2) geochemical modeling, (3) analysis of potentiometric surfaces, and (4) numerical simulation of ground-water flow. Ground-water flow was simulated in two-dimensional vertical cross-sectional and horizontal plan-view models and in three-dimensional models.

Hydrochemical Facies

Hydrochemical facies are named for the ions that account for at least 50 percent of the total equivalent concentration as depicted in Piper diagrams (Back, 1966); mixed-cation and mixed-anion hydrochemical facies, as used here, are waters in which no one cation or anion is dominant. A map of hydrochemical facies is one tool used

to interpret flow patterns among hydrogeologic units. Associations of typical hydrochemical facies have been mapped for ground water in the Ogallala Formation by Nativ and Smith (1987), in the Dockum Group by Dutton and Simpkins (1986), and in San Andres carbonate rock by Dutton and Orr (1986). Similar maps for other hydrogeologic units, notably the Pennsylvanian System and Wolfcampian Series, have not been made. Sparse data coverage and highly variable water compositions result in considerable uncertainty for placement of the hydrochemical facies boundaries and total dissolved solids isopleths.

Geochemical Modeling

Several geochemical models were used to assist interpretation of water-sample chemistry. SOLMNEQ (Kharaka and Barnes, 1973) and AQ/SALT (Bassett and Griffin, 1982) were used to calculate saturation with respect to mineral phases. AQ/SALT uses a rearranged form of the Pitzer equation to calculate ion-activity coefficients and should be accurate at high salinity; its mineralogic data base, however, is limited and lacks thermodynamic parameters for calculating activities of dissolved carbonate species. EQ3NR (Wolery, 1983) also is based in the Pitzer equation and is more generally applicable than AQ/SALT; however, it was not available to researchers in the DOE program until 1987. SOLMNEQ uses the B-dot method to calculate ion-activity coefficients. Although it may be less accurate than AQ/SALT or EQ3NR in calculating activity coefficients for dissolved species in saline water, it includes carbonate species in its data base and is satisfactory for fresh to slightly brackish waters. PHREEQE (Parkhurst and others, 1982) and EQ6 (Wolery, 1984) are designed to model geochemical reactions, such as the evolution of solution composition in equilibrium with multiple minerals. PHREEQE is based on an ion-pairing model and is inaccurate in high-salinity solutions; EQ6 is based on the Pitzer equation and is applicable to fresh water and brine solutions.

Analysis of Potentiometric Surfaces

Hydraulic head, or potentiometric surface, maps (Hubbert, 1940, p. 909-913) have been constructed using equivalent-fresh-water hydraulic-head values. This approach has been made standard by numerous previous studies (McNeal, 1965; Hanshaw and Hill, 1969; Hitchon, 1969, 1984; Bond, 1972). However, the approach can lead to significant errors in interpretation of flow direction if applied where ground-water density greatly varies (Jorgensen and others, 1982) or where permeability contrasts exist. Both problems limit the straightforward interpretation of potentiometric surface maps and cross sections to predict flow direction in the Palo Duro Basin.

Geostatistical techniques were used in a few studies to determine regional, potentiometric surfaces, complementing conventional contouring of hydraulic-head data (Orr and Dutton, 1983; Smith and others, 1985; Dutton and Orr, 1986). The approach minimized variance caused by heterogeneous data and by measurement and analytical errors, previously summarized. Geostatistical techniques include variogram analysis and kriging (David, 1977; Clark, 1979; Royle and others, 1980) The variogram shows expected values of square differences in hydraulic head between pairs of wells within a given distance and at multiples of that distance. Variogram analysis was used to determine a set of parameters—range, nugget, and sill—that describe how hydraulic-head values vary spatially. Kriging computed spatially averaged values of hydraulic head with minimum estimation variance based on the parameters determined from the variogram. Computer programs used by Orr and Dutton (1983), Smith and others (1985), and Dutton and Orr (1986) for geostatistical analysis were by Knudsen and Kim (1978) and calculated experimental variograms having a spherical model and performed block kriging.

Hydrologic Modeling

Senger and Fogg (1987), Senger and others (1987a, 1987b) and Atwood and others (1986) modeled ground-water flow in two-dimensional vertical cross sections. Senger and Fogg (1987) and Senger and others (1987a, 1987b) used the generalized east-west geologic cross section of Bentley (1981), Bassett and others (1981), and Bassett and Bentley (1982, 1983), which depicts the major depositional facies from postulated recharge areas in New Mexico to an artificial boundary located just east of the Texas-Oklahoma border. That modeled section does not reach outcrops of hydrogeologic units or hypothesized discharge zones across North-Central Texas and South-Central Oklahoma. Senger and others (1987a) distinguished ten different facies in the model. They explicitly modeled a salt dissolution zone on either side of the High Plains and included the San Andres unit 4 carbonate bed as a separate layer. Permeability values used by Senger and Fogg (1987) and Senger and others (1987a, 1987b) were based on Myers (1969), U.S. Geological Survey open-file data, and Wirojanagud and others (1986) and included limited permeability data derived from pumping tests and drill-stem tests (Dutton and Orr, 1986) to obtain a permeability value for the San Andres unit 4 carbonate bed. The model was implemented with the finite-element computer code FREESURF (Neuman and Witherspoon, 1970) which solves for both hydraulic head h and stream function Ψ in steady state flow equations (Senger and Fogg, 1987; Senger and others, 1987a). The computer program FLUMPS, which also employs the finite-element method, (Neuman and others, 1982) was used to simulate transient ground-water flow during the Cenozoic (Senger and others, 1987a, 1987b).

Atwood and others (1986) used a cross section that extends from the Pedernal Uplift in New Mexico across the Tucumcari, Palo Duro, and Hardeman Basins. The evaporite confining system was subdivided into eight units and the deep-basin brine

aquifer system was subdivided into 19 units. That model did not include the upper aquifer system and did not separately discretize the San Andres unit-4 carbonate bed. Atwood and others (1986) extended the units in the evaporite confining system to their respective outcrop areas in the west without varying permeability to represent lithologic changes. The profile included several faults that died out in the evaporite section. Atwood and others (1986) used permeability data from analyses of complete and incomplete DST records (Smith and others, 1984) and assigned textbook values to rock units that lacked data. Permeabilities were assumed to be higher along the faults than in adjacent strata only where the faults coincide with displacement of basement rocks. Flow through the hydrogeologic profile was simulated using the program FEDAR, a steady-state finite-element code (Christian and Trudeau, 1980).

Wirojanagud and others (1986) and INTERA (1984, 1986) modeled flow in two-dimensional horizontal (plan-view) layers. Wirojanagud and others (1986) evaluated two different single-layer models: one representing the aquifer in Wolfcampian rocks and the other representing the entire deep-basin brine aquifer system, which includes the aquifer in Wolfcampian rocks. Both models included leakage from the overlying evaporite confining system. Wirojanagud and others (1986) calculated average permeabilities from drill-stem test and pumping-test data, core analyses, and textbook values. Spatial variations in rock type within eight lithologic facies were taken into account through the use of vertically-averaged hydraulic conductivities. The steady-state model was implemented using the finite-element program TRAVEL (Charbeneau and Street, 1972).

INTERA (1984) modeled ground-water flow in the shallow, fresh-water aquifers and in the deep-basin brine aquifers as separate flow systems but assumed there was no vertical leakage through the evaporite confining system. INTERA (1986) expanded the previous model to a fully three-dimensional model by including flow through the evaporite confining system, which was divided into an upper evaporite layer, the San

Andres unit 4 carbonate bed, and a lower evaporite layer. The upper flow system was represented as two to three distinct layers: Ogallala Formation, Dockum Group, and Dewey Lake Formation; and the deep-basin flow system was divided into four layers: Wolfcampian Series carbonate, Pennsylvanian and Wolfcampian shale, Pennsylvanian carbonate, and Pennsylvanian granite wash (INTERA, 1984, 1986). Permeability values used by INTERA (1984, 1986) were based on the data and suggested values reported by Bassett and others (1981, 1982). In addition, INTERA (1986) used data presented by Stone and Webster Engineering Corporation (1984a, 1984b), Core Laboratories (1972b, p. 206), and various data on salt permeability in locations outside the Palo Duro Basin to assign values for the evaporite confining system and the San Andres unit 4 carbonate bed. Spatial variability of hydraulic conductivity assigned to deep-basin aquifer units represented changes in depositional facies near the Amarillo Uplift and Oldham Nose.

Senger and others (1985) constructed a three-dimensional numerical model to simulate ground-water flow in the deep-basin aquifer system. The model allowed for leakage through the overlying evaporite section. The deep-basin aquifers were divided into two Wolfcampian units, a so-called Brown dolomite layer and a layer below the Brown dolomite, and two Pennsylvanian strata separated by a layer representing a basin-shale deposit. Spatially distributed permeabilities used in the three-dimensional model of Senger and others (1985) were based on permeability measurements of Wolfcampian core correlated with interpreted geophysical log data. Permeabilities of the Pennsylvanian strata were the same as those used by Wirojanagud and others (1986).

All modeling studies to date ignored effects of variable density brine and have calibrated model parameters and boundary conditions using equivalent-fresh-water hydraulic-head data. As previously mentioned, significant errors in predicted flow velocities and flow directions can result from discounting effects of variable density (Davies, 1987).

Hydrogeologic Units

High Plains Aquifer

The High Plains aquifer consists primarily of an aquifer unit in the Neogene Ogallala Formation, with permeable units also in Cretaceous Trinity and Fredericksburg Groups and Quaternary Blackwater Draw Formation (table 1). The High Plains aquifer beneath the Southern High Plains is separated from the Central High Plains by the Canadian River valley. Most information is derived from the Ogallala Formation, although it is possible that water-well drillers have not distinguished between Blackwater Draw and Ogallala rocks when reporting drilling logs. Cretaceous rocks do not underlie the Ogallala in the Palo Duro Basin. The High Plains aquifer is sometimes defined to include the Dockum Group (Knowles and others, 1984). Dutton and Simpkins (1986), however, argued that the ground water in Triassic rocks (table 6) should be classified separately from the High Plains aquifer because of marked differences in hydraulic head and hydrochemical composition. The Dockum is treated separately from the High Plains aquifer in this report.

The water table in the Ogallala Formation is inclined to the southeast beneath most of the Southern High Plains (fig. 51); north of Palo Duro Canyon it is inclined to the east or northeast due to the locally different slope of land surface (Gutentag and Weeks, 1980; Gutentag and others, 1984; Knowles and others, 1984). In most areas during the early 1980s, the unconfined water table was at depths from 100 to 300 ft (30 to 90 m) beneath land surface; water levels have declined by as much as 100 ft (30 m) from the 1940s to the 1980s due to ground-water production which greatly exceeded recharge (Knowles and others, 1984). The saturated thickness, specific yield, and hydraulic conductivity of the aquifer generally vary between alluvial fan and interfan deposits in the Ogallala Formation (Nativ and Smith, 1987).

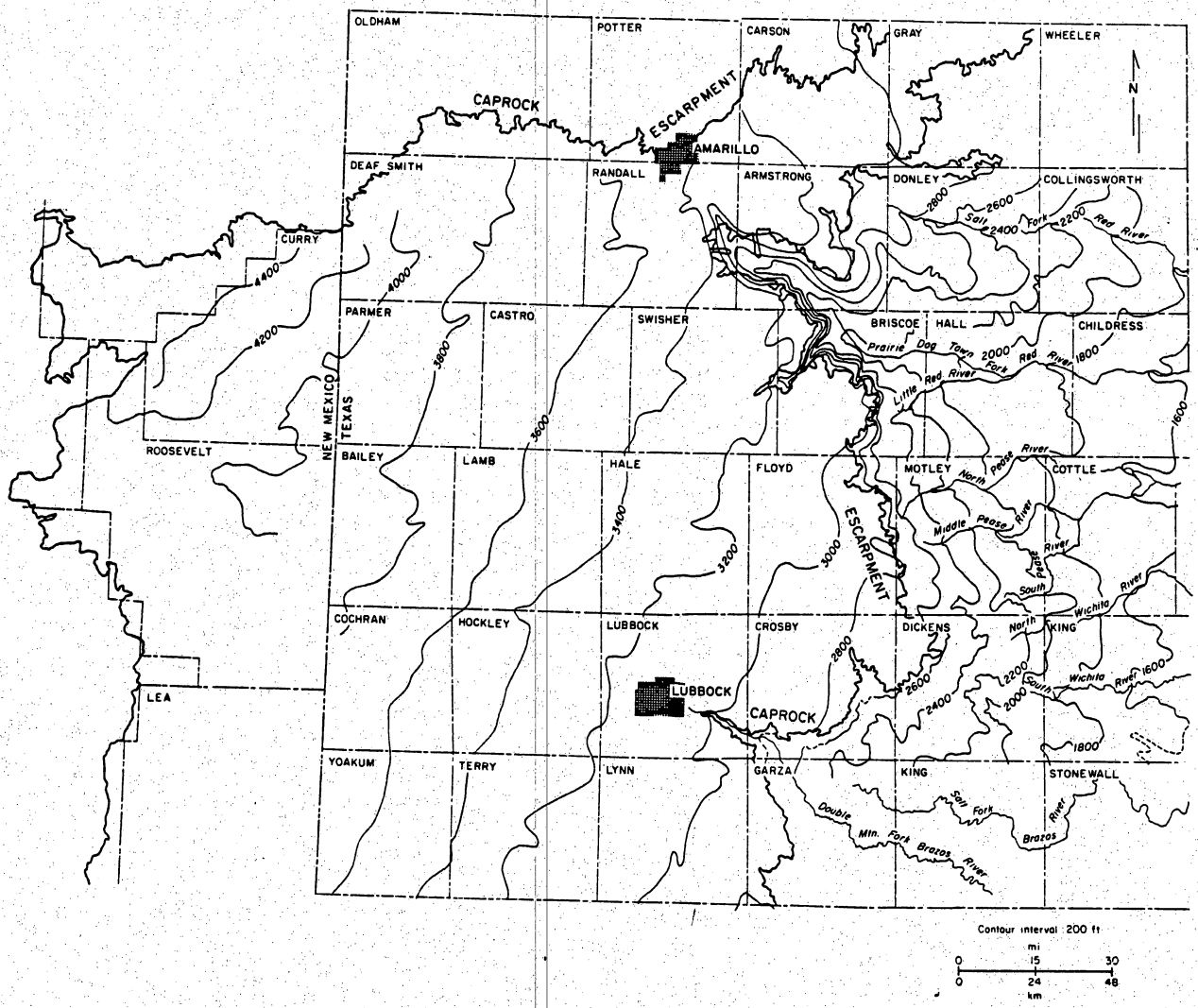


Figure 51. Hydraulic head in unconfined aquifers in the Ogallala Formation across the Southern High Plains and in Permian Formations across the Rolling Plains. From Bassett and others (1981).

Chemical composition of ground water also reflects the different sedimentary facies; for example, hydrochemical facies in Ogallala Formation ground water (Nativ and Smith, 1987) are strongly related to abundance of sand and gravel in the Ogallala Formation, mapped by Seni (1980). The mixed-cation–HCO_3 facies (fig. 52, area 7) beneath the northern part of the Southern High Plains coincides with an area where sand and gravel make up more than 60 percent of formation thickness. The Ca–HCO_3 hydrochemical facies in Randall County (fig. 52, area 6) overlaps an area where sand and gravel make up less than 50 percent of the formation. Total dissolved solids (TDS) concentration is less than 400 mg/L beneath most of the northern part of the Southern High Plains; TDS exceeds 600 mg/L in an area between the cities of Hereford and Canyon in Deaf Smith and Randall Counties, respectively, and in northeastern Carson and northern Gray Counties (Knowles and others, 1984, fig. 39). Samples from the aquifer beneath the central and southern parts of the Southern High Plains, where percentage of sand and gravel is variable but generally less than 60 percent (Seni, 1980), commonly have no predominant cations or anions (fig. 52, areas 14 and 19). Ground-water salinity increases to more than 600 mg/L in the southern part of the Southern High Plains generally in the area underlain by Cretaceous rocks. Hydrogeologic controls of the regional salinity increase are not understood, but might be related to the occurrence of saline lakes (Wood and others, 1986), irrigation return flow (Knowles and others, 1984), or discharge of ground-water from Cretaceous rocks (Nativ and Smith, 1987).

Recharge to the High Plains aquifer is limited, as is recharge in general in the Modern arid to subhumid climate of the southwestern United States. Infiltration of meteoric water is considered the primary source of Holocene and Modern water in the Ogallala Formation, based on measured values of $\delta^{18}O$ and δD (Nativ and Riggio, 1987; Nativ and Smith, 1987). Winograd (1974) and others suggested that effective recharge in this region is located only in intermittent and perennial stream

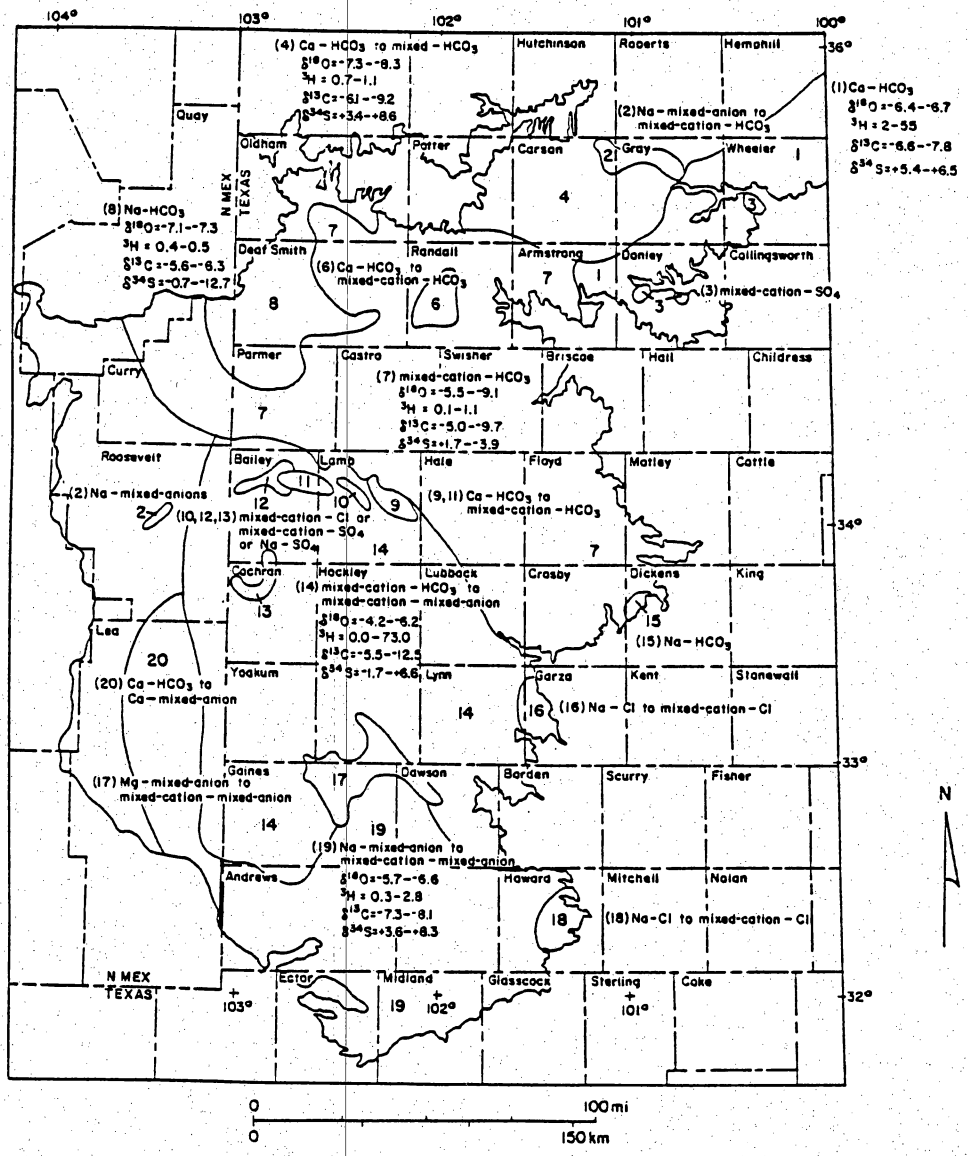


Figure 52. Hydrochemical facies, tritium activity, and stable isotopic composition of ground water in the Ogallala Formation. From Nativ and Smith (1987).

channels that cross aquifer outcrops and that infiltration of precipitation on interchannel areas contributes little recharge. On the Central and Southern High Plains, however, only a few widely separated stream channels are present and probably drain less than two percent of precipitation runoff (Ostercamp and Wood, 1984). Lotspeich and others (1971) observed that, owing to ground-water production rate being so much in excess of recharge rate, Ogallala water essentially is being mined without being replenished, and that the only meaningful source of recharge is in the more than 30,000 playa-lake basins on the Central and Southern High Plains which collect more than 80 percent of precipitation runoff (Dvoracek and Black, 1973). Playas are ephemeral lakes, generally less than approximately one mile in diameter, that occur in internally drained, roughly circular watersheds perched high above the water table (see DuBar, this volume).

The proportion of Ogallala Formation ground water that is recharged from (1) infiltration of precipitation across the Southern High Plains and (2) percolation of runoff that collects in playa lakes is poorly known. Research on Modern and Holocene recharge rates in playa and interplaya areas are incomplete. Previous studies have determined models for the origin and development of playa-lake basins and estimated recharge rates across the High Plains. A reasonable model for playa origin is through deflation, ponding, infiltration, chemical solution, and karstic subsidence (Caran and DuBar, in press; Osterkamp and Wood, 1987; Wood and Osterkamp, 1984, 1987). This explanation for playa-basin formation is consistent with a hydrologic model that assumes that these lake basins focus recharge to aquifers. In other hydrologic models, water that accumulates in playa-lake basins is assumed to evaporate subsequently and not recharge ground water; in such models, recharge is distributed throughout "interbasin" areas. Local recharge estimates vary widely from 0.01 to 1.6 inch/yr (0.025 to 4.1 cm/yr) between interplaya and playa areas, respectively (Cronin, 1964; Havens, 1966; U.S. Bureau of Reclamation, 1982; Stone, 1984, 1987;

Wood and Osterkamp, 1984). In comparison, regional recharge estimates of approximately 0.1 to 0.2 inch/yr (0.24 to 0.5 cm/yr) based on numerical model calibration (Knowles and others, 1984; Luckey and others, 1986) fall within this range, possibly reflecting spatial weighting of several recharge sources.

Evaporation and infiltration are the two major mechanisms that control water loss from playa lakes. If permeability of soils and caliche beneath playa lakes is very low, significant infiltration is prevented and much of the water must eventually evaporate. Estimates of evaporation on the basis of pan measurements and lake-level change range from 15 to 81 percent of available water (Reddell, 1965; Havens, 1966; Ward and Huddlestone, 1979; U.S. Bureau of Reclamation, 1982). Permeability of clay-rich soils across the floor of playa-lake basins may be very low, at least shortly after the lakes are filled with water (White and others, 1946; Harris and others, 1972; Knowles and others, 1984). Caliche that occurs throughout the Southern High Plains is a second possible barrier of low permeability beneath the lake sediments (Knowles and others, 1984).

However, lack of evaporites among playa-floor sediments (Harris and others, 1972), similarity of clay mineral assemblages between playa basins and adjacent areas (Allen and others, 1972), absence of halophytic flora (Reed, 1930), and low water salinities in playa lakes (Wells and others, 1970; Felthy and others, 1972; Lehman, 1972) suggest that playa-lake basins are at least partly flushed by deeply percolating vadose water and do not accumulate salts resulting from evaporation. Nativ and Riggio (1987) and Nativ and Smith (1987) determined that isotopic composition varies between rain, playa, and ground waters. Approximately 80 percent of playa- and ground-water samples have slightly heavier isotopic compositions than do rain waters, possibly reflecting a small evaporation effect that might develop either in playa-lake basins or in the soil zone in interplaya areas. Stone (1984) and Wood and

Osterkamp (1984) noted that dissolved solute concentrations in soil samples were significantly lower beneath the playas than beneath interplaya areas. This suggests either rapid rates of infiltration of playa water not affected by evaporation, or rapid drainage of vadose water beneath playa-lake basins, relative to rates of infiltration and subsurface flow in interplaya areas.

Other studies challenged the assumption that all playa basins are entirely underlain by impermeable sediments. Caliche is found beneath most playas but is sand-rich and relatively permeable beneath many, based on several hundred test holes drilled in the beds of playa lakes (White and others, 1946). Lotspeich and others (1971) found that caliche under playa lakes was partially dissolved and permitted leakage. Wood and Osterkamp (1984) concluded that recharge from a playa-lake basin occurs in an annular ring around the basin floor where soil is more permeable with karst-like solution pipes in the subsurface caliche. Desiccation cracks in playa-floor clays in conjunction with solution pipes in the underlying caliche may provide passageways for downward movement of water (Osterkamp and Wood, 1987).

Wood and Osterkamp (1987) observed that "few data exist on dissolved solids in the unsaturated zone adjacent to playa-lake basins to support interpretations of recharge processes." Despite numerous previous investigations, few data exist on hydrogeologic variability of playa-lake basins or distribution of vadose-water flow through the thick unsaturated section beneath playas across the High Plains. Hydrologic and geochemical cross sections of the unsaturated zone beneath playa and interplaya areas have not been made. Variability of recharge from playa lakes and of mass flux and vadose-water distribution in the vicinity of playa lakes have not been defined. Whether the proportion of recharge distributed between playa and interplaya areas has been reasonably constant during the Holocene Period is also unknown. The rate of vadose-water flow and percent of irrigation water applied during the 1900s that eventually returns to the aquifer through the thick vadose zone are unknown.

Recharge from irrigation might account for the rise in water-table elevation during the past several decades in the southern part of the Southern High Plains (Knowles and others, 1984).

Another source of recharge is local cross-formational flow. Recharge from underlying Cretaceous strata beneath the southern part of the Southern High Plains was inferred from similarities in salinity, hydrochemical facies, and stable-isotopic compositions, as well as from hydraulic-head differences (Nativ and Smith, 1987). Based on similar lines of evidence and the observation that Lower Dockum Group hydraulic head locally is higher than the water table in the Ogallala Formation, Nativ and Smith (1987) suggested that Ogallala water is recharged from the Dockum Group in western and southeastern Deaf Smith County and in Randall County. In these areas the Ogallala Formation is thin and hydraulic head of Dockum ground water locally is above the Ogallala water-table elevation. Ogallala Formation ground water along the Caprock Escarpment might be chemically modified by discharge of ground water from Permian rocks where the latter directly underlie the Ogallala Formation in Donley and Carson Counties (Nativ and Smith, 1987).

At present, most ground-water discharge from the aquifer in the Ogallala Formation is by well pumpage for irrigation and municipal water supply; as previously mentioned, overpumpage has caused significant decline in elevation of the water table. Before the pumping overdraft, Ogallala Formation ground water discharged in springs and seeps and in cross-formational flow into the underlying aquifer in the Dockum Group along the Eastern Caprock Escarpment. Springs have largely stopped flowing, indicating a change in configuration of the water table near the Caprock Escarpment (Brune, 1981). Amount of cross-formational flow downward into upper Dockum Group shale is poorly known; it is predicted from hydraulic-head relationships (Fink, 1963; Bassett and Bentley, 1982, 1983; Dutton and Simpkins, 1986; Wirojanagud, 1986; Senger and others, 1987a, 1987b) but poorly constrained because permeability data for upper Dockum shale and sandstone do not exist.

In the conceptualization of the shallow aquifer system used in a numerical model by INTERA (1984), no flow could cross the western boundary, which was assumed to be the western extent of the Southern High Plains. Recharge to Ogallala Formation ground water was assumed to occur solely through infiltration at a rate of 0.3 inch/yr (0.76 cm/yr), obtained by model calibration. INTERA (1984, Simulation B) estimated that discharge of Ogallala ground water in the Palo Duro Basin is distributed as follows: 38 percent to springs in Palo Duro Canyon, 29 percent across the southern boundary of the model (coinciding with the southern boundary of the Palo Duro Basin; see fig. 1), 4 percent out of the Eastern Caprock Escarpment, and 29 percent downward into the Dockum. In the INTERA (1986) model, a different western boundary, coinciding with the state border between Texas and New Mexico, allowed ground water into the modeled region. Recharge to Ogallala ground water was assumed to occur as infiltration (89 percent) and as inflow across the western boundary of the model (11 percent). Discharge was distributed as follows: 24 percent to the Palo Duro Canyon, 26 percent across the model's southern border, 17 percent across the model's northeastern border, and 36 percent to the Dockum Group.

Dockum Group Hydrogeology

Dutton and Simpkins (1986) determined that the potentiometric surface of unconfined and confined ground water in the Triassic Lower Dockum Group closely reflects topography (fig. 53). Ground-water-basin divides were located beneath the Caprock Escarpment along the northern and western margins of the Southern High Plains (Fink, 1963; Dutton and Simpkins, 1986). Beneath the Southern High Plains, the potentiometric surface indicates that ground-water flow direction is to the east or southeast. Hydraulic head of lower Dockum Group ground water is 300 to 700 ft (90 to 210 m) lower than the water table of the Ogallala aquifer beneath much of the

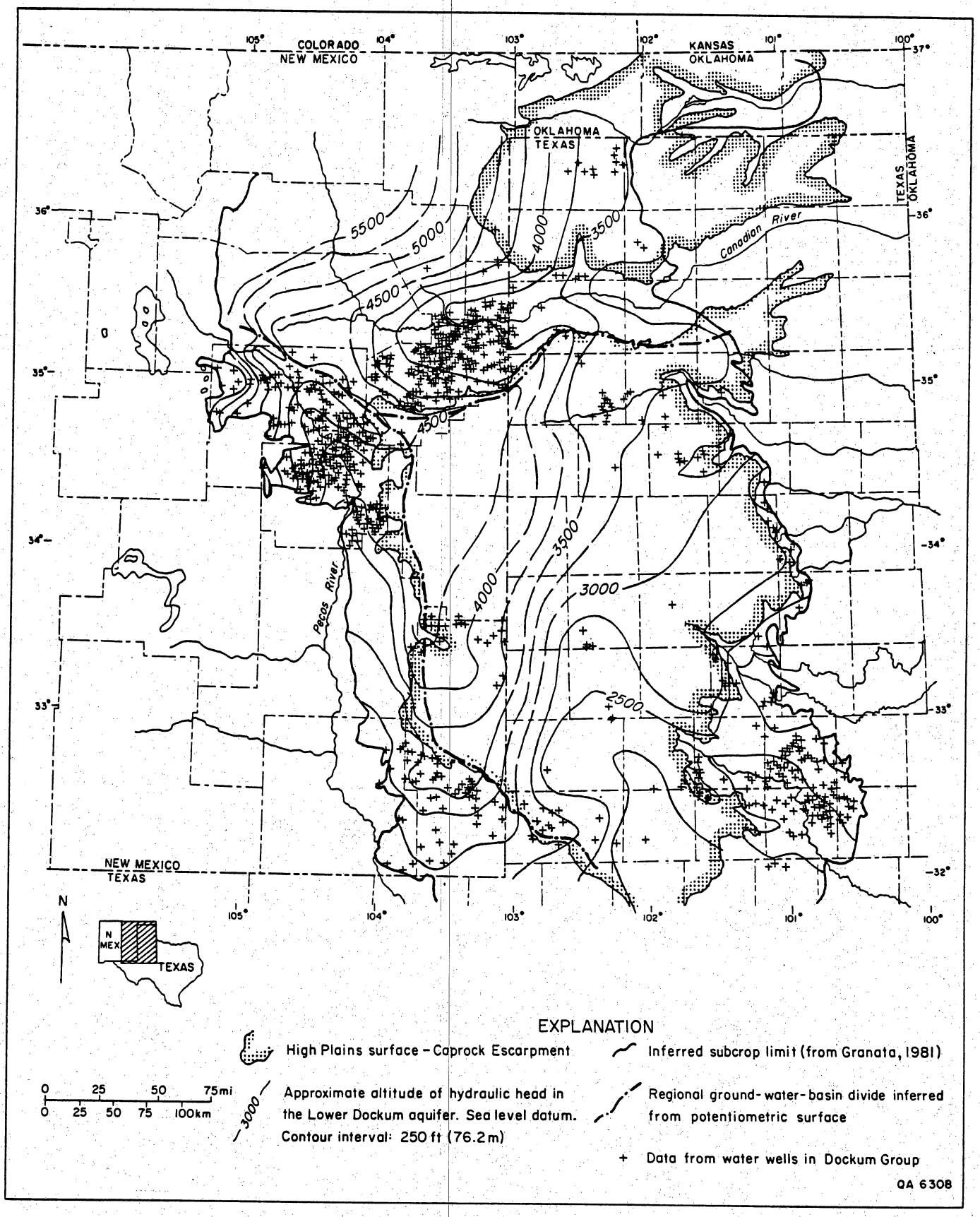


Figure 53. Potentiometric surface of lower Dockum Group ground water and ground-water-basin divides. From Dutton and Simpkins (1986). Map based on approximately 1,050 water-level measurements; dashed contours signify areas with little well control.

Southern High Plains (Fink, 1963; Dutton and Simpkins, 1986); the difference increases toward the southern part of the Southern High Plains. Dutton and Simpkins (1986) interpreted the magnitude of hydraulic-head difference as indicating that downward ground-water flow between the aquifers is restricted by thick, low-permeability shale in the upper Dockum and upper part of the lower Dockum Group. They did not quantify or model the amount of leakage to be expected from head gradient and did not determine whether ground-water basin divides could result from cross-formational flow focused for some reason along the western and northern limits of the Southern High Plains. The potentiometric surface of ground water in sandstones of the upper Dockum Group has not been mapped to date because few wells are completed solely in these thin and mostly discontinuous deposits. The potentiometric surface of ground water in the upper Dockum Group is probably intermediate between the elevation of the water table in the High Plains aquifer and the potentiometric surface of lower Dockum Group ground water.

Dutton and Simpkins (1986) hypothesized that the ground-water basin divides (fig. 53) prevent recharge in the Dockum Group outcrop in the Pecos River and Canadian River valleys from entering the confined aquifer beneath the Southern High Plains. If that is correct, Modern recharge of Lower Dockum Group ground water beneath the Southern High Plains is limited to cross-formational leakage. The ground-water divides most likely developed in conjunction with excavation of the Pecos River valley, which presumably took place before the Holocene, based on studies by Gustavson and Finley (1985) and Gustavson (1986a). Before thick lower Dockum Group sandstones were eroded from the area west of the present outcrop limit and before the ground-water-basin divide beneath the Western Caprock Escarpment was fully established, the rate of recharge, located at an elevation of 6,000 to 8,000 ft (~1,800 to ~2,500 m) in eastern New Mexico may have been significantly greater than effective recharge from precipitation across the Southern High Plains.

Dutton and Simpkins (1986) supported this hydrologic interpretation with geochemical data. Hydrochemical facies and salinity generally differs between ground waters in the Ogallala Formation and Lower Dockum Group. In general, Lower Dockum Group ground water is from two to more than ten times more saline than Ogallala Formation ground water. Dutton and Simpkins (1986) recognized three general associations of hydrochemical facies in Lower Dockum Group ground water. First, in the northern part of the Southern High Plains ground-water basin, TDS of Na–HCO_3, mixed-cation–HCO_3, and Na–mixed-anion hydrochemical facies is generally less than 3,000 mg/L (figs. 54 and 55). This good quality water occurs in the thicker deposits of Lower Dockum Group sandstone (Dutton and Simpkins, 1986). In this area, the Ca–HCO_3 hydrochemical facies is absent from the Dockum Group (fig. 54) but is common in the Ogallala Formation (fig. 52, area 6). Beneath western Deaf Smith County and northern Parmer County, ground waters in both the Ogallala Formation and Lower Dockum Group consist of Na–HCO_3 hydrochemical facies. Second, a Na–Cl facies having TDS of 10,000 to more than 20,000 mg/L underlies the central part of the Southern High Plains (fig. 55) where Dockum Group sandstones are thought to be thin and discontinuous. Third, Na–SO_4 and Na–mixed-anion waters having TDS from 1,000 to 5,000 mg/L characterize the southern part of the Southern High Plains; the dominance of the sulfate anion varies gradually across the area and these latter two facies are not distinct. Hydrochemical facies are variable in the Dockum Group outcrop areas in the Pecos and Canadian River valleys and east of the Eastern Caprock Escarpment.

The $\delta^{18}O$ and δD values of Dockum Group ground water plot along the meteoric water line but are 2 °/oo to 4 °/oo and 15 °/oo to 25 °/oo lighter, respectively, than those of Ogallala Formation ground water. Dutton and Simpkins (1986) concluded that the isotopic and hydrochemical differences support the hydrologic interpretation that (1) confining, low-permeability mudstone in the Upper Dockum Group

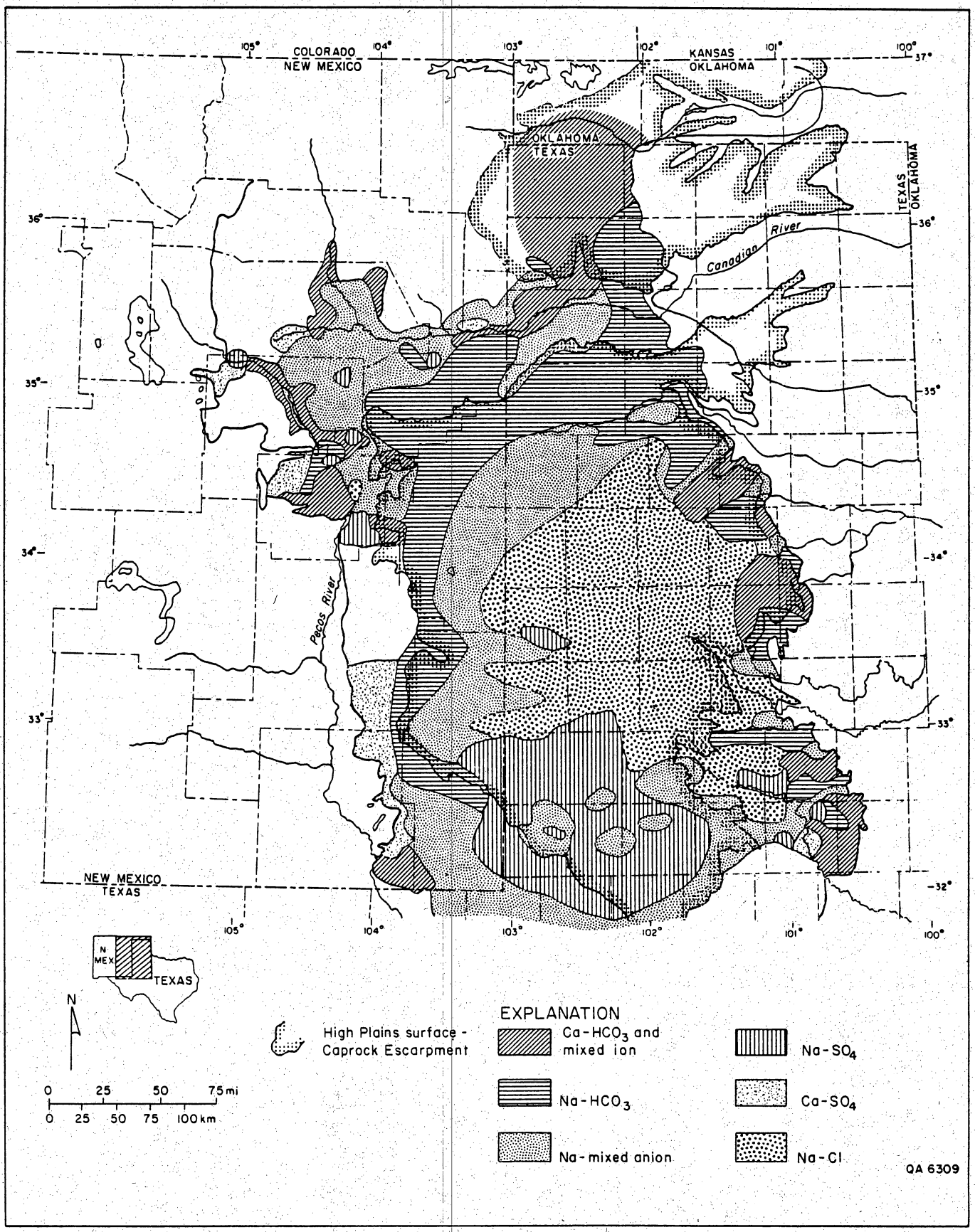


Figure 54. Hydrochemical facies of lower Dockum Group ground water. From Dutton and Simpkins (1986). Only key well data are located; additional well locations shown in Figure 55.

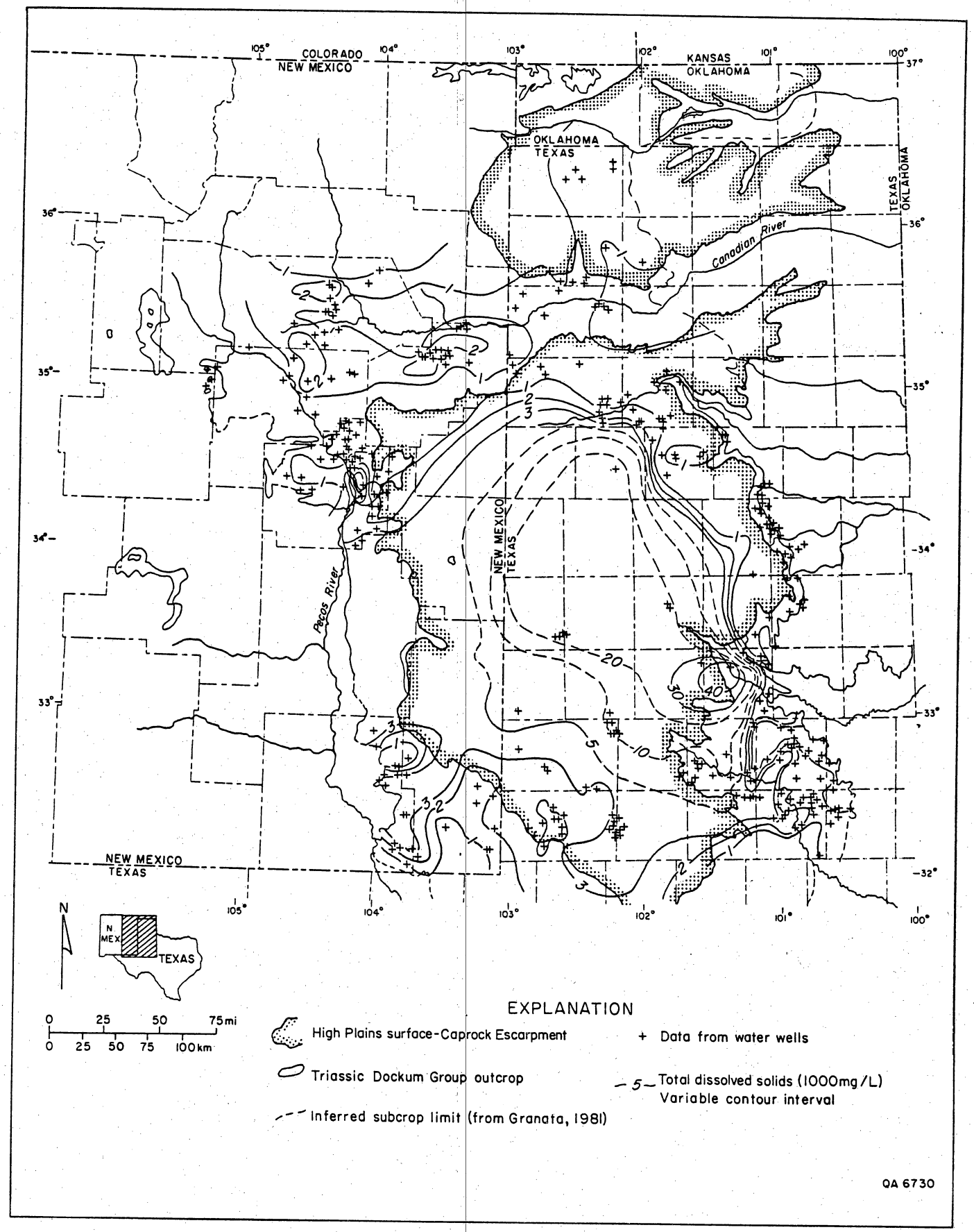


Figure 55. Total dissolved solids in lower Dockum Group ground water. From Dutton and Simpkins (1986).

and upper part of the Lower Dockum Group separate the two aquifer units and limit cross-formational recharge to a negligible amount and (2) ground water present in the confined Dockum aquifer was recharged by lateral flow from higher elevations to the west before the Pecos River valley was fully excavated. An alternate explanation is that the depleted isotopic values resulted from cross-formational recharge from the overlying Ogallala aquifer during a cooler and wetter climate than exists at present. However, in order to maintain the isotopic difference, this mechanism implies that the confining mudstones transmitted isotopically depleted water during a relatively brief interval of the Pleistocene, but retarded Holocene recharge; such a change in hydrologic behavior is improbable. The Dutton and Simpkins (1986) hypothesis has not been tested by flow modeling or by chemical analyses of radioactive isotopes from Dockum Group ground water.

The greater salinity and dissolved sulfate concentration in Dockum Group ground water beneath the central and southern parts of the Southern High Plains suggest that Dockum Group ground water in these areas might have been recharged by upward movement of ground water that had dissolved halite and anhydrite in Permian rocks (Dutton and Simpkins, 1986). The $\delta^{34}S$ values of dissolved sulfate of +10 °/oo to +12 °/oo which are typical of some Dockum Group ground waters, particularly in outcrop areas, might reflect dissolution of Permian anhydrite or gypsum; few samples for isotopic analysis to be used in determining the recharge source are available from areas where Dockum Group ground waters are saline. Dutton and Simpkins (1986) thought that diluted and modified connate water could exist in sandstone-poor parts of the Dockum Group for prolonged time periods due to inefficiencies of flushing by meteoric water.

Hydrochemical evidence of cross-formational discharge from the Ogallala Formation is ambiguous or nonexistent. Because chemical and isotopic composition of Dockum Group ground water generally differs from that of Ogallala Formation ground water,

either there is little downward discharge from the Ogallala Formation or composition of downward moving ground water is significantly changed by mineralogic reactions in the Upper Dockum; the latter explanation is unlikely because there is no reasonable geochemical reaction to deplete the abundance of heavy oxygen and hydrogen isotopes along a downward flow path. Dutton and Simpkins (1986) inferred from the potentiometric surface map and the spatial variability of hydrochemical facies that unconfined or partly confined Dockum Group ground water within the Pecos River and Canadian River valleys discharges to bottomlands and east of the Eastern Caprock Escarpment discharges into into near-surface aquifers or surface-water bodies. As in the High Plains aquifer, confined ground water in the Dockum Group beneath the Southern High Plains most likely discharges in springs and seeps, percolates downward into the underlying Upper Permian deposits along the Eastern Caprock Escarpment, or is pumped from wells (Dutton and Simpkins, 1986). Similarity of stable isotopic composition suggests that there is downward movement of ground water from both the Ogallala Formation and Dockum Group beneath the Caprock Escarpment into salt-dissolution zones (Dutton, 1987a); the vertical hydraulic-head gradient near the Caprock Escarpment is great enough to drive Ogallala Formation ground water downward through Dockum rocks into the upper Permian evaporite section. Near the western edge of the Southern High Plains and along reaches of Tierra Blanca Creek in Deaf Smith and Randall Counties, where the Ogallala Formation is thin due to erosion, hydraulic head in the Dockum Group locally may be higher than the water-table in the Ogallala Formation, indicating a potential for upward ground-water discharge from the Dockum Group (Nativ and Smith, 1987). Chemical composition of the two aquifers is more similar in these areas.

Ground water in the Dockum Group has been modeled as part of larger system models. An INTERA (1984) model assumed that infiltration of precipitation in Dockum outcrop areas directly recharges the confined aquifer beneath the Southern

High Plains. Fink (1963) and Dutton and Simpkins (1986), as previously mentioned, present data that suggest that this assumption is incorrect. INTERA (1984) calculated that 40 percent of the influx was from outcrop-area recharge and 59 percent was from vertical cross-formational leakage from the Ogallala. The numerical model framework is probably inaccurate in assigning too high a vertical permeability to the upper Dockum, inasmuch as Dockum and Ogallala water composition are quite distinct. The model calculates that 54 percent of discharge from the Dockum Group flows to the Canadian River and 46 percent discharges across the southern boundary of the model. Vertical leakage to the underlying evaporite section was not included in the conceptual model. In a model by Senger and Fogg (1987), a small percentage of the flow (approximately 0.2 percent in Simulation B-2) discharges downgradient into the evaporite confining system. INTERA (1986) calculates that 76 percent of recharge to a Dockum/Dewey Lake layer occurs as vertical leakage from the Ogallala Formation and 24 percent as inflow across the western boundary of the model. The latter model estimates that 47 percent of the flow discharges across the southern boundary of the model, 17 percent to the Canadian River, 36 percent to the Rolling Plains, and less than 0.1 percent to the underlying evaporite section.

Evaporite Confining System

Permeabilities are much lower in the Upper Permian evaporite section than in the overlying aquifers and make the Permian deposits of shale, carbonate rock, anhydrite, and halite function as a confining system in the Palo Duro Basin (table 6). Whether the confining system should be classified as an aquitard or an aquiclude, which reflects how much ground water actually leaks through the section, has not been resolved (Dutton, 1987b). Because potentiometric surfaces are higher in the overlying aquifers than in deeply buried aquifers in Wolfcampian and Pennsylvanian rocks

beneath the confining system in the Palo Duro Basin, there is a potential for downward ground-water flow across the evaporite confining system, as depicted in figure 50. The amount and mechanism of flow are controversial (Dutton, 1987b). Given this uncertainty, the more general term "confining system", which does not imply a given leakage rate, is used instead of "aquitard" in this section.

Equipotential cross-sections suggest that the steepest drop in hydraulic head might occur in the upper part of the Clear Fork Group within the confining system (Bair, 1987a; Senger and Fogg, 1987). However, estimation of flow velocity is difficult because water density varies greatly with changes in salinity. Also, the few available measurements of hydraulic head within the confining system are insufficient to map flow potential in detail. Hydrologic studies of the confining system (Orr and Dutton, 1983; Dutton and Orr, 1986; Dutton, 1987b) focused on the two hydrogeologic units in the evaporite confining system that appear to have higher permeabilities relative to the majority of the units in the evaporite confining system: (1) salt dissolution zones that exist around the perimeter of the Southern High Plains (Dutton, 1987a, in press) and (2) the informally named "unit 4 carbonate bed" (Fracasso and Hovorka, 1986), the thickest carbonate bed in the San Andres Formation in the Palo Duro Basin. In addition, a hydrologic test was conducted in the Queen and Grayburg Formations, which make up a permeable unit above the San Andres Formation. Fluid pressure of brine in that test zone (SWEC J. Friemel No. 1, zone 9) at an elevation of 2,392 ft (729.1 m) was 462 psi (3.2 MPa) (Stone and Webster Engineering Corporation, 1985). Fluid pressure in the underlying San Andres Formation (SWEC J. Friemel No. 1, zone 8) at an elevation of 1,207 ft (367.9 m) was 1,090 psi (7.52 MPa) (Dutton and Orr, 1986).

Salt-Dissolution Zones

The uppermost part of the evaporite confining system around the perimeter, and to a lesser extent, beneath the northern part, of the Southern High Plains contains extensive salt-dissolution zones where ground water has dissolved halite and anhydrite (see Gustavson, this volume). Large concentrations of dissolved salt in rivers draining the area, recent development of sinkholes and depressions (Johnson, 1981; Gustavson and others, 1982), and geochemical characteristics of salt springs and saline ground water (Richter and Kreitler, 1986) indicate that evaporite dissolution is a continuing hydrologic process. Johnson (1981), Gustavson and others (1980, 1982), and Simpkins and Fogg (1982) hypothesized that ground water flows downward from the overlying aquifers into the confining system, causing salt dissolution. Little is known about specific flow paths of ground water that contacts and dissolves bedded halite. The hydrogeologic properties of the Permian formations are, of course, changed by dissolution of bedded salt, leaching of halite and anhydrite cements from sandstone and carbonate rock, and by fracturing of rock collapsing into dissolution cavities. Flow paths from salt-dissolution zones to discharge areas at ground surface in topographically low areas are undoubtedly complex and include flow through joints in low-permeability formations, through fractures enlarged by solution and collapse, and locally through sinkholes, caverns, and breccia-filled chimneys (Johnson, 1981; Dutton, 1987a, in press). There are too few wells to map potentiometric surfaces of ground water in the salt-dissolution zones. However, Dutton (1987a, in press) assumed that the general shape of potentiometric surfaces in salt-dissolution zones reflects topography and might display ground-water-basin divides that are similar to those in the aquifer in the Dockum Group (fig. 56).

Dutton (1987a) collected hydrologic and geochemical data from six test wells to determine hydrogeochemical processes involved and the source and flow paths of

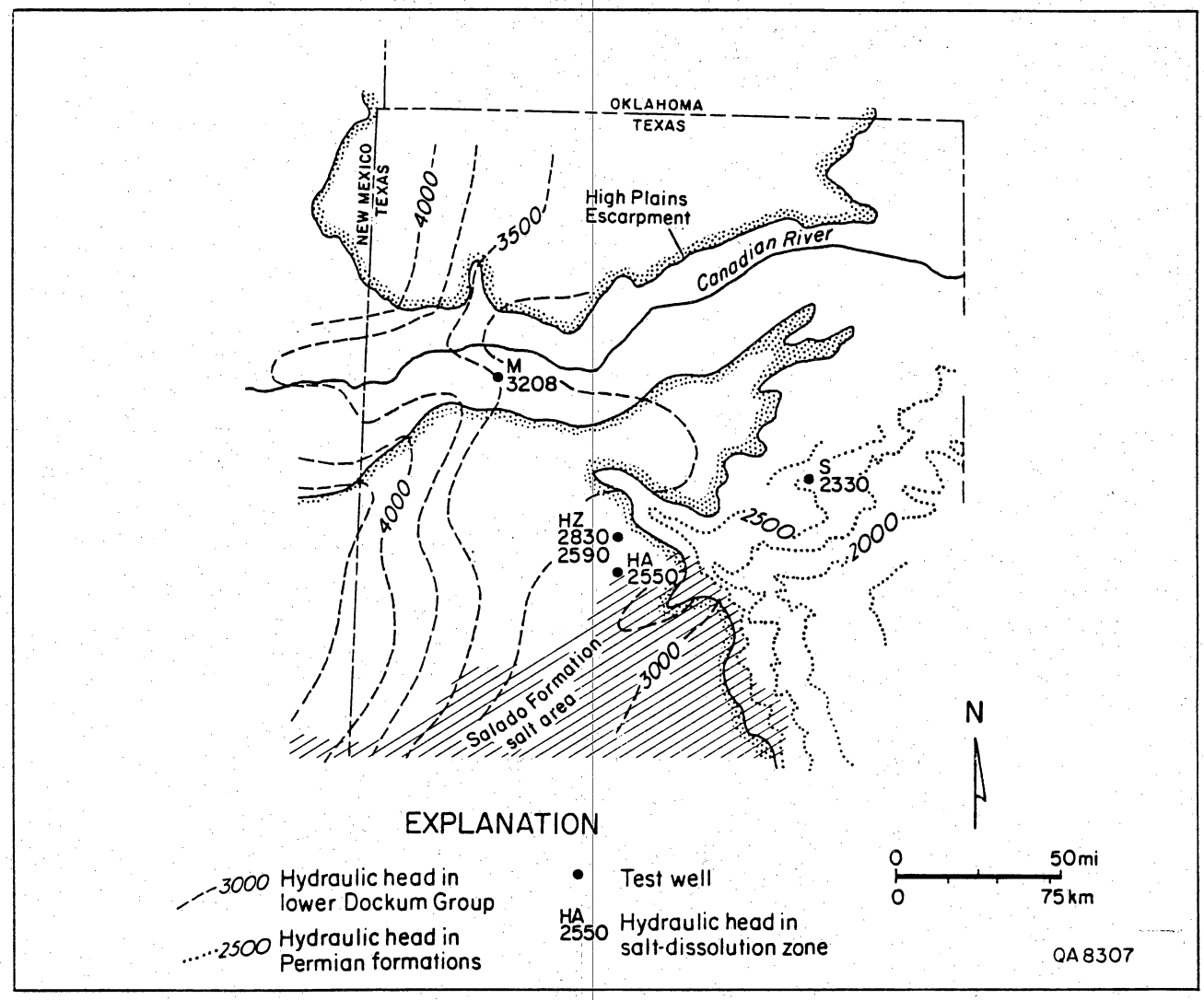


Figure 56. Hydraulic head of ground water in aquifers overlying salt-dissolution zones and point values of hydraulic heads in salt-dissolution zones. From Dutton (in press). Northern edge of Salado Formation salt area assumed to be the boundary of transmissive salt-dissolution zones.

ground water moving in salt-dissolution zones. Samples from salt-dissolution zones in the Artesia and Ochoan Groups in the Palo Duro Basin consist of Na–Cl hydrochemical facies having TDS values from approximately 65,000 to 288,000 mg/L (Dutton, 1987a, in press). The position of $\delta^{18}O$ and δD values along the meteoric water line, the Na/Cl (0.60 to 0.65) and Cl/Br (2,560 to 5,830) ratios, the $\delta^{34}S$ values (+8 °/oo to +12 °/oo), and the saturation indices with respect to gypsum (~0.0) and halite (-1.6 to -1.9) clearly indicate that salinity was derived by dissolution of gypsum and halite by meteoric water (Dutton, 1987a, in press). Values of $\delta^{18}O$, δD, and $\delta^{13}C$ in ground waters sampled from salt-dissolution zones in Artesia and Ochoa Groups are similar to those from the Lower Dockum Group and Ogallala Formation (Dutton, 1987a).

Hydraulic heads in salt-dissolution zones in the San Andres unit 4 carbonate bed in Donley County, in the Lower Seven Rivers Formation in Oldham and southern Randall Counties, and in Salado and Tansill Formations in Swisher and southern Randall Counties are lower than the hydraulic heads of the overlying aquifers and are within or slightly above the top of the Permian evaporite section (Dutton, 1987a). A potential exists for downward ground-water flow from overlying aquifers into salt-dissolution zones, regardless of differences in water density. Hydraulic head in the salt-dissolution zone in the Seven Rivers Formation appears lower than that in the overlying Salado and Tansill Formations, suggesting that the downward decrease in hydraulic head continues into the upper part of the evaporite confining system (Dutton, 1987a). Possible reasons for the continuous downward decrease in hydraulic head include: (1) elevations of recharge and discharge areas for salt-dissolution zones around the Southern High Plains are lower than those for overlying aquifers and limit the maximum hydraulic head in each hydrologic unit, and (2) modern discharge rates from salt-dissolution zones are greater than recharge rates (Dutton, 1987a). The latter implies that hydrologic conditions in salt-dissolution zones might be in a transient state.

The isotopic similarity and the downward decrease in hydraulic head through the Dockum Group and salt-dissolution zones confirm that the meteoric ground water in salt-dissolution zones originated as cross-formational recharge from the shallow aquifer system. Ground-water flow paths through Upper Permian rocks must have changed as salt-dissolution zones expanded and effected physiographic changes (Gustavson and Finley, 1985).

Dutton (1987a, in press) hypothesized that salt-dissolution rates decreased from the Tertiary and Pleistocene to the Holocene because of regional changes in physiography that probably decreased the amount of recharge to salt-dissolution zones; whether climatic change also influenced recharge rates is unknown. His qualitative results suggested that present as well as paleohydrologic ground-water velocities and salt-dissolution rates are probably less beneath the Southern High Plains ("interior" salt-dissolution zones) than in adjacent, "peripheral" salt-dissolution zones because of lower hydraulic conductivities and lower hydraulic-head gradients. The comparison of interior and peripheral salt-dissolution zones appeared supported by data from test wells: salinities in peripheral salt-dissolution zones are low (67,000 to 95,000 mg/L) despite high solubility of halite, reflecting relatively open circulation of ground water, whereas in interior salt-dissolution zones beneath the Southern High Plains, ground-water circulation is low and water composition tends to reach halite saturation (Dutton, 1987a).

In the models of Simpkins and Fogg (1982) and Senger and Fogg (1987), flow in salt-dissolution zones originates from the shallow aquifer system and travels predominantly horizontally through the dissolution zones. Recharge and discharge rates are not specified. Simulated flow patterns (Senger and Fogg, 1987) suggest that recharge that occurs west of the Southern High Plains discharges to the Pecos River and cycles through the upper part of the evaporite confining system. Atwood and others (1986) refer to salt-dissolution zones below the Pecos River valley in New

Mexico and the Red River valley in Texas. The salt-dissolution zone in the vicinity of the Pecos River was assigned a ratio of horizontal to vertical hydraulic conductivities (k_h/k_v) of 2 (in all other units, k_h/k_v is 10). A local flow system is evident in the vicinity of the Pecos River; the modeled salt-dissolution zone in the vicinity of the Red River had no discernible effect on local flow.

INTERA (1986) included a salt-dissolution zone in one model simulation to investigate the possible effects of salt dissolution on flow in the San Andres unit 4 carbonate bed. The salt-dissolution zone was represented by increasing horizontal and vertical hydraulic conductivities in specified areas of the upper evaporite layer. Model results indicated that presence of a salt-dissolution zone influences hydraulic-head values in the underlying San Andres unit 4 carbonate bed.

Lower San Andres Formation--Unit 4 Carbonate Bed

The San Andres Formation includes carbonates, evaporites, and terrigenous clastics at approximately the center of the evaporite confining system. The unit 4 carbonate bed, which directly underlies the unit 4 halite, a possible repository horizon (U.S. Department of Energy, 1986), is stratigraphically equivalent to the P-3 porosity zone in the San Andres Formation in eastern New Mexico and to the Yellowhouse dolomite, the reservoir in the Yellowhouse oil field in the Northern Shelf of the Midland Basin (Ramondetta, 1982a). The halite- and anhydrite-cemented carbonate rock contains brine and has an average intrinsic permeability of 0.24 md (2.4×10^{-16} m^2) (Dutton and Orr, 1986). Pressures of ground water in the unit 4 carbonate bed at depths of 2,592 to 3,041 ft (790 to 927 m) in the Palo Duro Basin are nearly hydrostatic at 980 to 1,310 psi (6.76 to 9.03 MPa) (Dutton and Orr, 1986). Based on a map of equivalent-fresh-water hydraulic head (fig. 57), Dutton and Orr (1986) suggested that a ground-water-basin divide lies beneath the western border

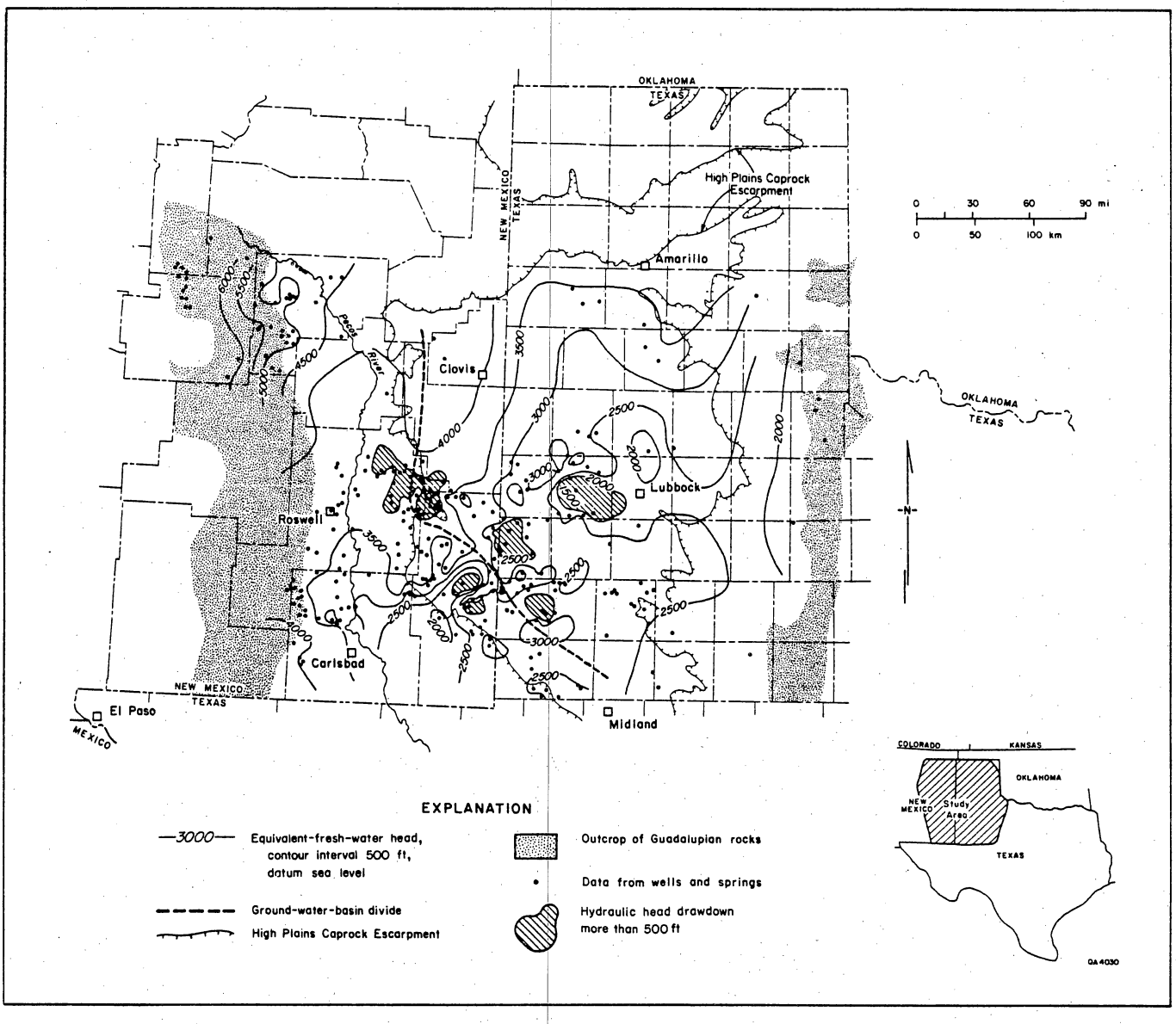


Figure 57. Potentiometric surface of ground water in the San Andres Formation prepared by hand contouring equivalent-fresh-water hydraulic head data from drill-stem tests and water-level measurements. From Dutton and Orr (1986).

of the Southern High Plains, in a position similar to the previously described divide in the Dockum hydrogeologic unit, and that the divide separates San Andres ground water present beneath the Southern High Plains from that in San Andres rocks in the Pecos Plains of Eastern New Mexico. This indicates that the Pecos River captures some or all of the recharge to Permian rock aquifers in the Pecos Plains, as hypothesized also by Summers (1981) and represented in modeling results (Senger and Fogg, 1987). The accuracy of this interpretation might be affected by use of fresh-water hydraulic head to predict flow potential in variable-density ground water. Potentiometric levels correctly indicate directions of ground-water flow only where water density and hydraulic conductivity are reasonably constant. Beneath the Southern High Plains, the interpreted flow potential within the San Andres unit 4 carbonate is variable, roughly reflecting the prevalent southeastward dip of land surface, but also influenced by withdrawal of billions of barrels of hydrocarbons and ground water from San Andres oil fields in the Permian Basin.

As previously mentioned, Dutton and Orr (1986) concluded that if recharge from rain and snowmelt to the San Andres aquifer in the Pecos Plains of eastern New Mexico discharges to the Pecos River and is prevented from recharging San Andres carbonates below the Southern High Plains owing to a ground-water divide. There are three remaining sources for ground water now found in the San Andres Formation: (1) intrastratal recharge from the west before the Pecos River valley was incised, (2) vertical movement from adjacent formations, or (3) preservation of formation water of Permian origin (Dutton and Orr, 1986; Dutton, 1987b).

The low permeability of halite- and anhydrite-cemented carbonate rock and small lateral gradients in hydraulic head probably limited specific lateral discharge to less than approximately 1 mi/m.y. (~1.6 km/m.y.) (Dutton and Orr, 1986) and minimized impact of intrastratal recharge on San Andres ground-water composition in the Palo Duro Basin. Significant potential for intrastratal recharge might have existed only

after Miocene uplift of the western part of the Palo Duro Basin until the Pleistocene when the Pecos River valley was excavated. Hydrologic modeling results (Senger and others, 1987a, 1987b) and existence of a potentiometric gradient suggest that vertical recharge might occur through the overlying part of the confining system, depending on permeability of the evaporite section and whether flow in fractures is significant at depth. The fact that there is a vertical pressure gradient through the evaporite confining system and that San Andres fluid pressures recover (although slowly) from pumping tests (Dutton and Orr, 1986, app. A) suggest that ground water in the confining system is not stagnant but rather capable of motion under hydrodynamic forces. If ground water is in motion, it is probable that any connate water present in San Andres carbonate rocks has been modified by circulation of meteoric water. However, the rate, distribution, and duration of ground-water flow and amount of connate-water modification are poorly known. White (1965) defined connate water as generally similar in age to, or somewhat younger than, the age of its associated rocks. Meteoric water refers to water that recently precipitated as rain or snow and that is young compared to the age of associated rocks.

Hydrochemical facies and salinity of San Andres Formation ground water in eastern New Mexico and the Texas Panhandle are related to rock type and position in the flow system (Dutton and Orr, 1986; Dutton, 1987b). Ca–HCO_3 water having TDS less than approximately 1,000 mg/L occurs in limestones and dolostones at the western limit of the Pecos Plains in eastern New Mexico, whereas Ca–SO_4 water with TDS from 1,000 to 10,000 mg/L exists beneath most of the Pecos Plains west of the Pecos River. The Ca–SO_4 facies reflects the occurrence of gypsum interbedded with carbonate rocks in the San Andres Formation. East of the Pecos River, Na–Cl, Ca–Cl, and mixed-cation–Cl waters with TDS from 10,000 to more than 300,000 mg/L prevail; salinity increases from the Midland Basin in the south to the Palo Duro Basin in the north and appears to be related to the abundance of halite in

the San Andres Formation and the overlying part of the evaporite confining system. Na/Ca and Na/Cl ratios vary significantly among brines from different parts of the Midland and Palo Duro Basins; the highest salinity water tends to have smaller Na/Cl and higher Ca/Cl ratios.

Dutton and Orr (1986) and Dutton (1987b) stated that explanation of the origin of chemical and isotopic composition of San Andres Formation brine must address (1) $\delta^{18}O$ and δD compositions of brines from two test wells (SWEC Zeeck No. 1, zone 4; SWEC J. Friemel No. 1, zone 8), (2) small Na/Cl and Cl/Br ratios, and (3) continuous variation in ionic ratios and salinity between the Palo Duro and Midland Basins. Dutton and Orr (1986) and Dutton (1987b) concluded that enriched hydrogen and oxygen isotopic compositions and the concentrations and ratios of major ions are consistent with an origin as a modified connate brine, which implies, however, that flow rates are negligible, as in an evaporite aquiclude. An alternate explanation is that these brines derived their chemical and isotopic composition by extensive water-rock interactions, which implies that flow rates are relatively significant, as in an aquitard (Dutton, 1987b). The extensive reactions are more complex and less substantiated for the Palo Duro Basin geochemical system than are diagenetic modifications of connate brine (Dutton and Orr, 1986; Dutton, 1987b). Also, San Andres formation brine from the two test wells in the Palo Duro Basin is isotopically similar to brines from fluid inclusions within San Andres halite beds (Knauth and Beeunas, 1986), which suggests the two solutions have a related origin (Dutton, 1987b). Roedder (1984) presents textural and geochemical evidence that indicates that the fluid inclusions are trapped connate fluids from the Permian evaporating sea from which the salts were precipitated. The San Andres formation brine in carbonate rocks, of course, has ionic ratios different from those in fluid inclusions because of diagenetic reactions, principally dolomitization.

Dutton (1987b) stated that ground water in the low-permeability San Andres carbonate bed in the Palo Duro Basin might be dominated by connate brine if (1) there has not been enough time for flow of meteoric ground water to flush connate brine from the carbonate bed since a significant cross-formational gradient in hydraulic head developed (Senger and others, 1987a, 1987b), (2) present cross-formational flow of ground water is unevenly distributed between fractured and unfractured areas, and (3) composition of brine sampled at the test wells differs from that of brine in fractured zones that have not been sampled. If fractures are the only pathway for movement of meteoric ground water across the evaporite confining system (Kreitler and others, 1985), ground-water flow would be unevenly distributed with a small or perhaps negligible amount moving across unfractured areas that were sampled at two test wells in the Palo Duro Basin (Dutton and Orr, 1986).

The San Andres unit 4 carbonate bed was a separate layer in the model of INTERA (1986) and Senger and Fogg (1987). Most of the recharge to that bed was limited to leakage from overlying units because of the western, northern, and southern boundaries of the unit were specified as "no-flow" boundaries (INTERA, 1986). Model results indicated that a small amount of recharge enters as upward-directed flow from the Wolfcamp layer in the southeastern corner of the Palo Duro Basin. Senger and Fogg (1987) and Senger and others (1987a) varied hydraulic-conductivity values of the San Andres unit 4 carbonate bed relative to average hydraulic conductivity of the remainder of the evaporite confining system in a numerical model to determine the sensitivity of model results to uncertainty in hydrogeologic properties. They found that to achieve significant lateral flow within the carbonate bed, the k_h/k_v ratio must exceed 10^3; data are inadequate to evaluate this finding. Also, as indicated by Dutton and Orr (1986), significant lateral flow is considered more likely in the Midland Basin than in the Palo Duro Basin. Discharge from the carbonate bed is simulated as leakage to underlying units beneath most of the area and as

surface discharge in the outcrop at the eastern side of the modeled area. Other aspects of modeling flow through the confining system are discussed in the following section along with models of the deep-basin brine aquifer system.

Deep-Basin Brine Aquifer

The potentiometric surface of ground water in Wolfcampian and Pennsylvanian rocks has a principal easterly to east-southeasterly dip across the Midland Basin from higher elevations in eastern New Mexico to lower elevations across north-central Texas and south-central Oklahoma, and a northeasterly dip in the Palo Duro Basin (McNeal, 1965; Handford, 1980b; Bentley, 1981; Bassett and Bentley, 1982, 1983; Smith and others, 1985). As for most hydrogeologic units beneath the Texas Panhandle and eastern New Mexico, the orientation reflects a primary topographic control of subsurface fluid pressures, as theorized by Tóth (1962, 1963, 1978). In addition, high-transmissivity arkosic ("granite-wash") sandstones along the southern flank of the Amarillo Uplift affect fluid pressures and flow paths. Most models have assumed that basement rocks of the Amarillo Uplift form an impermeable barrier along the northern side of the Palo Duro Basin. There is, however, some oil and gas produced from basement rocks in the Amarillo Uplift, indicating a finite permeability. Senger and others (1985) and Wirojanagud and others (1986) considered that Wolfcampian dolomite overlying the uplift may hydrologically connect the Palo Duro and Anadarko Basins and allow some interbasin flow. The potentiometric surface map suggests that the Pennsylvanian and Wolfcampian fan-delta sandstone deposits act as a hydrologic sink that draws ground-water flow paths to the north and northeast (Wirojanagud and others, 1986; Senger and others, 1987a, 1987b); the region of influence might extend as far south as the Midland Basin (fig. 58). The high-transmissivity sandstones (Conti and others, 1988) probably cause the apparent underpressuring of ground water

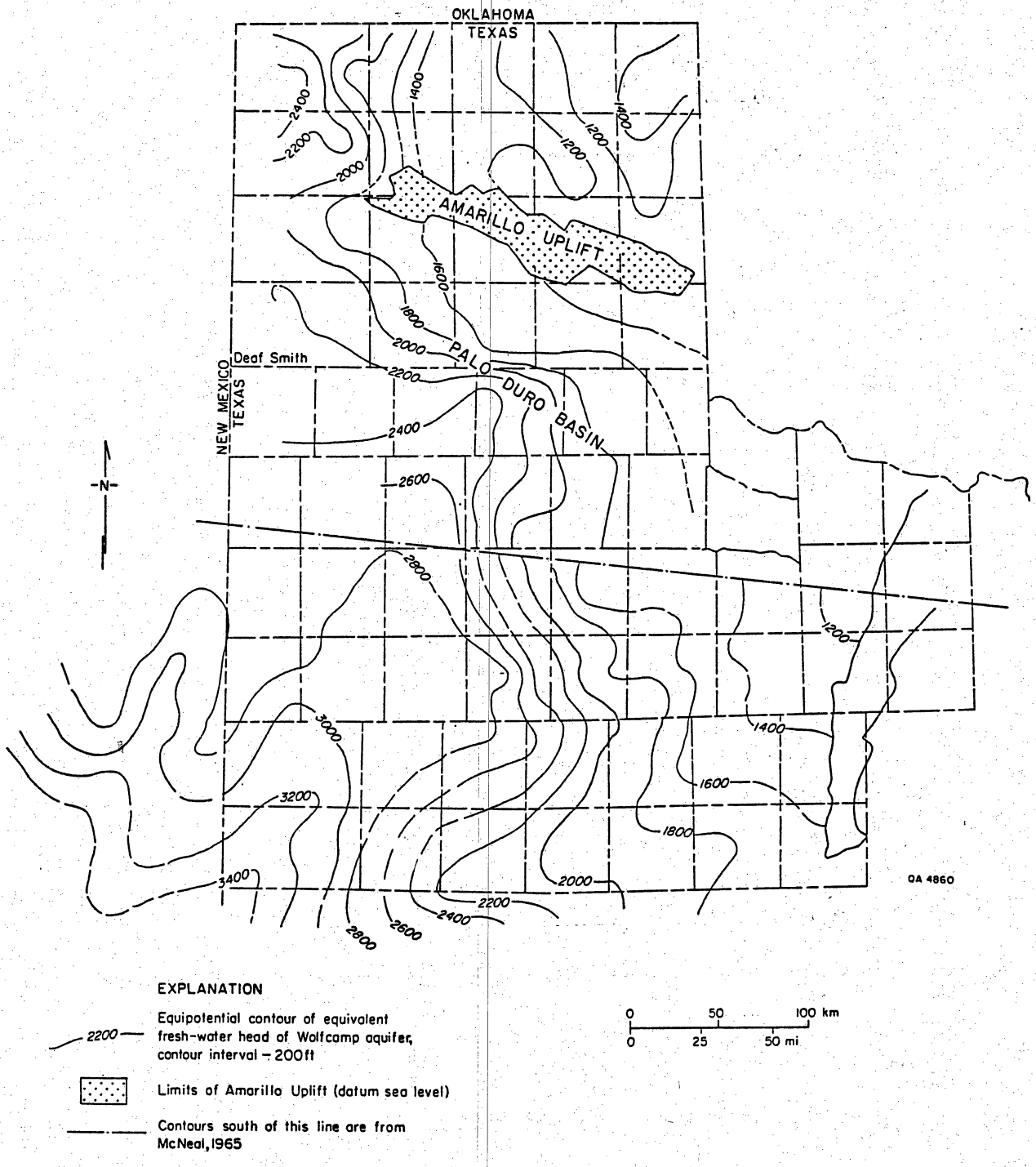


Figure 58. Potentiometric surface of ground water in Wolfcampian carbonate rocks based on geostatistical analysis of equivalent-fresh-water hydraulic head data from drill-stem tests. From Smith and others (1985).

in Wolfcampian rocks (Senger and others, 1987a, 1987b). Thus, potentiometric surface maps limited to the Palo Duro Basin (Handford, 1980b; Bassett and Bentley, 1982, 1983; Bair and others, 1985; Bair, 1987b) display an incomplete picture of regional flow patterns in Wolfcampian rocks.

There are three possible sources of ground water currently found in the deep-basin brine aquifer in the Palo Duro and Midland Basins: (1) intrastratal flow (recharge) from eastern New Mexico, (2) preservation of connate brine, and (3) vertical cross-formational leakage across the evaporite confining system (Fisher and Kreitler, 1987a, 1987b). Evaluation of these alternatives is largely based on geochemical evidence because confidence in ground-water modeling is limited by uncertainty in assumptions of steady state flow, porous or fractured hydraulic properties, leakage through confining beds, and homogeneity and isotropy of hydraulic properties (Dutton, 1988); hydraulic-head data alone cannot distinguish between the alternative explanations of origin of water in the basin.

Ground water within Wolfcampian rocks is thought to have been recharged in eastern New Mexico where the rocks subcrop beneath Leonardian and Guadalupian sandstones and carbonates, as along the eastern flank of the Pedernal Uplift (Fisher and Kreitler, 1987a, 1987b; Senger and others, 1987a, 1987b). However, there are insufficient wells completed in Wolfcampian rocks in the vicinity of the Western Caprock Escarpment and Pecos River west of the Palo Duro Basin to empirically evaluate whether the topographically controlled ground-water-basin divide interpreted from potentiometric maps of ground water in the Permian San Andres Formation and Triassic Dockum Group also extends at present into the brine aquifer in Wolfcampian rocks. The presence of such a divide would impact calculations of Holocene and future ground-water flow rates in the Palo Duro Basin. Results of numerical modeling do not show such a hydrologic divide in the deep-basin brine aquifer system (Senger and others, 1987a, 1987b). The model of Atwood and others (1986) indicates that

influence of the Pecos River does not extend deeper than the Queen and Grayburg Formations (Atwood and others, 1986). Atwood and others (1986) calculate that the amount of ground water captured by the Pecos River is less than the amount calculated in the model of Senger and Fogg (1987) because the former assigned lower permeabilities to the evaporite confining system west of the Pecos River. The potentiometric surface is not obviously influenced by topographic relief across the Red River valley (Atwood and others, 1986) or across the Eastern Caprock Escarpment, although retreat of the escarpment during the Cenozoic probably caused fluid pressure in the deep-basin brine aquifer to decrease (Senger and others, 1987a, 1987b).

Based on high Cl/Br ratios and the fact that the isotopic composition of hydrogen and oxygen are apparently nonequilibrated with the local carbonate rock, brines from Wolfcampian and Pennsylvanian rocks at the western side of the basin have been interpreted to be derived from intrastratal recharge of meteoric water that moved into the basin through clastic formations in eastern New Mexico and that acquired salinity by halite dissolution at the western margin of the basin (Fisher and Kreitler, 1987a, 1987b; Bein and others, in preparation). Brines from the SWEC Zeeck No. 1 well, SWEC Sawyer No. 1 well, and other deep oil wells in the center and at the eastern side of the basin have low Cl/Br ratios and oxygen isotopic compositions that appear to be equilibrated with carbonate host rock at *in situ* temperatures. These hydrochemical characteristics could reflect either a connate brine evolved from Permian seawater by diagenetic reactions or meteoric water that was recharged and became enriched in bromide because of halite recrystallization along a northeastward flow path from the Midland Basin (Fisher and Kreitler, 1987a, 1987b; Bein and others, in preparation). The different interpretations concerning source of water in the center and at the eastern side of the basin are significant because they imply greatly different flow rates and would reflect greatly different heterogeneity and other hydrogeologic properties in the basin.

Zaikowski and others (1987) interpret analyses of noble gases in some deep-basin brines to suggest ground-water residence times of 100 to 300 m.y. These must be considered preliminary estimates, however, because the composition of the host rock has not been adequately characterized and the sensitivity of the calculated ages to uncertainties in rock and water composition, contamination during sampling, and the possibility of extraformational sources of noble gases such as flux from basement rocks has not been critically evaluated (Fisher and Kreitler, 1987a, 1987b).

The hypothesis of uniform (that is, not fracture-controlled) cross-formational recharge across the evaporite confining system is inconsistent with major ion composition, ionic ratios, and stable isotopic composition of deep-basin brines (Fisher and Kreitler, 1987a, 1987b). The possibility that deep-basin brines are derived from flow across the evaporite confining system through fractured zones, however, has not been excluded. Ground-water samples from fractured zones within the confining system are needed for comparison with samples from unfractured areas. Extensive cross-formational flow, on the scale suggested by numerical models (30 to 50 percent of deep-basin brine aquifer discharge [Kreitler and others, 1985]), would significantly alter brine composition and brine in Wolfcampian rocks should show a clear hydrochemical imprint from cross-formational recharge, including high salinity approaching halite saturation in the horizons closest to the confining system. Also, Wolfcampian formation fluids should be more saline than underlying Pennsylvanian formation fluids (Fisher and Kreitler, 1987a, 1987b). In all test wells in which more than one sample was taken, however, samples near the evaporite section have lower salinity and lower halite saturation index than deeper brines (Fisher and Kreitler, 1987a, 1987b). Maps of regional trends in salinity and ionic ratios also show stratification in water composition in the hydrogeologic units, with Wolfcampian formations having lower salinity than deeper units.

To achieve the observed isotopic composition of deep-basin brines from ground water leaking downward through the evaporite confining system, brine in the confining system, assumed to have a composition similar to that in the San Andres Formation (Dutton, 1987b), would have to gain isotopically light oxygen (Fisher and Kreitler, 1987a, 1987b). The proportions of dissolution zone fluid and San Andres unit 4 fluid required to produce deep-basin brine compositions is different for each of the ionic species considered, thus the chemical evidence does not indicate that cross-formational flow is significant (Fisher and Kreitler, 1987a, 1987b). Cross-formational flow, however, may be highly localized along fault planes or fracture zones, giving local, undetected anomalies in water composition.

Samples from Wolfcampian and Pennsylvanian hydrogeologic units beneath the evaporite confining system are Na–Cl and Ca-Cl brines that contain 140,000 to 290,000 mg/L total dissolved solids (Kreitler and others, 1985; Fisher and Kreitler, 1987a, 1987b; Bein and others, in preparation). Chemical and isotopic composition and salinity of brine in Pennsylvanian rocks differ from brine composition in overlying Wolfcampian rocks. These differences could indicate either that (1) brine in the deep-basin aquifer system is density-stratified within a vertically continuous aquifer system or (2) ground-water flow rate in the Wolfcampian unit is greater than that in the Pennsylvanian unit and flow has more completely flushed connate brine out of Wolfcampian rocks (Fisher and Kreitler, 1987a, 1987b). Analysis of pressure distribution with depth in the deep-basin brine units indicates that the potential for downward flow from Wolfcampian to Pennsylvanian rocks exists only southwest of the Palo Duro Basin (Orr and others, 1985). In the northeast and northwest parts of the basin, there is a small component of ground-water flow directed upward.

The age and rate of movement of brine in Wolfcampian and Pennsylvanian rocks are controversial. Interpretation of noble gas contents suggests that residence time is 100 to 300 million years (Zaikowski and others, 1987). Bein and others (in

preparation) suggested that Pennsylvanian-formation brines are derived from lower Paleozoic seawater and evaporatively concentrated Permian brine. Results from numerical model calculations are equivocal. Atwood and others (1986) report estimated flow velocities and travel times from recharge areas in eastern New Mexico to presumed discharge areas in Oklahoma; in Wolfcampian rocks average velocity was calculated as 2.84 mi/m.y. (4.57 km/m.y.) and total basin travel time was approximately 124 m.y., whereas in Pennsylvanian rocks average velocity ranged from 7.38 to 10.6 mi/m.y. (11.88 to 17.06 km/m.y.) and travel time ranged from 42 to 57 m.y. Although these rates are consistent with results of age estimates based on noble-gas data (Zaikowski and others, 1987), other geochemical data suggests flow rate in Wolfcampian rocks is greater than in Pennsylvanian rocks. Conti and others (1988), Senger and Fogg (1987), and Senger others (1987a, 1987b) suggest that residence time of ground water is only several million years. Discrepancies probably are due to the influence of fan-delta sandstone deposits included in the various cross-sectional numerical models and definition of hydrostratigraphic units. Steady-state models such as used by Atwood and others (1986) and Senger and Fogg (1987) calculate inherently unrealistic values of travel time because tectonic and hydrogeologic setting likely change within the elapsed time and significantly affect ground-water flow (Senger and others, 1987a, 1987b).

As previously mentioned, potentiometric surfaces of all hydrogeologic units are inclined generally eastward roughly parallel to topography, with local variation affected by highly transmissive deposits along the Amarillo Uplift, and can be traced through the basin to formation outcrops. This suggests that there is a potential for discharge of the deep basin brine to the east of the Palo Duro Basin and other parts of the Permian Basin. Differences in Na/Cl, Cl/Br, and I/Cl ionic ratios and differences in stable isotopic composition have been cited by various authors as distinctive brine signatures reflecting deep-basin brine discharge. Richter and Kreitler (1986)

distinguished between (1) saline-water in springs at the northern part of the Rolling Plains in the Texas Panhandle that derived from circulation of meteoric water through shallow salt-dissolution zones, such as those studied by Dutton (1987a), and (2) saline water sampled from test wells at the southern part of the Rolling Plains that is discharged from regional, deeply circulating flow paths in the Permian Basin. TDS content of ground waters just tens of miles to the west of Paleozoic formation outcrops in North Texas varies from 50,000 to more than 150,000 mg/L (Core Laboratories, 1972a), indicating that there is intrastratal movement of brine from the basin to near land surface driven by regional potentiometric gradients (Richter and Kreitler, 1987b; Dutton and others, in press).

The principal discharge area for brine in the Palo Duro Basin is thought to be in south-central Oklahoma where Paleozoic formations crop out. The potentiometric surfaces of Wolfcampian and other hydrogeologic units are continuous from west to east across the Midland and southern part of the Palo Duro Basins, extending to formation outcrops in north-central Texas (McNeal, 1965). It is reasonable to expect that brine continues to flow generally eastward in Wolfcampian rocks eastward to where the rocks crop out where it probably mixes with local, shallowly circulating ground water (Dutton and others, in press). Ground water in Wolfcampian rocks from the Palo Duro Basin might also discharge into the Anadarko Basin through permeable carbonate rock that covers the Amarillo Uplift, based on results of numerical modeling (Senger and others, 1985; Wirojanagud and others, 1986) and studies of lithologic continuity in Wolfcampian rocks across the uplift (Conti and others, 1988).

Most of the flow through the regional ground-water flow system beneath the Southern Great Plains (fig. 49; Jorgensen and others, in press) occurs in shallow aquifers in the Ogallala Formation and Dockum Group. In all of the numerical models, specific discharge through the deep-basin brine aquifer is two orders of magnitude less than in the shallow aquifer system. Senger and Fogg (1987) calculat-

ed that only one percent of the total flow passes through the deep-basin brine aquifer. INTERA (1984, 1986) estimated that flow through the deep-basin brine aquifer makes up 0.2 percent to 0.35 percent of the total flow through the hydrologic systems.

Ground-water flow in the deep-basin brine aquifer system, including hydrogeologic units in Wolfcampian and Pennsylvanian rocks, has been modeled with the assumption that the aquifer system is recharged by cross-formational flow at the subcrop of Guadalupian and Leonardian rocks in the Pecos Plains of eastern New Mexico and by leakage through the evaporite confining system. Ground-water residence time in the aquifer system estimated from modeling varies with flow path and assumed hydrologic properties in the models; all model results suggest that several storage times (mean duration that a unit volume of water remains in the hydrogeologic unit) have elapsed. Calculated flow across the evaporite confining system makes up between 9 to 53 percent of the flow through the deep-basin brine aquifer. The wide range reflects uncertainty in conceptual models and scarcity of hydraulic-conductivity and hydraulic-head data. Darcy velocities through the evaporite confining system are two to three orders of magnitude less than those in the deep-basin brine aquifer and four to five orders of magnitude less than those in the shallow aquifer system. Darcy velocities in the evaporite confining system are essentially proportional to the vertical permeability (k_v) of the salt, because flow is nearly vertical across these units. INTERA (1986, p. 100) found that Darcy velocities in the deep-basin brine aquifer are relatively insensitive to changes in salt permeability: horizontal Darcy velocities in the deep-basin brine aquifer varied by only a factor of 2 as salt permeabilities were varied by up to 5 orders of magnitude.

In the model by Senger and Fogg (1987), approximately 26 percent of the flow through the deep-basin brine aquifer in Simulation A-2 was contributed by vertical

leakage through evaporite confining system. The remainder is contributed by surface infiltration in outcrops in the west. Results of simulations show that much of the water that recharges shallow aquifers beneath the Pecos Plains discharges to the Pecos River and does not flow beneath the Southern High Plains. Flow rate is locally high west of the Pecos River because of the relatively high permeability assigned to the mudflat and alluvial/fan delta system (Senger and Fogg, 1987). Ground water within the deep-basin brine aquifer moves from the Pecos River valley to beneath the Southern High Plains as if a ground-water basin divide does not exist at depth. Discharge occurs as horizontal subsurface flow through the eastern boundary. In sensitivity analyses, Senger and Fogg (1987; Senger and others, 1987a) found that the k_h/k_v ratio of hydraulic conductivity in the Dockum group must be greater than 10^4 for their models to reproduce the vertical hydraulic-head gradient between the Ogallala and the Dockum hydrogeologic units. As previously noted, data are inadequate to evaluate this ratio, which at present has been studied using models only. Lower hydraulic head in the unit above the evaporite confining system has the effect of reducing vertical leakage through the evaporite section by approximately 17 percent in the model but does not significantly change hydraulic head in the deep-basin brine aquifer. In another simulation (Senger and Fogg, 1987; Senger and others, 1987a, 1987b) permeability of the evaporite confining system was increased by one order of magnitude to 2.8 $x10^{-3}$ md (2.8 $x10^{-18}$ m^2); the resulting poor match between computed and measured hydraulic heads in the deep-basin brine aquifer suggested that this average value might be too large and that 2.8 $x10^{-4}$ md (2.8 $x10^{-19}$ m^2) might be an upper limit of possible permeability values for the evaporite confining system.

However, a model simulation by INTERA (1986) with average k_v of the evaporite confining system increased from 10^{-5} to 10^{-4} md (10^{-20} to 10^{-19} m^2) resulted in a poor match between computed and measured hydraulic-head values for aquifers in Wolfcampian and Pennsylvanian rocks. INTERA (1986) concluded that either this

average vertical permeability value was too high or that horizontal permeability values (k_h) used for the deep-basin brine aquifer in the model were too low. Because k_h values are substantiated by field test data, INTERA (1984) favored the former explanation.

In the "base-case" model simulation by INTERA (1986), inflow from the evaporite confining system contributes nine percent of the total flow through the deep-basin brine aquifer. In a series of similar simulations to analyze sensitivity of model calculations to assumed parameter values, the percent of flow through the deep-basin brine aquifer contributed by the evaporite confining system varied between 35 and 61 percent.

In the models of INTERA (1984, 1986), intrastratal recharge and discharge in the deep-basin brine aquifer was allowed only across the western and eastern boundaries. Most of the discharge occurs at the northern part of the eastern boundary (near the Amarillo Uplift), where relatively high hydraulic conductivities were assigned to arkosic sandstones. Mass-conservation requirements resulted in unexpected and probably unrealistic flow directions across some parts of the boundaries of the deep-basin brine aquifer that might not occur if the model were to allow recharge from the Midland Basin and discharge to the Dalhart Basin (INTERA, 1984, 1986).

Atwood and others (1986) calculated that leakage through the evaporite confining system between the Pecos and Red Rivers contributes approximately 39 percent of the flow through the deep-basin brine aquifer. No sensitivity analyses were performed. The vertical hydraulic-head gradient across the evaporite confining system was probably more representative of the actual gradient than was used in the other models because Atwood and others (1986) used hydraulic-head data for the Dockum Group, which is lower than Ogallala hydraulic head. Discharge from the deep-basin brine aquifer was modeled both as horizontal subsurface flow through the eastern boundary (86 percent) and as surface discharge (14 percent) in the vicinity of the North Fork of the Red River.

Wirojanagud and others (1986) concluded that leakage through the evaporite confining system accounts for 33 percent of the total flow through the deep-basin brine aquifer. Lateral flow into and out of the deep-basin brine aquifer was allowed across the western and eastern boundaries and, in some simulations, across the western half of the northern boundary, representing the Amarillo Uplift. In the simulation of ground-water flow in the aquifer in Wolfcampian rocks, 41 percent of the flow discharged through the northern boundary and 59 percent discharged through the eastern boundary. In the deep-basin brine-aquifer simulation, six percent and 94 percent of the flow discharged through the northern boundary and the eastern boundary, respectively. These results indicate that ground-water flow in the deep-basin brine aquifer as a whole is more eastward than northward, unlike flow in the aquifer in Wolfcampian rocks beneath the northern part of the Southern High Plains.

Paleohydrology

In the Palo Duro Basin, extensive modification of physiography during the past 15 million years has resulted from tectonic and geomorphic processes, including effects of salt dissolution at top of the evaporite confining system (Gustavson and Finley, 1985). Because land-surface topography influences location of recharge and discharge zones and general flow directions, these physiographic changes were most likely important controls on paleohydrology and keys to understanding what future changes in ground-water flow regimes should be expected. Fluctuations in climate from 1.8 m.y. to 10,000 yr ago, related to episodic glaciation of the North American Great Plains (Flint, 1971, p. 19), also might have influenced paleohydrology.

Tertiary uplift of the western part of the Southern Great Plains following deposition of the Ogallala Formation created a land surface inclined to the southeast. Ground-water flow direction in the High Plains aquifer probably has always been toward the southeast because depositional fabric is aligned with the regional topographic and water table slopes. Ground-water flow velocities probably were greatly disrupted by excavation of the Pecos River watershed during the Neogene and retreat of the Eastern Caprock Escarpment during the Quaternary (Gustavson and Finley, 1985). It is likely that before the ground-water-basin divide developed beneath the Western Caprock Escarpment, the High Plains aquifer, the confined aquifer in the Dockum Group beneath the Southern High Plains, and the deep-basin aquifer in Wolfcampian rocks were recharged by intrastratal flow from the west. During the Holocene, recharge of the High Plains aquifer has been limited to precipitation in the semi-arid climate and downward flow through the vadose zone (Knowles and others, 1984; Luckey and others, 1986; Nativ and Smith, 1987; Ostercamp and Wood, 1987; Wood and Ostercamp, 1987). Recharge to confined parts of the Lower Dockum Group during the Holocene is restricted by low-permeability confining mudstones in the Upper

Dockum Group and upper part of the Lower Dockum Group and by the ground-water-basin divide located to the west. Whether a ground-water-basin divide influences recharge to the Wolfcampian rocks is not established.

To the extent that hydrogeology of the Dockum Group controls hydrology of salt-dissolution zones, the Pleistocene-to-Holocene changes in physiography and climate probably effected major changes in salt-dissolution rates. Decreased recharge to the overlying aquifers suggests that Holocene and Modern recharge rates to salt-dissolution zones might be significantly different from pre-Holocene rates (Dutton, 1987a).

Hydraulic heads of ground water in the Ogallala Formation and Dockum Group might be transient during Modern time, responding to changing physiography, climate, and recharge and discharge rates (Dutton and Orr, 1986). Hydraulic head in the High Plains aquifer, of course, has been changed by extensive development of water resources. It is also possible that hydraulic heads in the deep-basin brine aquifer represent transient conditions that are still responding to topographic and climatic changes (Senger and others, 1987a, 1987b).

Qualitative and quantitative estimates of ground-water age have been made based on numerical modeling and hydrochemical studies. Actually, because a given volume of ground water may include components that were recharged at different places and at different times, it may be more appropriate to discuss ground-water age distributions; but generally this has not been done in ground-water studies in the Palo Duro Basin. Distinct $\delta^{18}O$ and δD isotopic compositions of Dockum Group ground water have been used qualitatively to infer Pleistocene recharge (Dutton and Simpkins, 1986). Tritium concentrations in Ogallala Formation ground water are below detection limits, qualitatively indicating that vadose water recharged during the 1960s has not reached the water table and circulated through the aquifer (Nativ and Smith, 1987). A third example of a qualitative estimate of ground-water age concerns whether brine in San Andres unit 4 carbonate bed is a modified connate brine of Permian origin or

a meteoric recharge water that extensively reacted with evaporite and clay minerals (Dutton and Orr, 1986; Dutton, 1987b). If the former interpretation is correct, the ground-water age distribution is probably weighted towards 250 m.y.; if the latter is correct, the age distribution includes a large component less than 5 m.y., based on timing of latest uplift and development of a cross-formational hydraulic gradient (Senger and others, 1987a, 1987b). A fourth ground-water age issue is the comparison between (1) brine that is isotopically nonequilibrated with Wolfcampian rock at the western side of the basin and (2) brine that is equilibrated with rock at the center and eastern side of the basin (Fisher and Kreitler, 1987a, 1987b; Bein and others, in preparation). The former suggests a short flow path from a western recharge area, relative to a long flow path from the southeast leading to the center and eastern side of the basin, which hydrochemically confirms the distribution of flow paths calculated by numerical models.

Quantitative estimates of ground-water age in salt-dissolution zones and in hydrogeologic units of the deep-basin brine aquifer system, based on radioactive isotopes, are uncertain because many of the required assumptions are poorly constrained. For example, Dutton (1987a) reported measured carbon-14 activities of 1.7 and 3.3 % (Modern) and $\delta^{13}C$ of -3.1 °/oo and -1.9 °/oo, respectively, in ground water from two salt-dissolution zones. Accuracy of ground-water age estimates based on carbon-14 data is affected by how the following factors are taken into account: (1) addition of inorganic carbon from mineral solution, (2) precipitation of carbonate when calcite saturation is exceeded because of gypsum solution and the common ion effect, (3) isotope fractionation between water and carbonate minerals, (4) mixing of ground waters from different hydrogeologic units, and (5) generation of carbon-14 in the subsurface (Dutton, 1987a, in press).

For ground waters older than 30,000 years, only experimental methods are available for estimating ground-water age (Davis and Bentley, 1985). Zaikowski and

others (1987) interpreted analyses of noble gases in some deep-basin brines to suggest ground-water residence times of 100 to 300 million years. These must be considered preliminary estimates, however, because the composition of the host rock has not been adequately characterized and the sensitivity of the calculated ages to uncertainties in rock and water composition, contamination during sampling, and the possibility of extraformational sources of noble gases have not been critically evaluated (Fisher and Kreitler, 1987a, 1987b). Correlation of geochemical age estimates and results of numerical models of ground-water travel and residence times are poor. Numerical models, constrained to use a simplified physical configuration that homogenizes aquifer properties, have calculated a wide range of travel times. Lack of hydrogeologic continuity might significantly limit uniform circulation and account for the discrepancy between modeling results and hydrochemical interpretations.

There have been preliminary models of ground-water flow as a transient system. INTERA (1986) evaluated the effect of dewatering the Ogallala aquifer and the effect of hydrologic changes that could occur during the next 10,000 years. Senger and others (1987a, 1987b) performed computer simulations to test the hypothesis that hydraulic heads observed in the deep-basin brine aquifer are still responding to physiographic changes caused by tectonic and erosional processes that have occurred within the last 15 million years. The more recent occurrence of reservoir-pressure decline due to hydrocarbon production was also considered. The Senger and others (1987a, 1987b) model was used to simulate transient flow velocity and hydraulic head; temporal changes in hydraulic conductivities were not included. Results of simulations by Senger and others (1987a, 1987b) suggested that regional hydrodynamics of the Palo Duro Basin will not be significantly affected by hydrocarbon production. However, parts of the ground-water flow system could be affected locally by retreat of the Caprock Escarpment and by hydrocarbon production with reduction in reservoir pressure.

INTERA (1986) performed computer simulations to determine the effect of selected hydrologic changes on the ground-water flow regime. Three cases were analyzed: 1) a return to pluvial conditions, 2) a return to drought conditions, and 3) accelerated uplift of the Rocky Mountain region.

Pluvial conditions were simulated by instantaneously doubling the rate of recharge to the High Plains from 0.12 to 0.24 inch/yr (0.3 to 0.6 cm/yr). A 100,000-yr transient simulation was performed to investigate the transient response of the hydrologic system to a surface perturbation. The results indicated that, as expected, the hydrologic response is much slower for the San Andres carbonate bed and Wolfcampian layers than for the two layers in the upper flow system. After 1,000 years, the Ogallala and Dockum/Dewey Lake layers had achieved 69 percent of the steady-state hydraulic-head increase and lower layers were unchanged. After 10,000 years, the upper flow system had reached steady state; ground water in the San Andres unit-4 carbonate had achieved 21 percent of the steady-state hydraulic-head increase; and the Wolfcamp layer was unchanged. After 100,000 years, the San Andres carbonate layer had reached 90 percent and the Wolfcamp layer 67 percent of their respective steady-state hydraulic-head increases. These calculations are, of course, affected by the same assumptions and conceptual model limitations previously discussed.

A return to drought conditions was simulated by prescribing a zero recharge rate, which is equivalent to the dewatering of the aquifer by man. The ground-water flow model predicted that there would be widespread dewatering of the Ogallala aquifer along the Canadian Breaks and the Eastern Caprock Escarpment. Discharge to springs in Palo Duro Canyon ceased. Hydraulic heads in the Ogallala aquifer remained higher than those in the Wolfcamp carbonate throughout the model region. A drop in Ogallala water levels, therefore, does not reverse the downward hydraulic gradient and cause water to flow upward across the evaporite confining system.

Accelerated uplift of the Rocky Mountain region was simulated by increasing the western-boundary heads in the subsalt strata by 500 ft (152.4 m). The effect of the perturbation on the upper flow system was negligible. The hydraulic-head gradients in the San Andres carbonate bed and in Wolfcamp layers was shown to increase. Beneath Deaf Smith County, the Darcy velocity of the Wolfcamp layer increased by 40 percent. Heads in the deep-basin aquifer system remained below those in the High Plains aquifer, maintaining the downward-directed hydraulic gradient.

Issues and Uncertainties

Uncertainties in conceptual models result partly from scarcity of available data and from the spatial variability of hydrogeologic parameters. The uncertainty in hydrologic properties and continuity of hydrogeologic units is greatest for the evaporite confining system, the deep-basin brine aquifer in its updip recharge area west of the Pecos River (figs. 49 and 58), and peripheral and internal salt-dissolution zones. There is considerable uncertainty regarding the depth to which the Pecos River in New Mexico significantly affects hydraulic-head distribution and thereby affects recharge of the evaporite confining system and deep-basin brine aquifer beneath the Southern High Plains. Lithology and representative hydraulic properties are poorly defined in deeply buried hydrogeologic units where they extend toward outcrops east of the Palo Duro Basin, so that locations and rates of ground-water discharge from the evaporite confining system and deep-basin brine aquifer are uncertain. How the ground-water flow system is influenced by the prominent structural elements that bound the Palo Duro Basin and by associated faults and localized coarse-grained deposits also is poorly understood.

Confidence in results of ground-water modeling is limited by uncertainty in assumptions about steady-state flow, porous and fractured rock, leakage through the evaporite confining system, and homogeneity and isotropy of hydraulic properties. Differences in flow velocities and flow paths predicted by each numerical model result from differences in the adopted conceptual model, including areal extent of hydrologic units, hydrogeologic framework, hydrologic properties, and boundary conditions. Basic assumptions in regional models used to date for travel-time calculations include: (1) flow through the evaporite confining system and deep-basin brine aquifer can be adequately simulated as porous-media flow and available permeability estimates are valid, (2) variations in fluid density and salinity do not significantly affect flow

velocity, and (3) the ground-water flow system does not vary with time and is essentially at steady state. The assumption that variable density is inconsequential is incorrect (Davies, 1987) and will have to be changed in future models. The assumption of steady-state flow has been questioned because of the decline of water-table surface elevation beneath the northern part of the Southern High Plains, changes in fluid pressure from hydrocarbon production and enhanced recovery operations, and possible long-term changes in climate and physiography. If the regional ground-water flow system has been changing during the Holocene and is in a transient state, time-varying boundary conditions might be required for models of local parts of the flow system.

Interconnection between aquifer systems above and below the evaporite confining system and the amount and homogeneity of ground-water flow within the confining system depend on the prevalence of fractures in the evaporite strata. A Ground-water Modeling Panel convened by the U.S. Department of Energy Salt Repository Project in 1985 evaluated possible ground-water flow through fractured Permian rocks in the Palo Duro Basin, based on pertinent publications, geological reasoning, and testimony from DOE contractors. The Modeling Panel based its conclusions on a conceptual model that all formations in the evaporite confining system contain connate water of Permian origin and that the main component of flow is vertically downward, with a horizontal flow component within permeable beds of the salt section. The panel concluded that there is no direct evidence that regionally significant ground-water flow in the Palo Duro Basin in the Deaf Smith County area is controlled by fractures that might have higher permeability or faster flow paths than the bulk rock porous permeability. The Modeling Panel pointed out that there are indirect suggestions of subsurface faults that penetrate all or most of the stratigraphic sequence from basement to land surface in some locations, but believed that the fractures are not likely to contribute to travel times that are significantly shorter than in other path ways. This belief was based

on evidence that fractures in brittle strata tend to be filled with salt minerals and that fractures in halite beds at depth tend to close and heal. The Modeling Panel recommended that ground-water travel time analysis (1) be based on a porous medium model with stochastically evaluated hydrologic parameters; (2) consider a low-probability vertical fracture-controlled pathway in brittle rocks within the confining system; (3) not consider fracture-controlled pathways in halite beds due to lack of credible evidence for their existence; and (4) include results of site characterization tests of location, extent, and hydraulic significance of both surface lineament features and deep-basin structural features.

Uncertainties regarding the conceptualized relationship between hydrogeologic units include (1) amount and mechanism of leakage through the evaporite confining system, (2) lateral and surface hydrologic boundary conditions for the deep-basin brine aquifers, (3) mechanism and rate of recharge to the High Plains aquifer, including role of playa lakes in focusing recharge and nature of transport processes in the thick vadose zone overlying the water table, (4) age and residence time of ground water in the Dockum Group and salt-dissolution zones, (5) extent and effect of stratigraphic heterogeneities on uniformity of flow and flushing of connate water in all hydrogeologic units, and (6) analysis of flow potential where water density significantly varies because of salinity. Extent of cross-formational ground-water movement between the previously defined hydrostratigraphic units is controlled by differences in potentiometric levels and hydraulic conductivity. The potentiometric level of each hydrostratigraphic unit reflects: (1) the regional southeast slope of the Southern Great Plains physiographic province; (2) amount of cross-formational flow; (3) influence of Cenozoic tectonism and physiographic changes; (4) intrastratal variations in hydraulic conductivity reflecting depositional patterns or diagenesis; and (5) variations in salinity.

Estimation of degree of circulation within hydrogeologic units is uncertain, partly because vertically distributed data is usually absent and partly because stratigraphic

data on rock heterogeneity is incomplete. Numerical models are constrained to use a simplified physical configuration that homogenizes aquifer properties. However, lack of stratigraphic continuity might significantly limit uniform circulation within a hydrogeologic unit and account for some discrepancies between modeling results and hydrochemical interpretations. Domenico and Robbins (1985) concluded that recharge water can completely displace connate fluids only if the recharge area includes most of the width and depth of the hydrogeologic unit. In heterogeneous sedimentary deposits in the Palo Duro Basin, ground-water origin may vary from meteoric water, in continuous transmissive zones that are hydrologically connected to recharge areas, to modified connate water in low-permeability confining beds or in discontinuous transmissive beds that are poorly interconnected with the regional flow system. Interpretation of hydrochemical data and numerical modeling must be integrated to develop complete and consistent conceptual models of regional flow systems in and around the Palo Duro Basin.

DISSOLUTION OF PERMIAN BEDDED SALT

T. C. Gustavson

Introduction

Permian bedded salt (halite) of the Palo Duro Basin occurs in sufficient thickness and at appropriate depth to have been considered as a potential repository for high-level nuclear waste. Low permeability, low moisture content, high plasticity, and high gamma-ray shielding are properties that make salt desirable as a host rock for a nuclear waste repository. The major drawback in using salt as a repository host rock is its extremely high solubility. Studies of salt dissolution included investigations of the characteristics, geographic and stratigraphic extent, processes, effects of dissolution, and timing of dissolution. The results of these studies, which were published in reports by Gustavson (1980), Gustavson and others (1980, 1981, 1982), Simpkins and others (1981), Gustavson and Budnik (1985), Gustavson and Finley (1985), Gustavson (1986a) Dutton (1987a, in press), Fracasso and Hovorka (1986), Hovorka and Granger (1988), and McGookey and others (1988) are summarized here.

Subsurface dissolution of salt has been widespread throughout much of the Permian Basin salt area and was recognized as early as 1901 in the Meade Basin in southwestern Kansas (Johnson, 1901). Playa lake basins on the High Plains surface were attributed to dissolution of salt and subsidence (Baker, 1915). Adams (1963), Jordan and Vosberg (1963), Brown (1967), Hills (1968), Bachman and Johnson (1973), Johnson (1976, 1981) all attributed certain subsurface structures and anomalous thinning of shallow buried evaporite and gypsum sequences to dissolution. Evans and Meade (1945) related folding of Triassic and Tertiary rocks locally to subsidence resulting from salt dissolution along the eastern margin of the Palo Duro Basin. Eck and Redfield (1963) and Bock and Crane (1963) identified numerous chimneys in the

excavation for Sanford Dam on the Canadian River northeast of Amarillo that were attributed to dissolution and collapse of gypsum beds in the Blaine Formation. Schultz (1977) attributed sinkholes filled with Ogallala sediments to collapse following dissolution of salt. Wilson (in press) recognized that Ogallala sediments exposed along the southern flank of the Canadian River valley north of Amarillo, Texas, were deposited in subsiding valleys that resulted from dissolution, and he recognized that the Ogallala deposits were later deformed by further dissolution-induced subsidence.

Geologic Setting

The Amarillo Uplift and the Matador Arch are the major positive structural elements bounding the Palo Duro Basin (fig. 59). Tectonic activity that produced these features apparently began in the Late Mississippian and continued into the Early Permian (Ruppel, this volume). Arkosic sands and gravels were derived from and deposited near the principal uplifts (Handford and Dutton, 1980). Away from principal uplifts, Late Pennsylvanian sedimentation was dominated by shelf carbonates, and fine-grained clastics, which filled the deeper parts of the basin. Salt, anhydrite, dolomite, limestone and red beds compose Permian strata in the Palo Duro Basin (fig. 60) (Presley, 1979a, 1979b, 1980a, 1980b; Fracasso and Hovorka, 1986). Fluvial, deltaic and lacustrine sandstone and mudstone of the Dockum Group were deposited during the Triassic (McGowen and others, 1979). Erosional remnants of Jurassic and Cretaceous marine sediments are preserved locally.

Regional uplift during the Late Cretaceous to Early Tertiary resulted in extensive erosion of Permian through Cretaceous sediments. During the Late Tertiary sediment derived from the Rocky Mountains to the west was deposited as the fluvial facies of the Miocene-Pliocene Ogallala Formation (Seni, 1980; Gustavson and Winkler, 1988). The upper part of the Ogallala consists of as much as 114.8 ft (35 m) of eolian

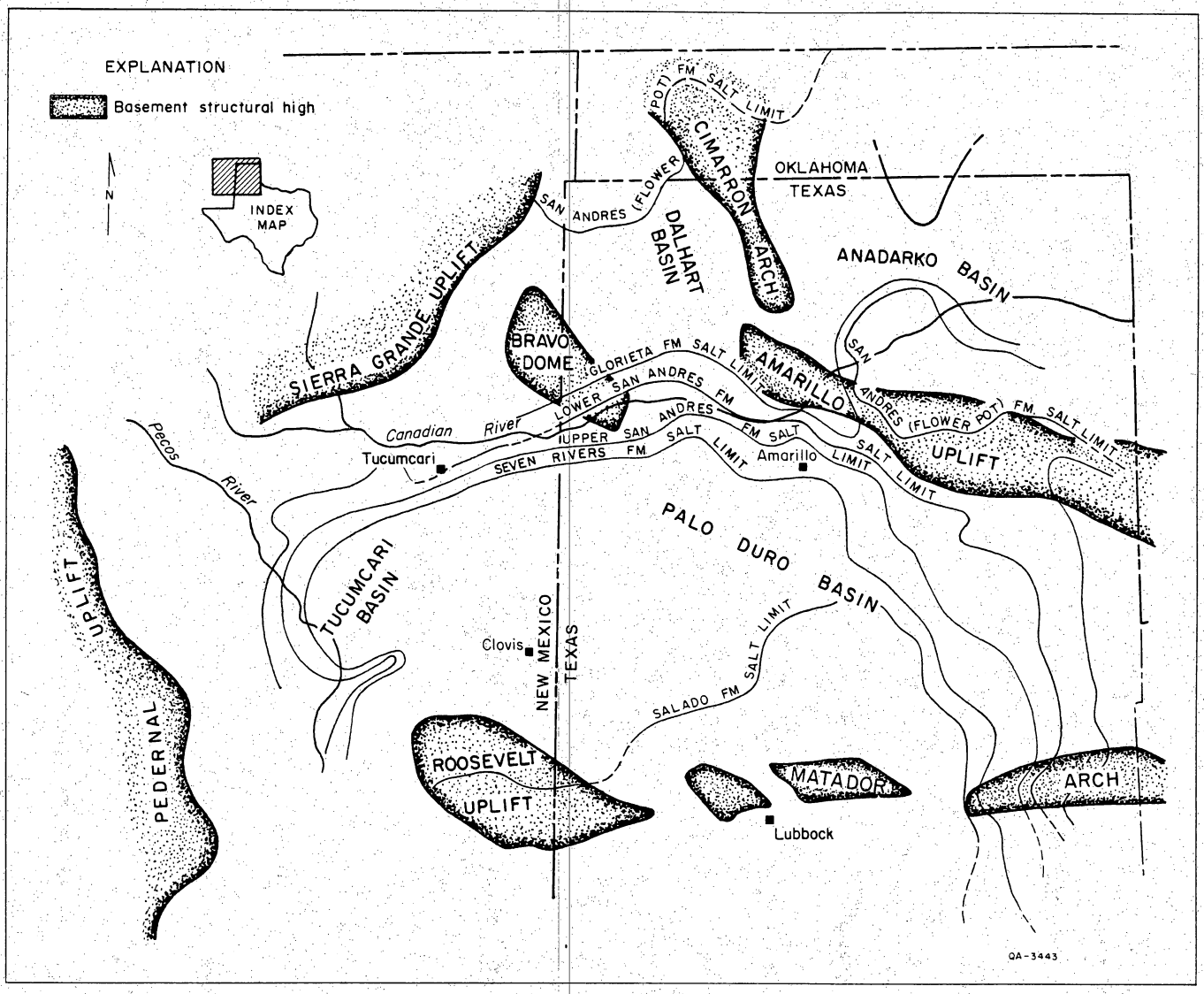


Figure 59. Major structural elements, Texas Panhandle and surrounding areas. After Nicholson (1960) and Gustavson and Finley (1985). Limits of Permian bedded salts are closely associated with the structural margins of the Palo Duro Basin. Structurally high salt units are most likely to be affected by dissolution.

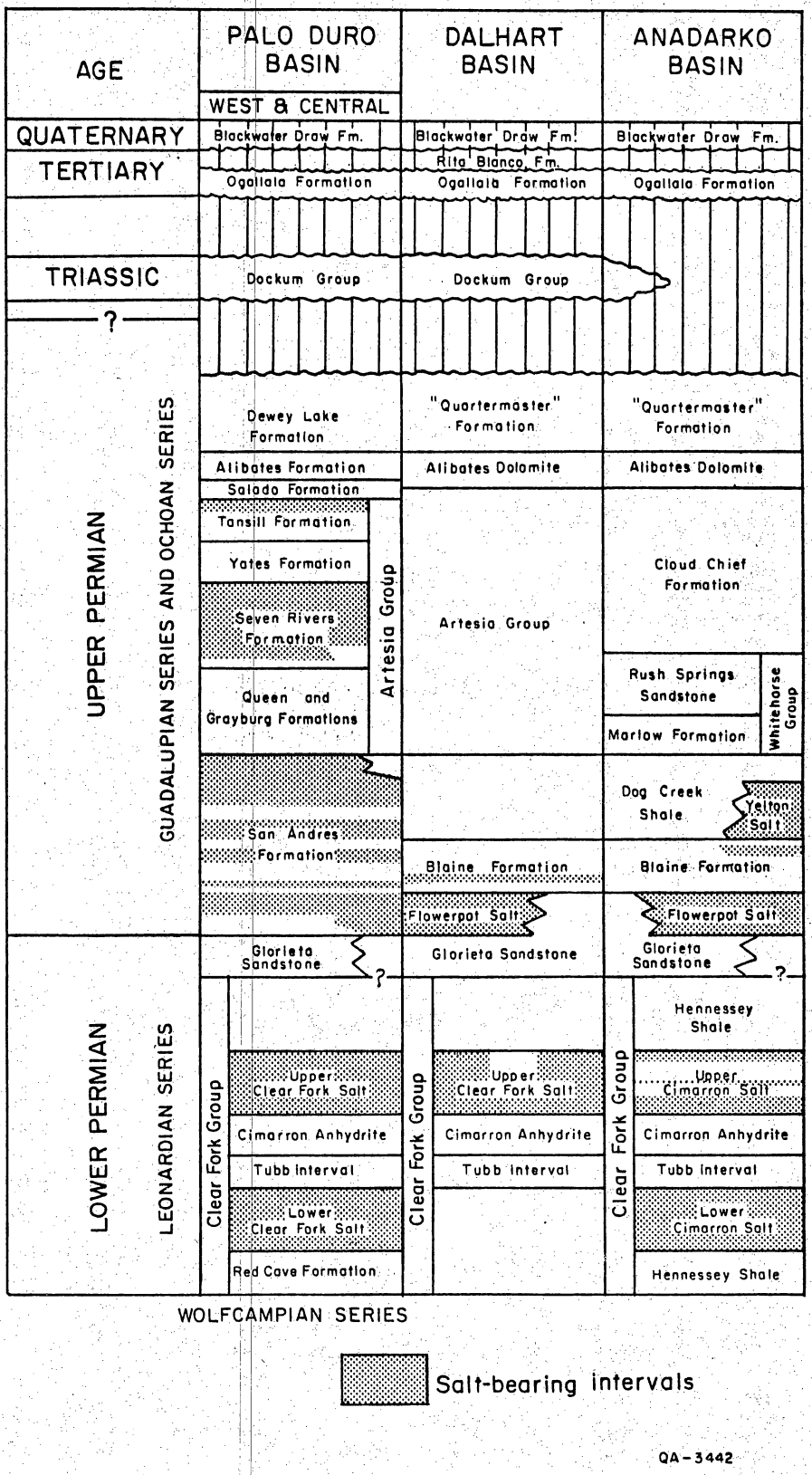


Figure 60. Stratigraphic nomenclature of Permian and younger strata in the Texas Panhandle and western Oklahoma. Principal salt units are shown in gray. From data in Tait and others (1962), Jordan and Vosberg (1963), Fay (1965), McKee and others (1967), and compiled by Johnson (1976).

sediments throughout most of the study area. Locally lacustrine sediments of the Pliocene Blanco and Rita Blanca Formations overlie the Ogallala. The Quaternary eolian Blackwater Draw Formation overlies the Ogallala Formation and locally the Blanco and Rita Blanca Formations.

Zones of extensive salt dissolution, collectively called the peripheral dissolution zone, have developed along the western, northern, and eastern structural margins of the Palo Duro Basin where salt-bearing strata are present at depths of approximately 328 to 656 ft (~100 to ~200 m) (fig. 59) (Gustavson and others, 1980; Gustavson and Finley, 1985; Gustavson, 1986a). The eastern and western Caprock Escarpments of the Llano Estacado or Southern High Plains and the southern margin of the Canadian River valley, the Pecos River valley, and the western part of the Rolling Plains overlie the peripheral dissolution zone. Dissolution beneath the High Plains, which overlies the central part of the Palo Duro Basin, is also widespread and is called the interior dissolution zone.

Recognition of Salt Dissolution

Recognition of zones of salt dissolution in the Palo Duro Basin comes from several lines of evidence including water chemistry, geophysical log, core, and outcrop data (Gustavson and others, 1980, 1982; Gustavson and Finley, 1985; Gustavson, 1986a, 1986b, 1986c; Johnson, 1981; DeConto and Murphy, 1986; Boyd and Murphy, 1987). The most compelling evidence for active salt dissolution is the high solute loads carried by streams that drain strata exposed at the updip margins of the Palo Duro Basin (figs. 59 and 61). Along the eastern margin of the Palo Duro Basin the Prairie Dog Town Fork of the Red River carries a mean annual solute load of 9.2×10^5 short tons (8.3×10^5 metric tons [t]) of chloride (U.S. Geological Survey, 1969-1983). Along the northern margin of the basin the Canadian River carries more

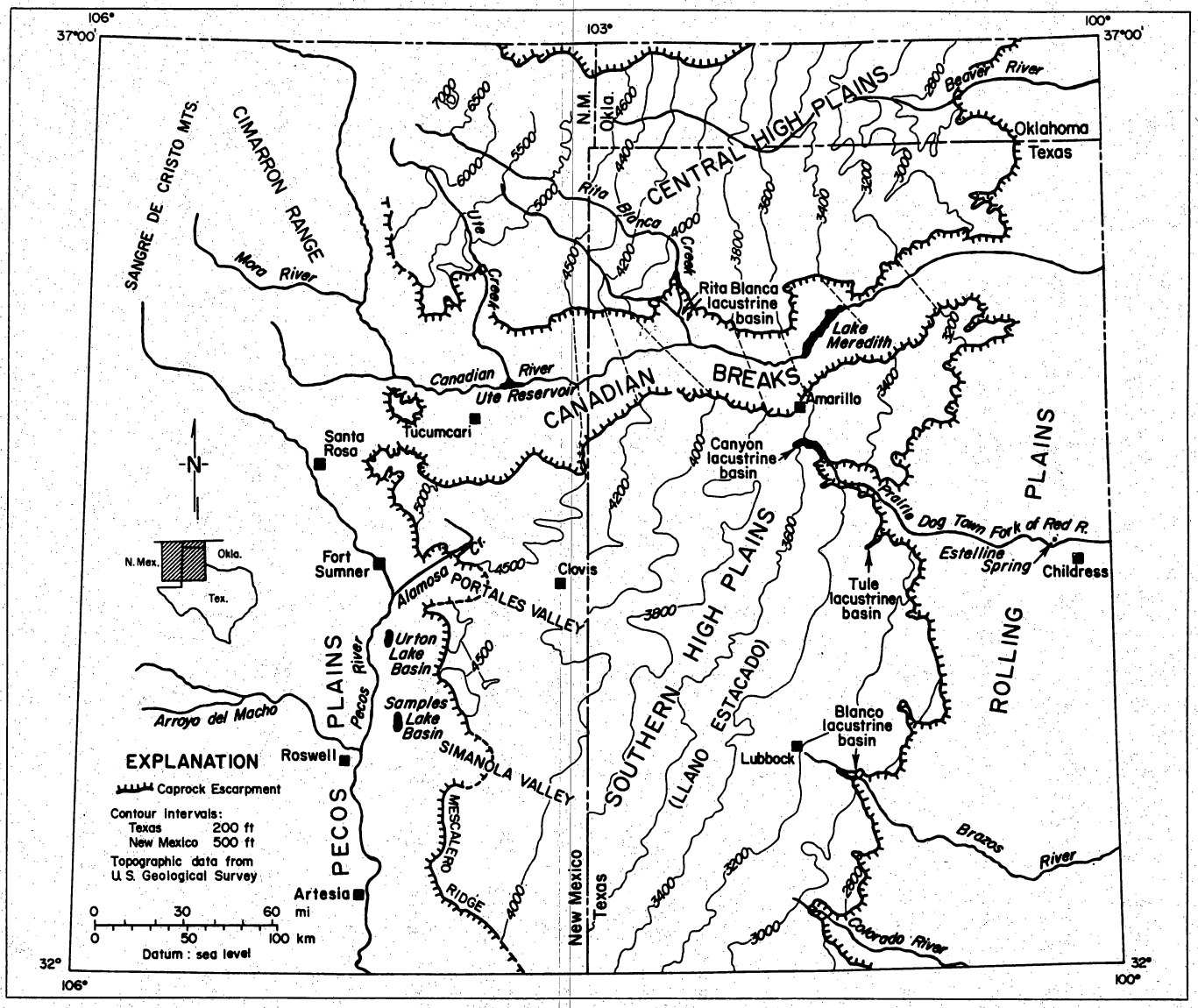


Figure 61. Physiography of eastern New Mexico and the Oklahoma and Texas Panhandles.

than 5.4 x 10^4 tons (4.9 x 10^4 t) of sodium chloride annually (U.S. Bureau of Reclamation, 1979). The western margin of the Palo Duro Basin is drained by the Pecos River, which was estimated to carry as much as 2.7 x 10^5 tons (2.4 x 10^5 t) of sodium chloride annually at Artesia, New Mexico (Morgan, 1941). Brine springs, salt seeps, and salt pans are present locally along these valleys.

Hydrologic testing of strata immediately above uppermost salts in the Stone and Webster Engineering Corporation (SWEC) Mansfield No. 2 well in Oldham County and the SWEC Sawyer No. 2 well in Donley County produced sodium chloride brines, which suggests that dissolution of the Seven Rivers salt is probably under way in the Canadian River valley at a depth of 811 ft (247 m) in the Mansfield No. 2 well, and that dissolution of San Andres salt is probably under way in the Rolling Plains at a depth of 830 ft (253 m) in the Sawyer No. 2 well. On the basis of radiocarbon dates these Na-Cl ground waters are no older than 16,200 and 23,000 yr respectively (Dutton, 1987a). Dissolution of Seven Rivers salt beneath the Southern High Plains is probably under way in the SWEC Detten No. 2 well at a depth of 1,332 ft (405 m) and in the SWEC Harman No 1. well at a depth of 1,225 ft (373 m) (Dutton, 1987a). The ages of these Na-Cl ground waters were not determined.

The abrupt loss of salt sequences between closely spaced hydrocarbon exploration wells can be interpreted from geophysical logs. If the loss of salt were due to thinning towards the basin margin or to facies changes, non-salt as well as salt units should be affected. In the absence of facies or thickness changes of interbedded non-salt units dissolution is a likely cause of salt thinning especially where collapse of overlying strata in wells that are missing salt is evident (figs. 62, 63, and 64). Regional maps of the extent of salt dissolution (fig. 65) were prepared from interpretations of geophysical logs (Presley 1979a, 1979b, 1980a, 1980b) as were regional cross sections (McGookey and others, 1988).

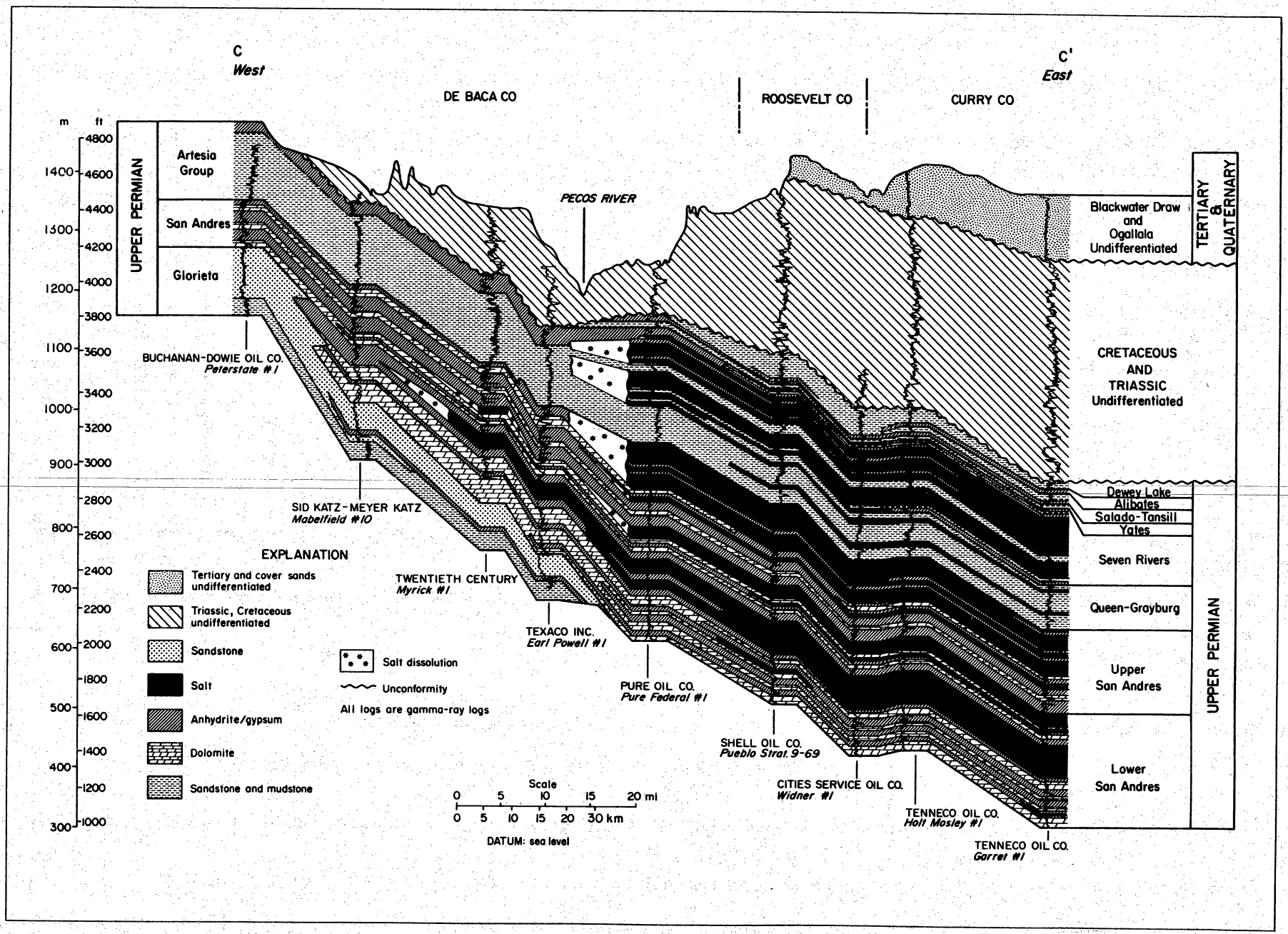


Figure 62. Stratigraphic cross section illustrating salt dissolution and collapse of strata beneath the Pecos River Valley. See figure 65 for the location of section C-C'.

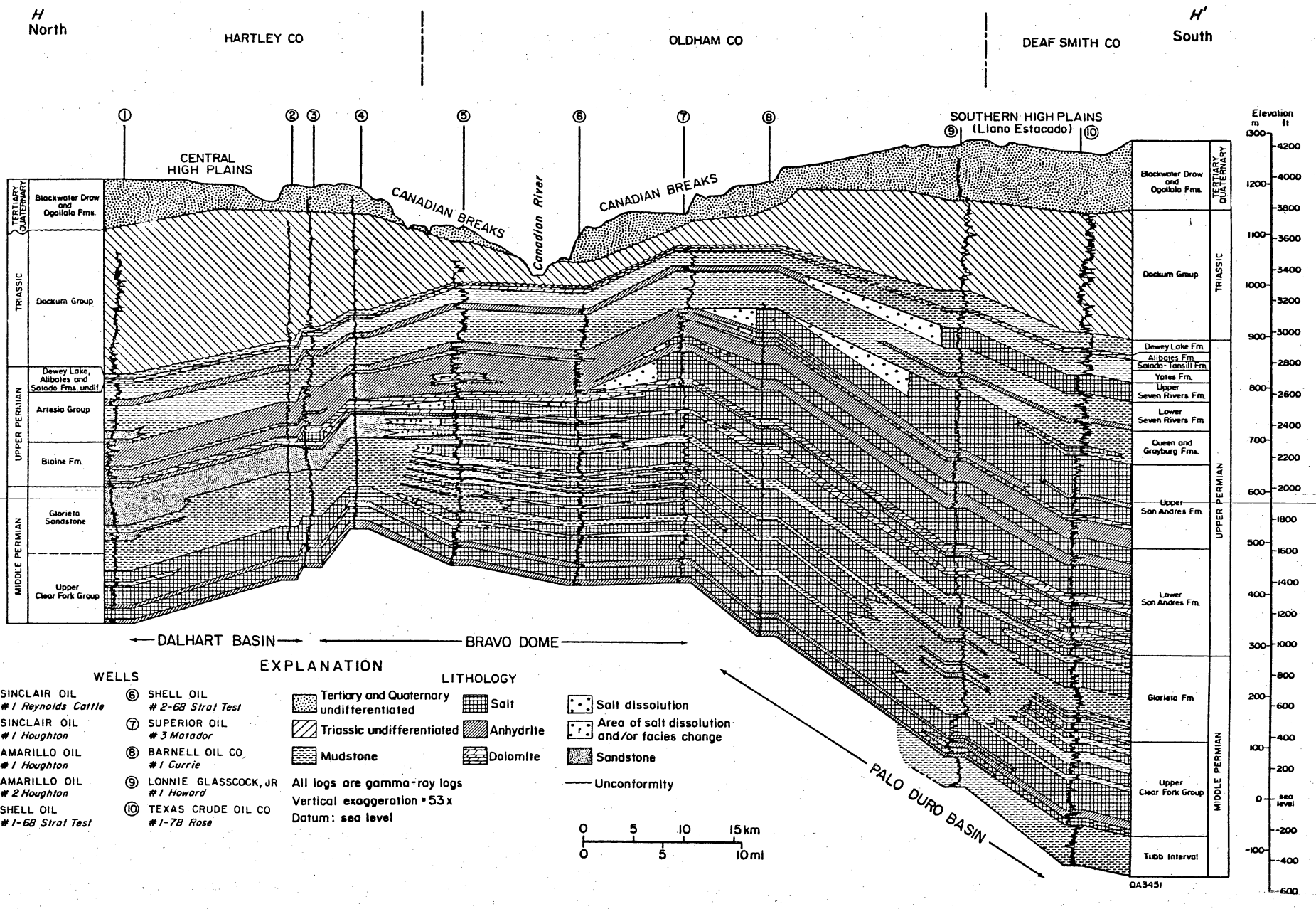


Figure 63. Stratigraphic section H-H' illustrates the loss of 600 ft (> 200 m) of salt along the northern margin of the Palo Duro Basin and beneath the Canadian River Valley. Structural collapse of overlying units and the continuation of nonsalt units in the upper part of the section across the structurally high Bravo Dome without significant changes suggests that salts have thinned because of dissolution and not because of facies change. A facies change from salt to sandstone occurs in the upper Glorieta formation. See figure 69 for the location of section H-H'. Cross section modified from McGookey and others (1988).

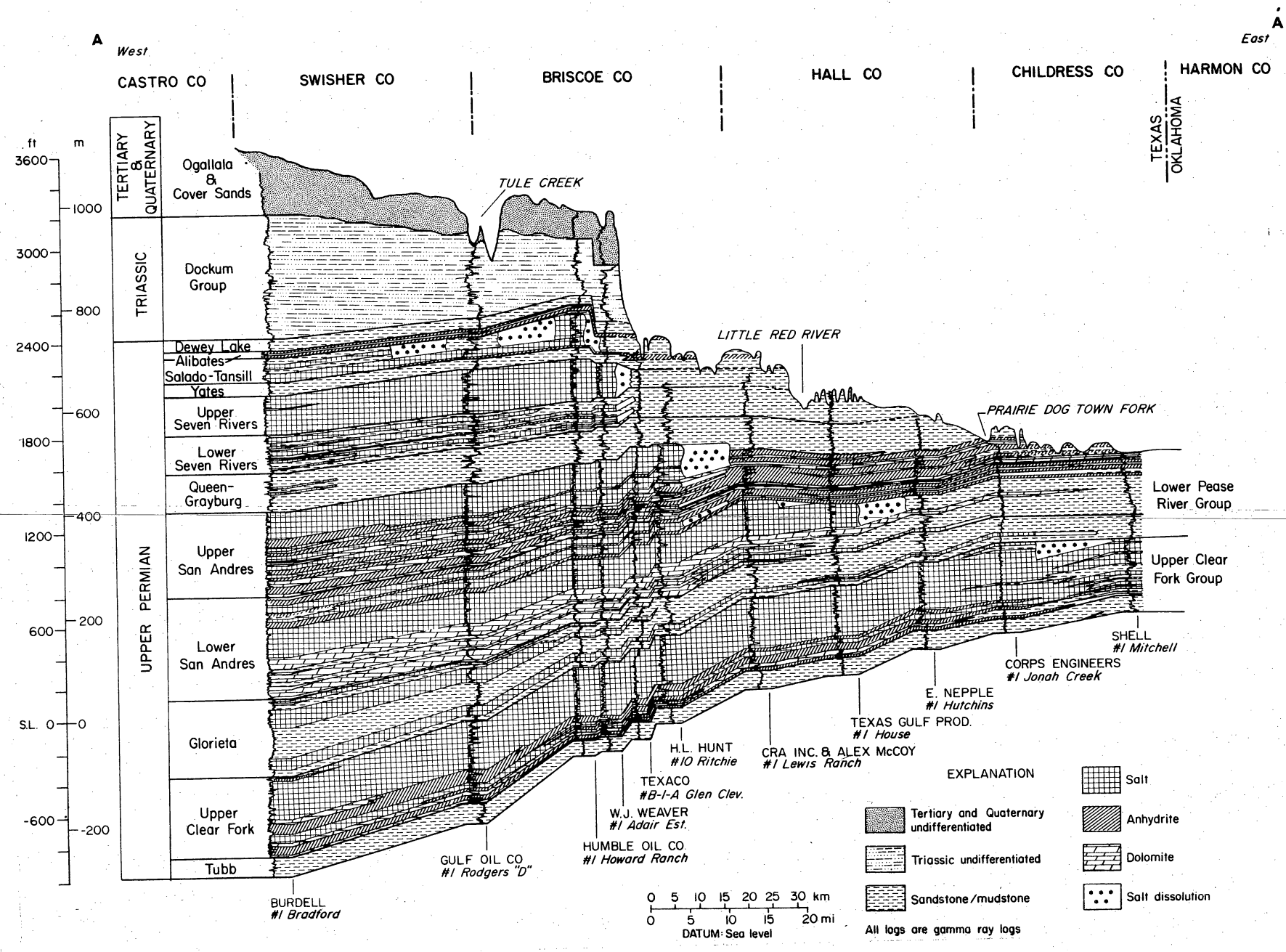


Figure 64. Stratigraphic section A-A' illustrating salt dissolution and collapse of strata beneath the Caprock Escarpment and the Rolling Plains. See figure 72 for the location of section A-A'. After Gustavson and others (1980).

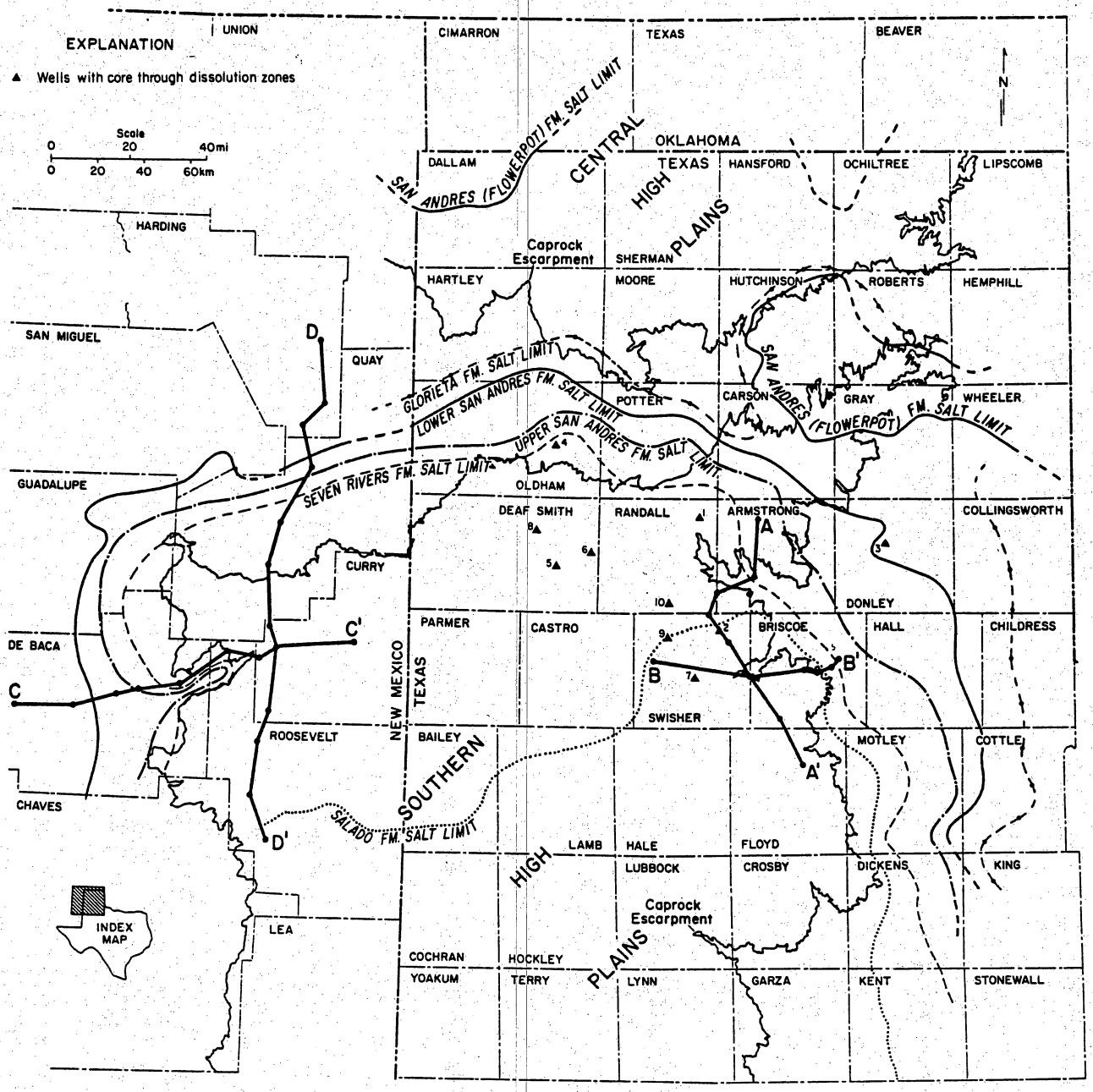


Figure 65. Zones of salt dissolution in eastern New Mexico and the Texas and Oklahoma Panhandles. Lines indicate the present extent of salt in the study region. In stratigraphic succession upward from the Glorieta formation to the Salado Formation, increasing amounts of salt are preserved toward the southwest corner of the Texas Panhandle. Some San Andres Formation salts are preserved in northwestern Dallam County, and some Glorieta and San Andres salts are preserved near Hutchinson County. A peripherial dissolution zone is approximately located by salt-limit lines of the Seven Rivers, San Andres and Glorieta Formations. an interior dissolution is approximated by the area overlain by the Southern High Plains. Wells with core through strata from which salt has been dissolved are indicated by numbered triangles: (1) DOE-Gruy Federal Rex White No. 1, (2) DOE-Gruy Federal Grabbe No. 1, (3) SWEC Sawyer No. 1, (4) SWEC Mansfield No. 1, (5) SWEC Detten No. 1, (6) SWEC G. Freimel No. 1, (7) SWEC Zeeck No. 1, (8) SWEC J. Freimel No. 1, (9) SWEC Harman No. 1, (10) SWEC Holtzclaw No. 1. Line C-C' is shown in figure 62; other sections given by Gustavson and Finley (1985).

Brecciated zones, extension fractures filled with satin spar (gypsum) (Gustavson, 1980; Goldstein and Collins, 1984) and "fabrics of wavy-laminated to brecciated gypsiferous mudstone and coarse granule to boulder breccia with rotated allochthonous blocks" (Hovorka and Granger, 1988, p. 152-153) overlie the uppermost salt in cores of the dissolution zone. These features, which are interpreted to have resulted from dissolution of salt, do not occur below the uppermost salt bed, but occur only in the stratigraphic intervals above the uppermost salt beds.

Outcrops along the Pecos and Canadian River valleys and in the Rolling Plains contain systems of antitaxial crack-seal veins filled with satin spar (fig. 66) breccia-filled chimneys (fig. 67) breccia beds (fig. 68) and remnants of caverns, all of which are thought to have resulted from dissolution of salt and collapse of overlying sediments (Gustavson and others, 1980).

Dissolution-Induced Subsidence

Subsidence that resulted from dissolution of salt is most easily recognized by comparing a structure-contour map of a surface that underlies the salt-bearing interval to a structure-contour map of a surface that overlies the salts. The Tubb interval, which underlies all the salt-bearing units that have undergone dissolution, is an informally-named thin sequence of clastic sediments within the Clear Fork Group that is recognized in the subsurface throughout the Palo Duro Basin (figs. 69 and 70). Consequently, structures on the Tubb interval are due to tectonic movements and not to dissolution-induced subsidence. Notable structural elements on the Tubb interval along the northern margin of the Palo Duro Basin are the Bravo Dome, the Amarillo Uplift, the Bush and John Ray Domes, which are separated by 787.4 ft (240 m) of elevation, the Whittenburg Trough with relief of about 196.9 ft (~60 m), and the deep trough (~690 ft [~210 m]) connecting the Dalhart and Palo Duro Basins (fig. 69).



Figure 66. Fractures in mudstones of the Permian Quartermaster Formation filled with satin spar (gypsum). Fractures result from collapse owing to dissolution of underlying salt beds. Photograph is from Caprock Canyons State Park in Briscoe County.

(a)



(b)



Figure 67. (a) Oval collapse chimney in Permian mudstones filled with Tertiary or younger fluvial sand and gravel. The collapse chimney filling is slightly cemented. Note the downwarping of the Permian enclosing beds. Figure on collapse chimney for scale. Photograph is of an area on Texas Highway 136 approximately 3.1 mi (5 km) north of Borger, Hutchinson County, Texas. (b) Erosionally resistant collapse chimney breccia in Permian mudstones. Breccias are carbonate cemented and consist of a mixture of fluvial gravels and blocks of Permian sediments. Lake Meredith National Recreational Area, Canadian River Valley.



Figure 68. Collapse breccia composed of laminated gypsum in a gypsum matrix. See lens cap for scale. Photograph from the Caprock Canyons State Park, Briscoe County, Texas.

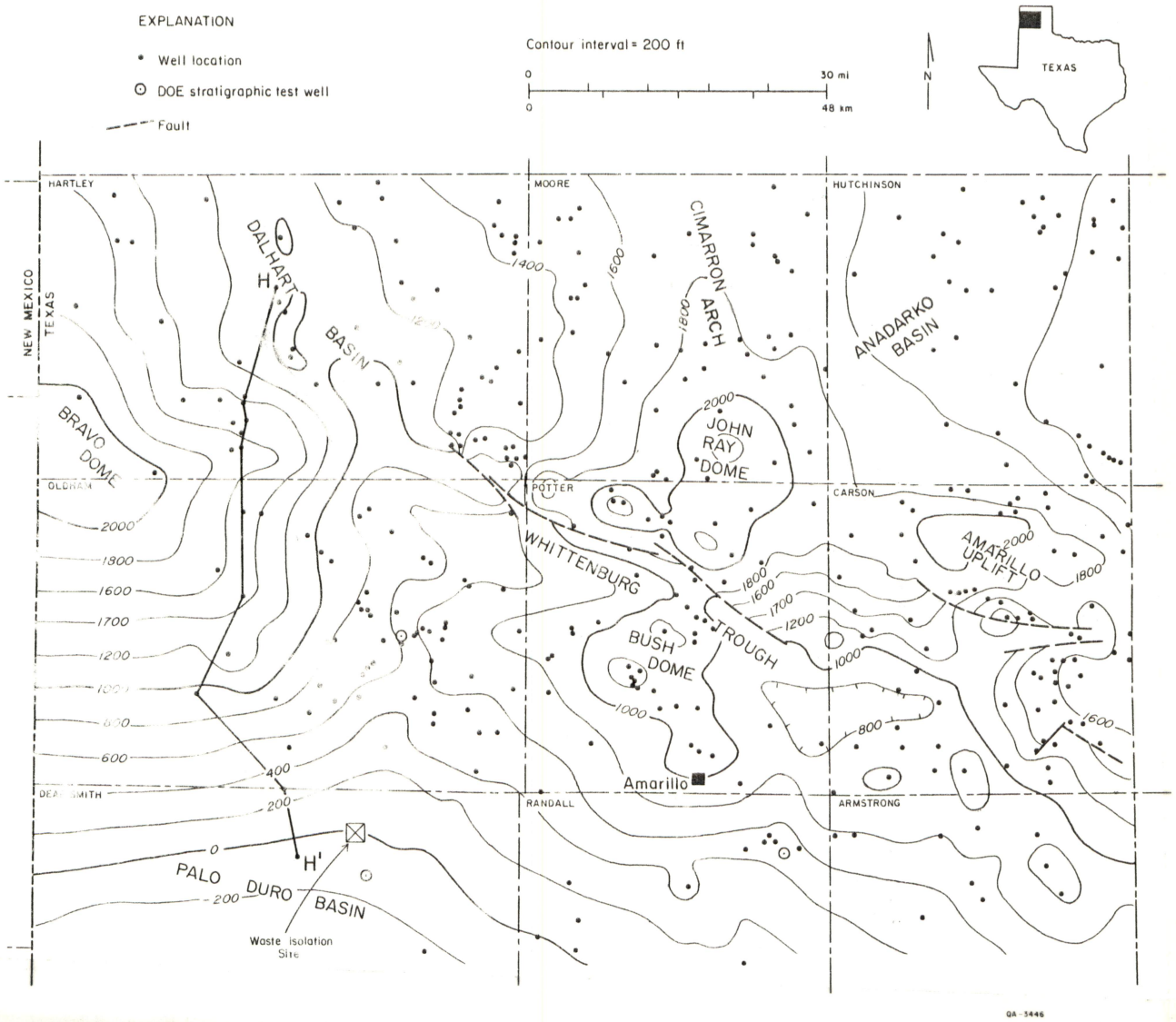


Figure 69. Structure-contour map on top of the middle Permian Tubb interval shows the major structural elements unaffected by salt dissolution at the northwestern margin of the Palo Duro Basin. After Budnik (1988). H-H' locates figure 63.

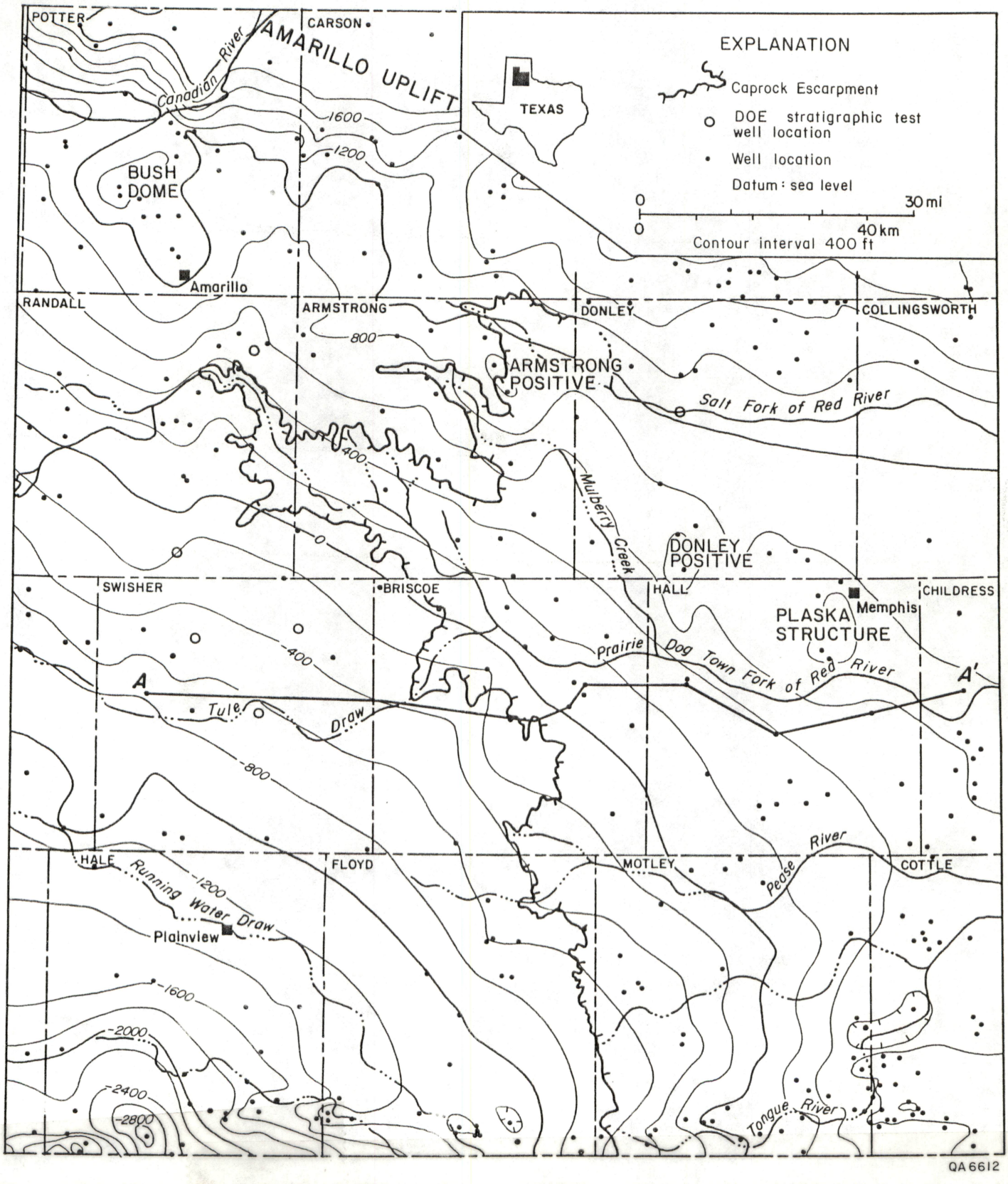


Figure 70. Structure-contour map on top of the middle Permian Tubb interval shows the major structural elements unaffected by salt dissolution at the eastern margin the Palo Duro Basin. Section A-A' shown in figure 64. After Gustavson and Simpkins (in press).

Notable structural elements along the eastern margin of the Palo Duro Basin are the Amarillo Uplift, the Armstrong County and Donley County positives, and the Plaska structure (fig. 70). The Tubb interval dips to the southwest from 26.4 to 132 ft/mi (5 to 25 m/km) in the Palo Duro Basin.

The thickness of salt remaining along the northern and eastern margins of the Palo Duro Basin is illustrated in figures 71 to 73. Along the northern margin of the Palo Duro Basin salts of the Clear Fork Group through the Seven Rivers Formation are preserved. Along the eastern margin salts from the Clear Fork Group through the Salado Formation are preserved. Collectively, salts along the northern margin thin as much as 1,082.7 ft (330 m) between central Randall and northern Potter County. Salts of the lower San Andres Formation thin most rapidly along the southwestern flank of the Amarillo Uplift. Along the eastern margin of the basin the collective salt thickness of the Clear Fork Group through the Salado Formation diminishes as much as 330 m between northeastern Swisher and northwestern Childress Counties. In this region salts thin most rapidly along the southwestern margin of the Amarillo Uplift and along and east of the Caprock Escarpment.

Comparison of the salt thickness map of the northern margin of the basin (fig. 71) to the structure-contour map of the Tubb interval for the same area (fig. 69) shows that (1) thick San Andres salts are missing over the structurally high areas of the axis of the Amarillo Uplift and John Ray and Bravo Domes, (2) Glorieta salts are missing over the John Ray Dome and Amarillo Uplift, (3) Clear Fork salts thin over the Amarillo Uplift and the John Ray Dome, (4) Upper San Andres and Seven Rivers salts thin over the Bush Dome, and (5) San Andres salts extend farthest north in the area of the Whittenburg Trough between the Palo Duro and Dalhart Basins in eastern Oldham County.

Structure on the Alibates Formation (figs. 74 and 75), which overlies all of the salt-bearing units, is markedly different from structure on the Tubb interval (figs. 69

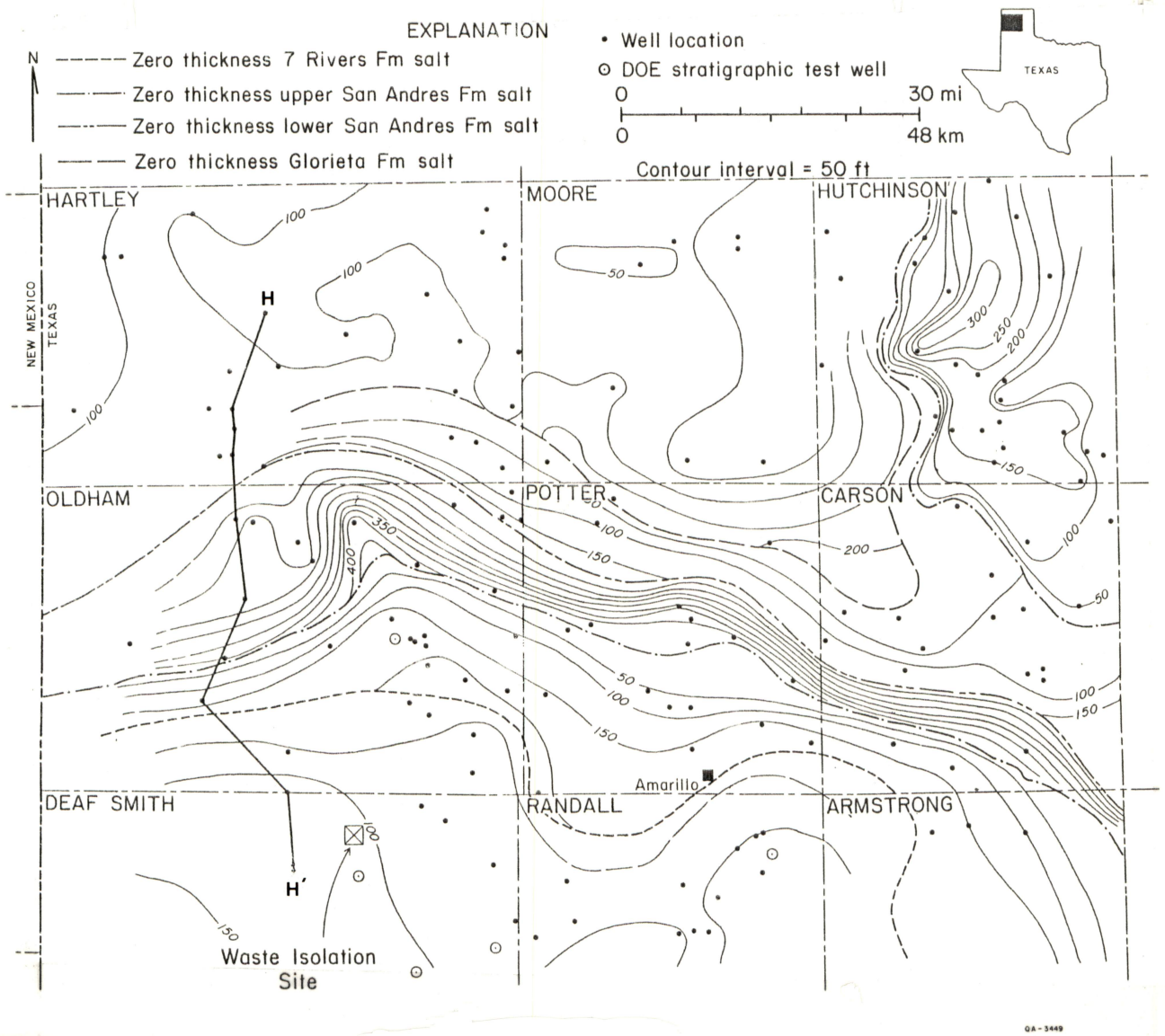


Figure 71. Net salt thickness map at the northwestern margin of the Palo Duro Basin, including the upper Clear Fork Group, Glorieta (Sandstone) formation, lower San Andres Formation (includes Blaine Formation salts in the Dalhart Basin and Flowerpot Formation salts in the Anadarko Basin), upper San Andres Formation, and Seven Rivers Formation. For each formation, only the salt-bearing interval which is not overlain by salt in younger strata is included. Salt thickness data for Hutchinson County are from McGookey (1981). H-H' locates figure 63.

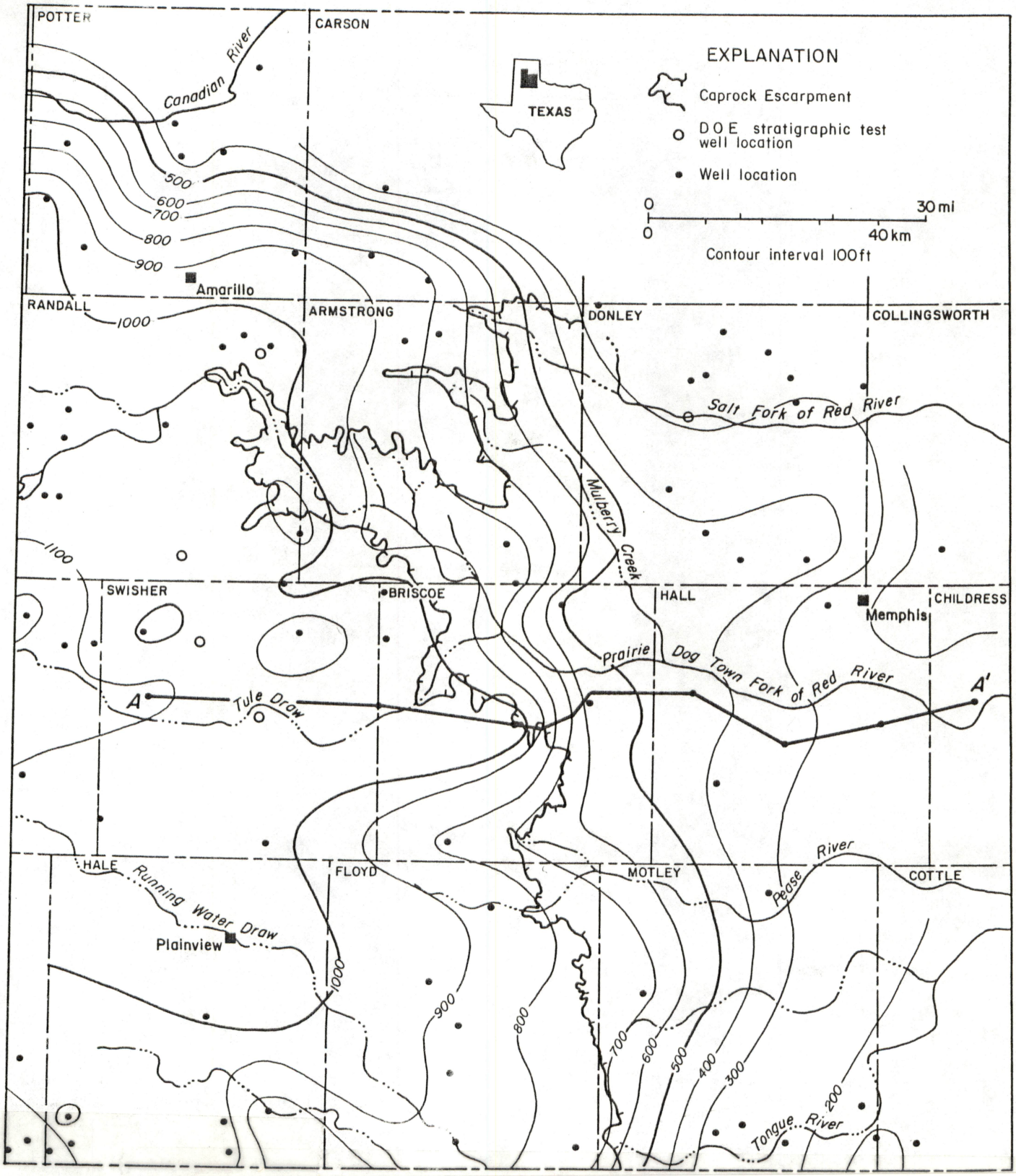


Figure 72. Net salt thickness map at the eastern margin the Palo Duro Basin, including the upper Clear Fork Group, San Andres formation, Seven Rivers Formation, and the Salado Formation. A-A' locates figure 64.

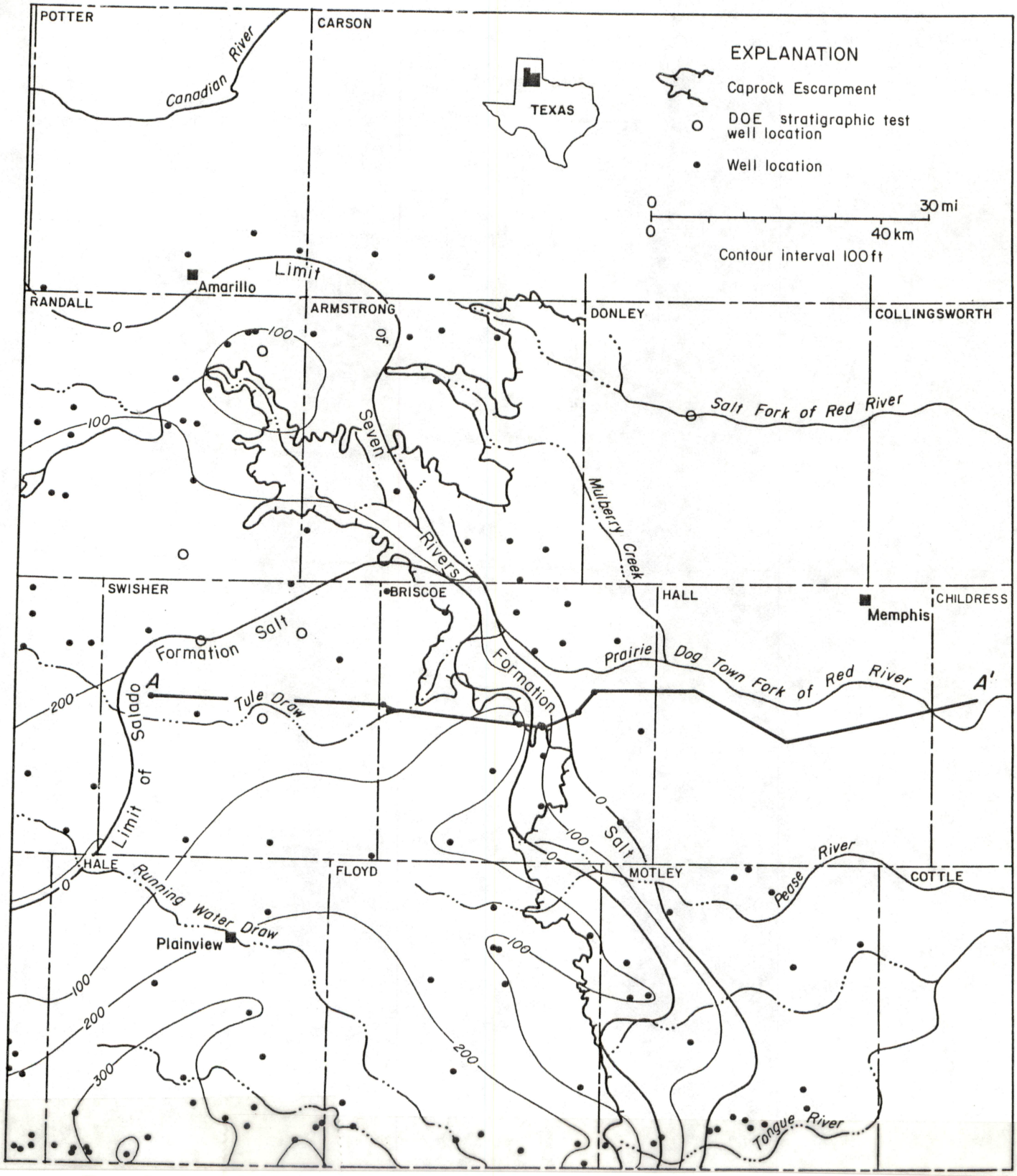


Figure 73. Net salt thickness map at the eastern margin the Palo Duro Basin, including slices of the Salado and Seven Rivers Formations. Salt thickness for the Salado Formation throughout the map area are shown in addition to salt thickness of the Seven Rivers Formation where there is no overlying Salado salt. A-A' locates figure 64.

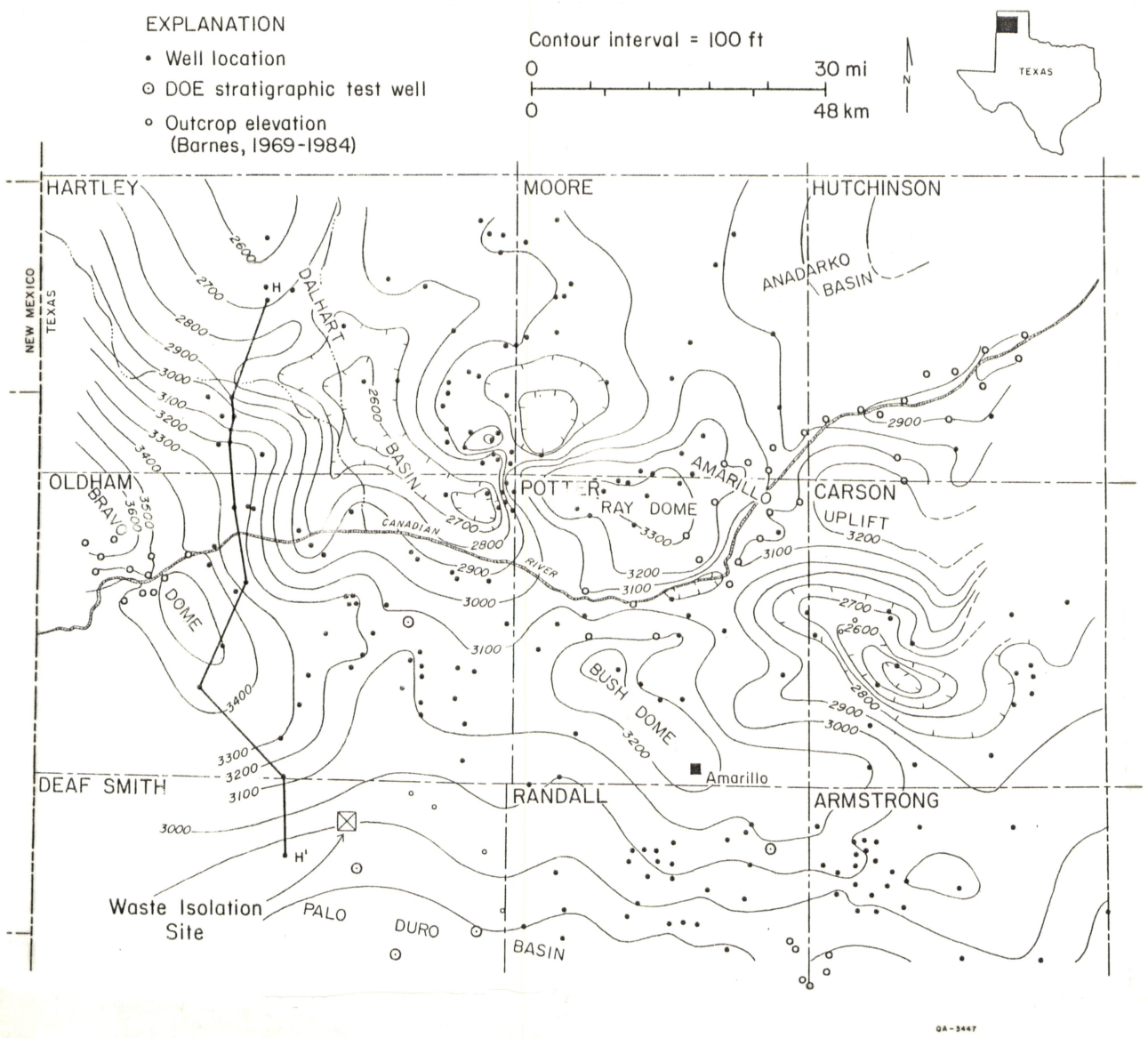


Figure 74. Structure-contour map on the upper Permian Alibates Formation at the northwestern margin of the Palo Duro Basin. H-H' locates figure 63.

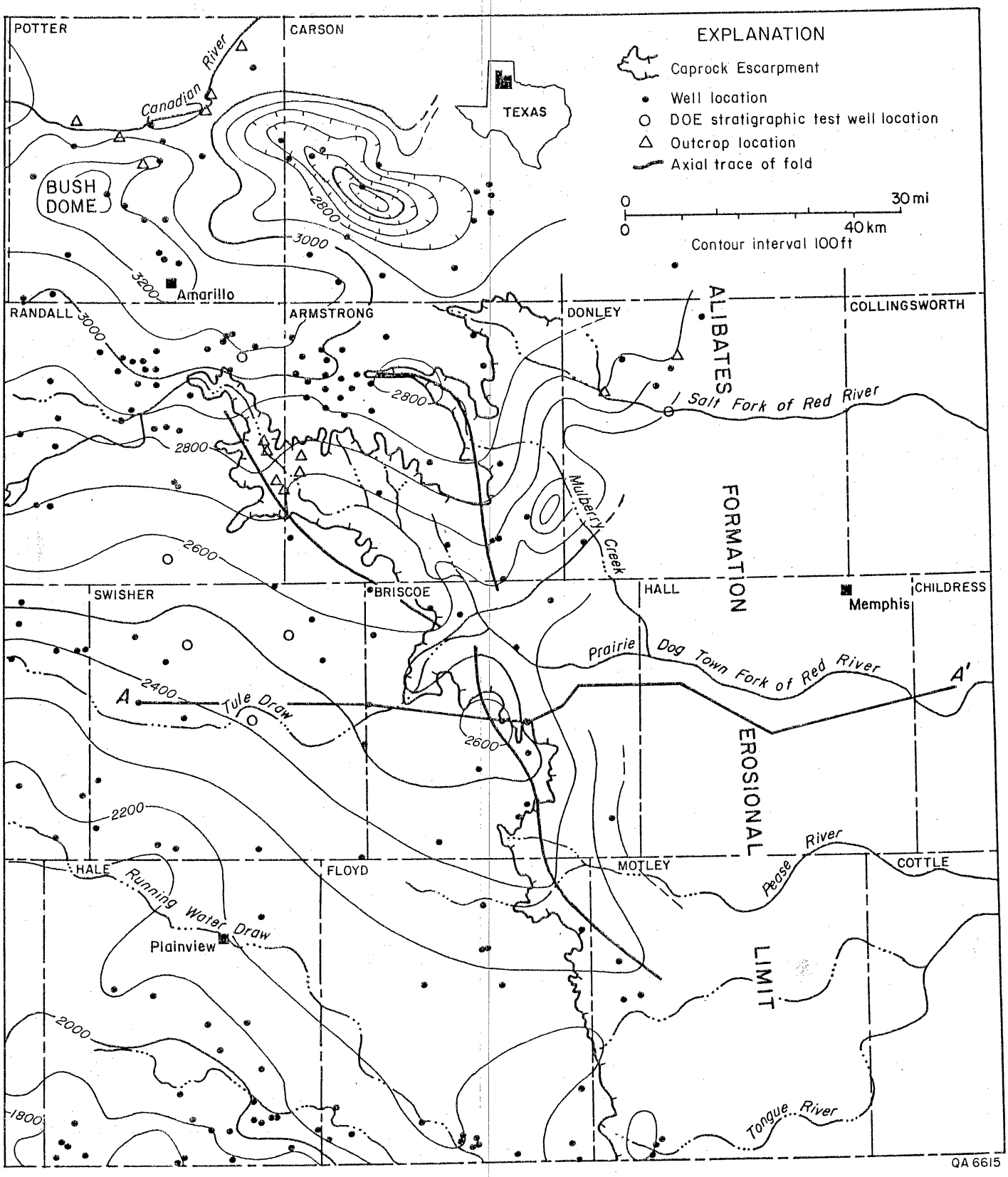


Figure 75. Structure-contour map on the upper Permian Alibates Formation at the eastern margin the Palo Duro Basin. A-A' locates figure 64.

and 70). At the level of the Alibates Formation in the northern margin of the Palo Duro Basin large closed basins occur in Oldham and Hartley Counties and in southwestern Carson County. The crest of the John Ray and Bush domes are at the same elevations, and the structural relief of the trough between the two domes has increased from 196.9 ft (60 m) on top of the Tubb interval to approximately 295.3 to 393.7 ft (90 to 120 m). At the level of the Alibates in the eastern margin of the Palo Duro Basin a regional dip reversal from southwest to east or southeast occurs.

The change in structural configuration from the Tubb interval to the Alibates Formation is apparently the result of post-Permian dissolution of salt from between the two structural horizons (compare figs. 69 and 70, figs. 71, 72, and 73, and figs. 74 and 75). For example, dissolution of approximately 590 ft (180 m) of salt from over the John Ray dome but not over the Bush Dome accounts for most the change in relief and elevation between the two features. Dissolution of Salado and Seven Rivers salt beneath and to the east of the Caprock Escarpment accounts for most of the dip reversal of the Alibates Formation.

Dissolution has occurred preferentially in the updip segments of Permian salt beds, which by virtue of their proximity to the ground surface at the structural margins of the Palo Duro Basin are easily accessible to relatively fresh ground waters.

Ground-Water Flow in the Peripheral Salt-Dissolution Zone

Downward cross-formational flow of groundwater from aquifers in the Ogallala Formation and Dockum Group drives salt dissolution (Fink, 1963; Gustavson and others, 1980; Dutton and Simpkins, 1986; Dutton, 1987a, in press, Gustavson and Simpkins, in press). The vertical hydraulic-head gradient is particularly steep beneath the Caprock Escarpment, reflecting the influences of topography on ground-water flow paths. The potentiometric surface data in figures 51 and 53 indicate that springs are

likely to develop at the base of the Caprock Escarpment where the aquifers in the Ogallala Formation and Dockum Group abruptly terminate and where flow encounters mudstones of low hydraulic conductivity in the upper Permian Formations. Closely spaced potentiometric contours in proximity to the Caprock Escarpment (fig. 51) reflect the convergence of ground-water flow paths in the region. The path of least resistance is to discharge at the boundary. Numerous springs are in fact present. Additional discharge from this flow system occurs farther to the east in the Rolling Plains where saline springs discharge a mixture of locally derived and deeper ground waters, as evidenced by geochemical and isotopic data (Richter and Kreitler, 1987a, 1986).

Figure 76 shows a conceptual model of the flow paths of low-salinity ground water through the dissolution zone in Permian bedded salt, including a steep hydraulic-head gradient imposed by the escarpment boundary as the major force driving ground-water flow downward into relatively shallow updip elements of Permian bedded salt. Low-salinity ground water discharges as springs along the escarpment. Permian salt contacted by low-salinity ground water is readily dissolved, and the resulting brines flow eastward in highly transmissive dolomite and anhydrite/gypsum beds to discharge in topographically low areas.

Effects of Salt Dissolution

Removal of several hundred meters of interstratal salt from the stratigraphic section produced widespread subsidence and collapse of overlying strata. Locally, subsidence resulted in regional troughs, synclinal folds, breccia beds, collapse chimneys, sinkholes, surface fractures, and extension fractures.

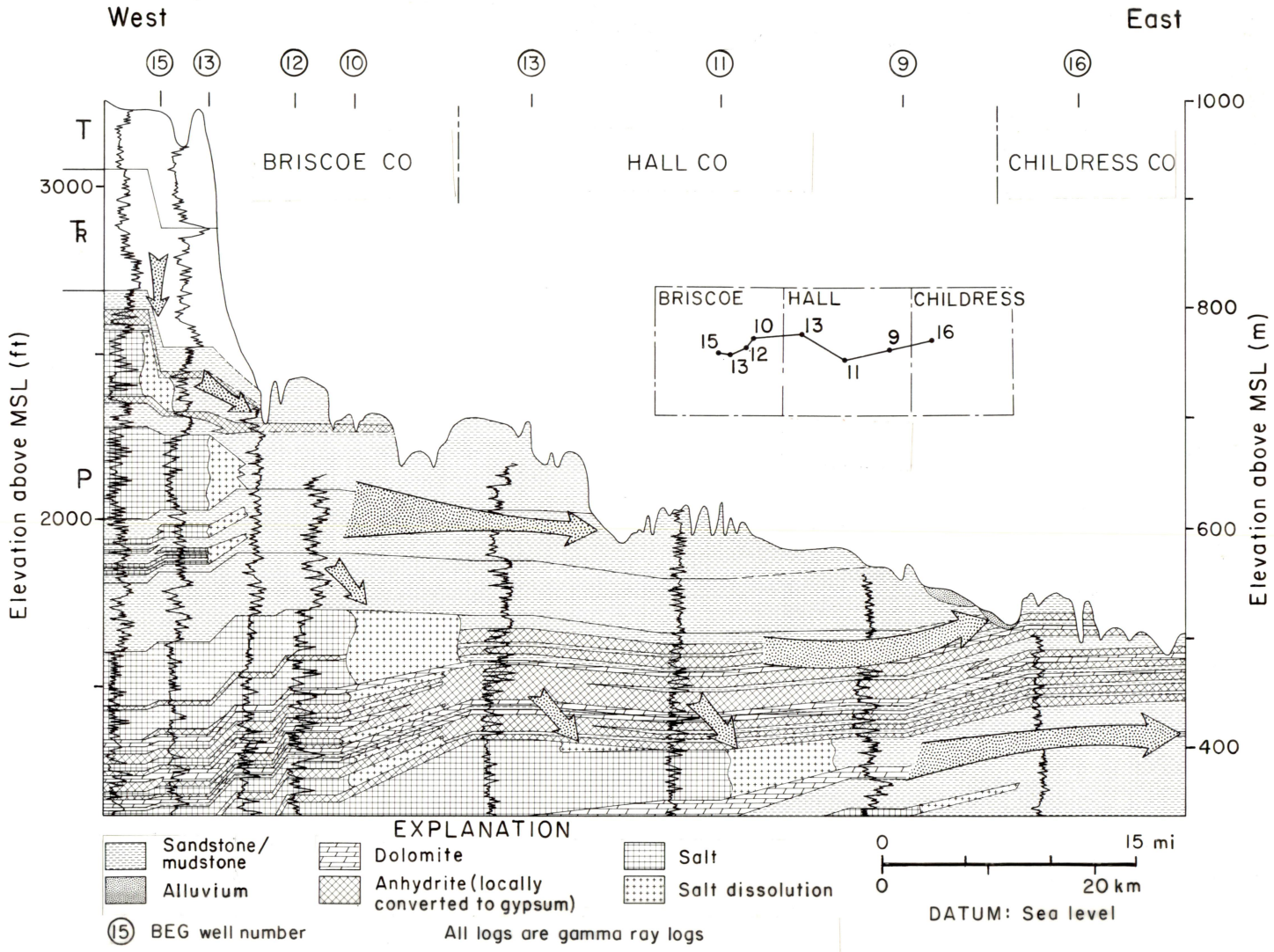


Figure 76. Conceptual model of ground-water movement and its impact on Permian bedded salt beneath the High Plains, Caprock Escarpment, and Rolling Plains. Inferred movement paths are indicated by arrows. Stratigraphic units are identified in figure 64.

Extension Fractures

Sandstones and mudstones of the Upper Permian Whitehorse and Quartermaster Formations and locally mudstones of the Upper Triassic Dockum Group in the Texas Panhandle are extensively fractured, and the fractures are commonly filled with satin spar gypsum (Gustavson, 1980, 1986b, 1986c; Goldstein and Collins, 1984; Collins and Luneau, 1986). Veins fill fractures along horizontal bedding planes, step fractures between bedding planes, vertical fractures and inclined fault planes (fig. 66). Satin spar fibers are vertical in all of these fractures except vertical fractures where fibers are horizontal. Veins are marked by a medial line. Satin spar-filled fractures are also present in cores of the Upper Permian Salado and Seven Rivers Formations taken from beneath the High Plains. From analyses of cores of upper Permian salt-bearing units it is clear that satin-spar veins only occur in strata overlying zones of salt dissolution. No evidence of these veins was found below the first salt bed in any core of the Permian section. These characteristics led Gustavson (1980) and Goldstein and Collins (1984) to attribute fractures filled with satin spar to dissolution-induced subsidence.

The fibrous texture of veins and the presence of a medial line in most satin spar veins suggests that these veins formed by the crack-seal mechanism described by Ramsey (1980) and Ramsey and Huber (1983). Fibers are built up during a series of successive small openings of the fracture followed by cementation of small increments of material derived from intergranular fluids and added to the vein (crack-seal mechanism of Ramsey (1980) and Ramsey and Huber (1983). Consequently, the orientations of the fibers mark the direction of maximum principal extension during the time that the fibers were forming. If the fibers are straight, fracture extension was unidirectional; if fibers are curved, as is sometimes observed, extension was accompanied by shear. However, examination of veins in thin section and outcrops

reveals only a few examples of incremental deposition of satin spar in veins such as the trails of inclusions of wall rock fragments described by Machel (1985). Fracture opening and fiber growth was either slow and continuous in most places, or if it was episodic, trails of wall rock fragments never developed.

Synclinal Folds, Dolines, and Surface Fractures

A large study encompassing all of Hall County, Texas, was undertaken to examine the relation between the processes of dissolution and subsidence or collapse and the karst topography that is developing in the Rolling Plains (Gustavson and others, 1982). In addition to numerous sinkholes, more than 200 internally drained, closed depressions (dolines) were recognized (fig. 77). These are broad, shallow, typically oblong depressions as long as 2.2 mi (3.5 km) and as deep as 35 to 50 ft (10 to 15 m). Many of these elongate depressions show preferential northwesterly, northerly, and northeasterly alignments (fig. 78). Two dolines, and at least 36 new sinkholes, formed in the county between 1940 and 1972. Dolines and sinkholes are surface expression of collapse chimneys, which are discussed in the following section.

The physiography of Hall County is characteristic of much of the rest of the Rolling Plains in the Texas Panhandle. Two major physiographic units are distinguished: (1) dissected areas of low to moderate relief underlain primarily by Permian strata and (2) areas underlain by low-relief relatively undissected Quaternary fluvial and eolian deposits of the Lingos Formation (Caran and Baumgardner, in press). Landforms resulting from dissolution and collapse are recognized primarily in undissected areas underlain by the Lingos Formation.

The shapes of the depressions depicted clearly illustrate preferred orientations of northwest-southeast, north-south, and northeast-southwest (fig. 77). These orientations correspond closely to the orientations of fractures within Triassic and

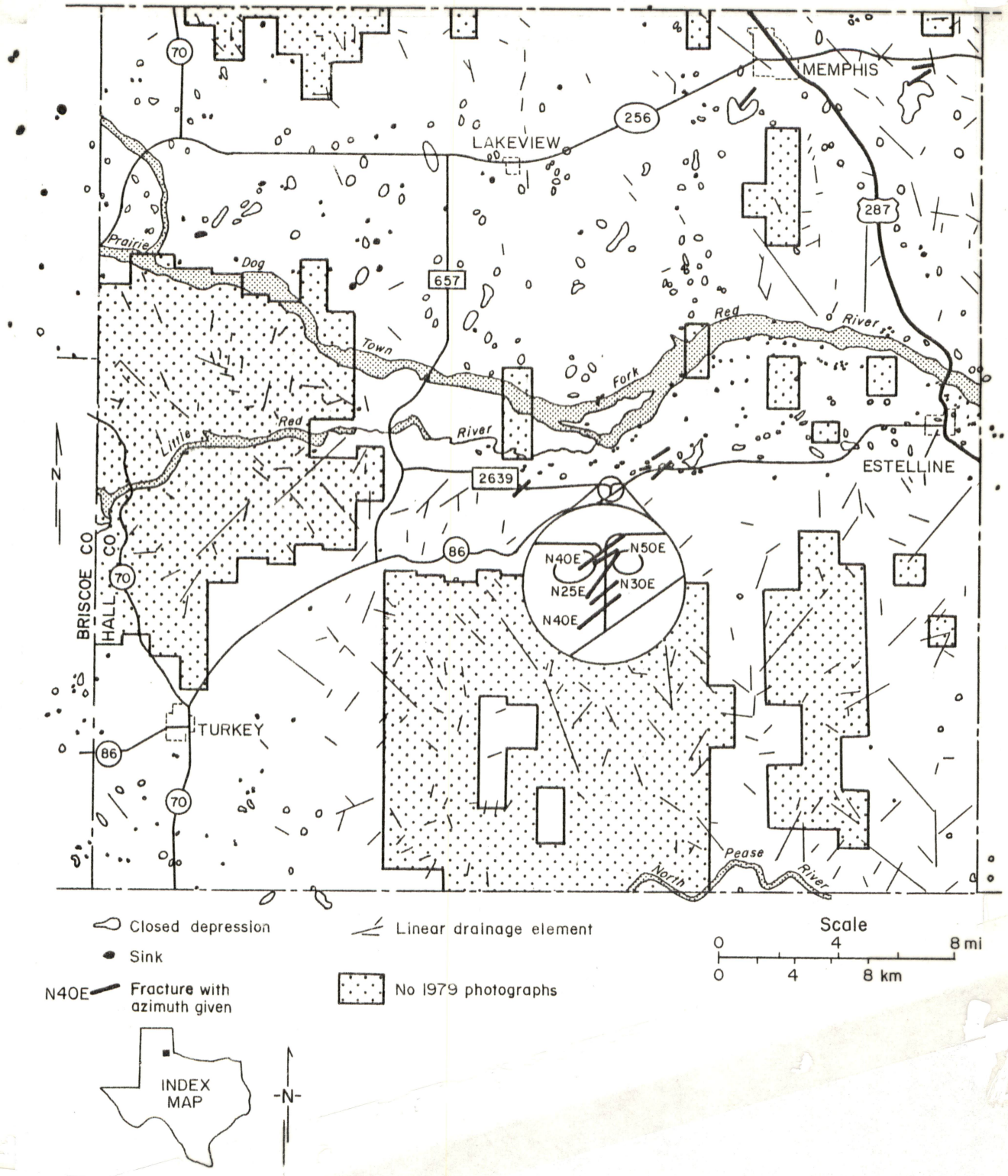


Figure 77. Location of sinkholes, closed depressions (dolines), and fractures in Hall and eastern Briscoe Counties, Texas. Dolines are drawn to scale; sinkholes, which are much smaller, are not drawn to scale. Dolines and sinkholes were recognized from color aerial photographs taken in 1979.

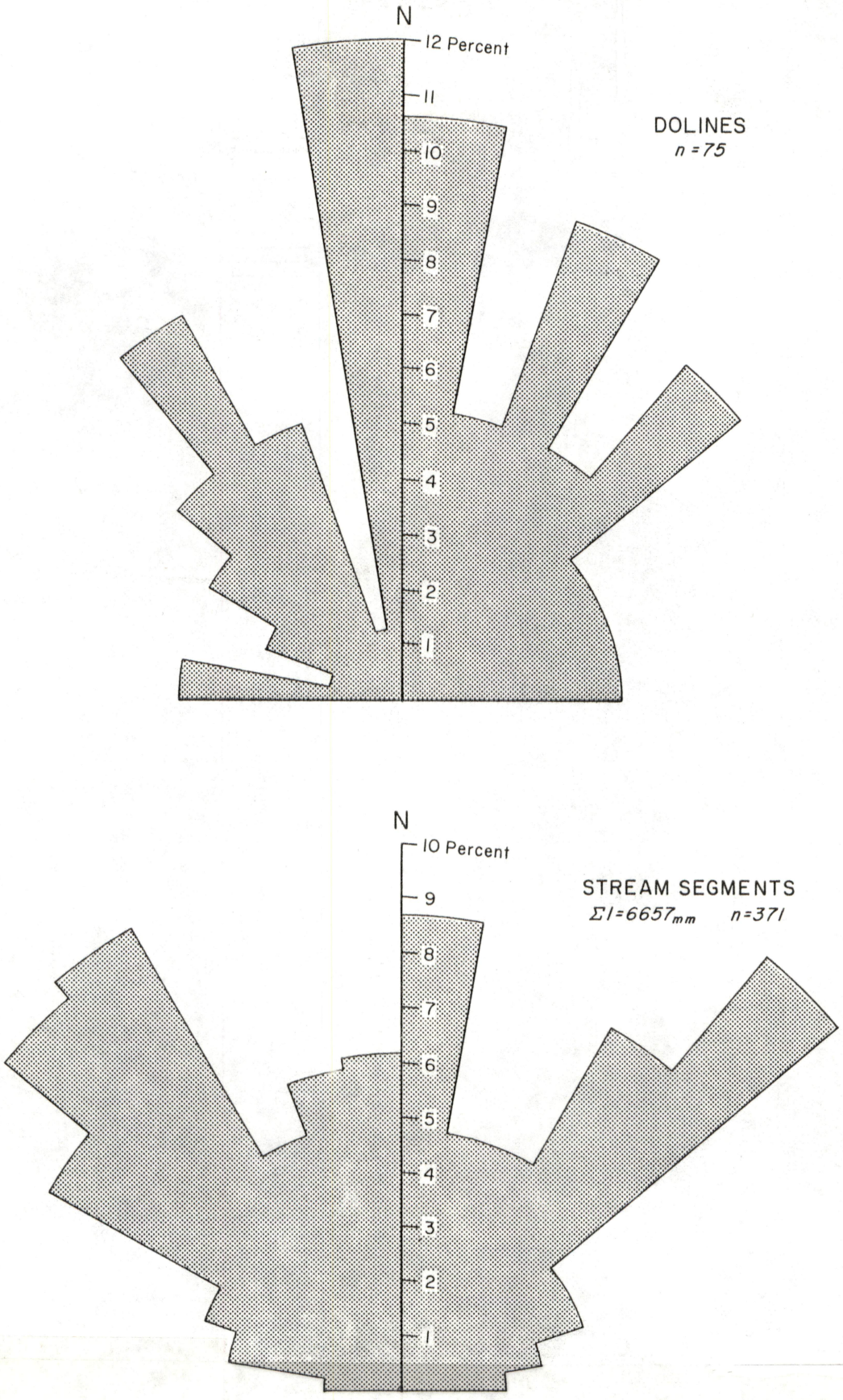


Figure 78. Diagrams indicating orientations of long axes of dolines and linear stream segments in Hall County, Texas. For each 10^{O} sector, linear data are plotted as a percentage of the total number of closed depressions and as a percentage of the total length of linear stream segments.

Permian rocks exposed to the west along the western margin of the Rolling Plains and along the Caprock Escarpment (Finley and Gustavson, 1981; Collins and Luneau, 1985). In areas of exposed Permian bedrock in Caprock Canyons State Park, a few kilometers west of the area covered by the Hall County study, Collins (1984) described the relation between joints and synclinal depressions. He concluded that dissolution and collapse developed preferentially along joint paths and caused the synclines. It is thought that the elongated, oriented depressions (dolines) observed on the surface of the Lingos Formation are underlain by synclines similar to those observed by Collins (1984).

Open fractures at the surface are difficult to observe over a large region because they tend to be covered through natural causes or are infilled by land owners in agricultural areas such as the Rolling Plains. However, evidence of surface fractures were noted along State Highway 6239 as well as several other are in Hall County (fig. 77). Two of the open surface fractures observed in Hall County were open intermittently between 1979 and 1984 to widths of 12 in (30 cm) where they crossed cultivated fields. The locations of four previously open tension fractures were indicated by diagonal patches across State Highway 2639; the open part of these fractures in adjacent fields had been filled by land owners. Fractures that cross State Highway 2639 strike from N.25°E. to N.50°E. and are nearly parallel an adjacent series of closed depressions that strike N.40°E.

To determine the region outside Hall County where karst features were being formed, questionnaires with photographs of sinkholes and open fractures were sent to soil scientists of the U.S. Department of Agriculture, Soil Conservation Service, the the Agricultural Stabilization and Conservation Service, and to State highway maintenance supervisors in the region. This was done for each county of the Texas Panhandle and for the two eastern tiers of counties in New Mexico. Soil scientists from 21 of the counties within the Pecos River valley, Canadian River valley, and

Rolling Plains reported that sinkholes had formed within their county (fig. 79). Highway maintenance supervisors from 9 of these 21 counties also reported that sinkholes had formed in the highway rights-of-way in their areas of jurisdiction (Simpkins and others, 1981). Neither sinkholes nor fractures were reported for counties in Texas and New Mexico that lie entirely within the Southern High Plains.

There is only a single historical example of collapse to form a sinkhole on the High Plains. The Cuneva Basin is a large, flat-floored depression in Quay County, New Mexico (T7N R28E) at the northwestern edge of the High Plains, that is approximately 5 mi (3.2 km) wide and 75 ft (22 m) deep (Judson, 1950). A sinkhole about 300 by 150 ft (91 by 46 m) wide and 40 to 60 ft (12 to 18 m) deep formed in the northwest corner of the depression early this century. The most recent collapse occurred in 1934, about 10 to 15 years after initiation of the feature. Cuneva Basin occurs over an area of abrupt thinning of Permian Bernal-Artesia Group strata in Quay and Curry Counties, New Mexico. Thinning of these units was attributed to salt dissolution (Gustavson and Finley, 1985). Thus, it is reasonable to infer, as Judson (1950) did, that the recent sinkhole and probably the Cuneva Basin are collapse features caused by dissolution.

Gustavson and others (1980) argued that large playas (>1.5 mi [>2.4 km] in diameter) along the northern margin of the Southern High Plains either resulted from or were modified by salt dissolution. They showed two examples of playa basins underlain by thin salts in the Blaine Formation and evidence of collapse beneath the basin (fig. 80). Ongoing studies of playa basins on the High Plains indicates that relationship between subsidence caused by salt dissolution and playa basins of any size is at best ambiguous. A few large playa basins are underlain by areas of thin salt and structural subsidence, but most are not.

The general absence of sinkholes and fractures in the Southern High Plains indicates that recent catastrophic subsidence or collapse as reflected by the formation

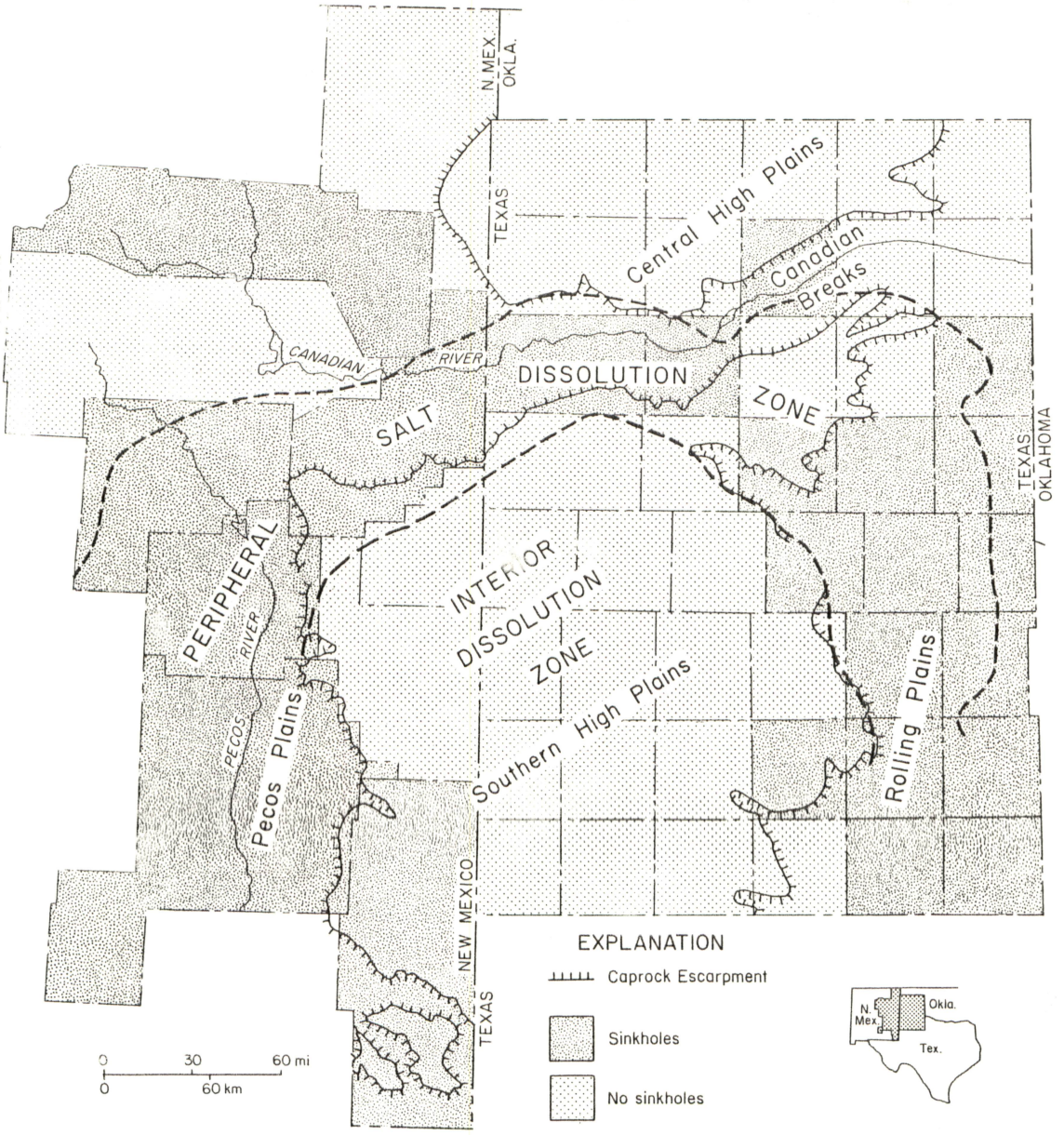


Figure 79. Texas and New Mexico Counties from which sinkholes and open surface fractures have been reported lie mostly within a peripheral dissolution zone that encompasses the Pecos Plains, the Canadian River Breaks, and the Rolling Plains. Sinkholes and open fractures have not been reported from the Southern High Plains or the Central High Plains within eastern New Mexico or the Texas Panhandle.

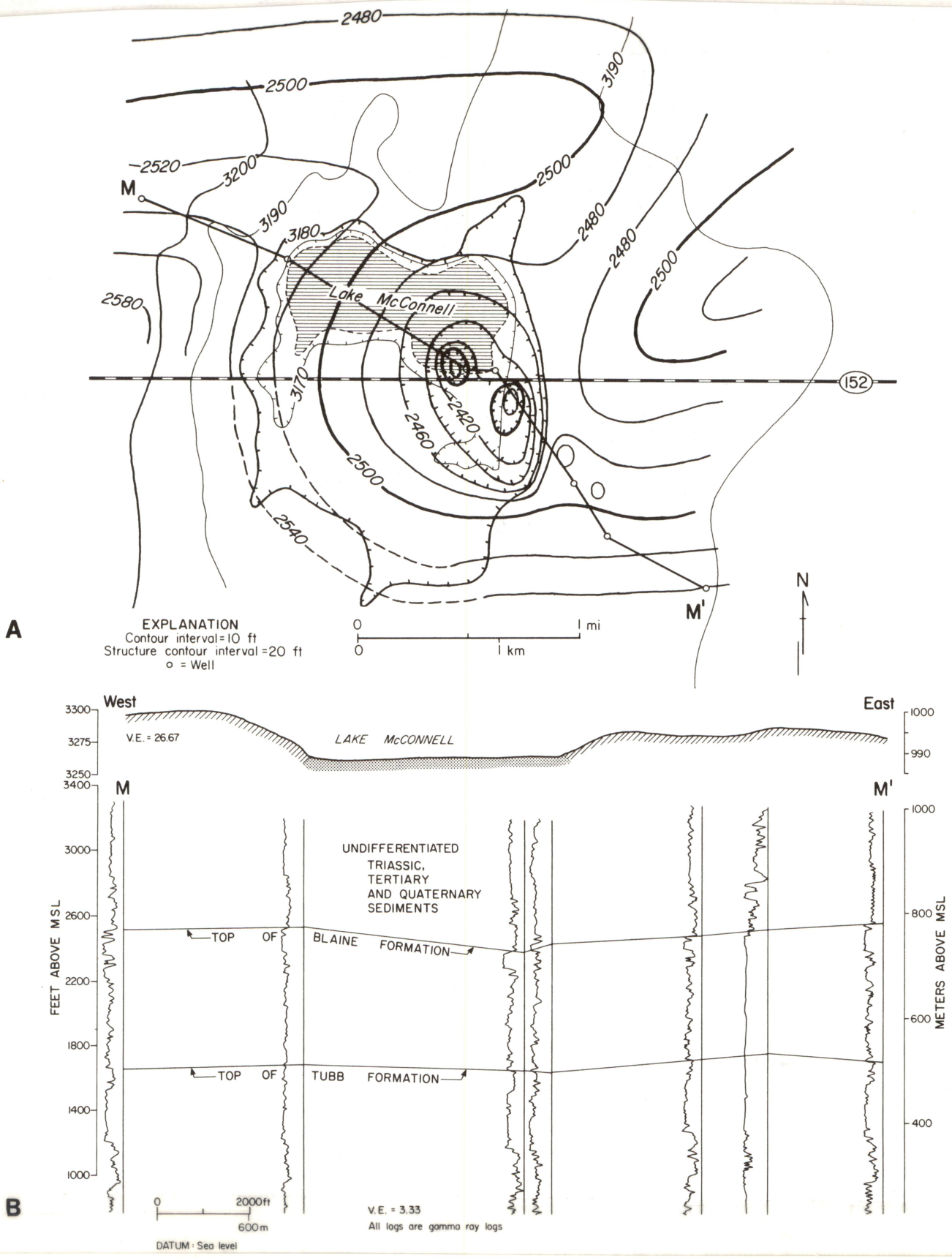


Figure 80. (a) Topographic map of the Lake McConnell playa superimposed on a structure-contour map on the top of the Blaine formation. The depression on the structure-contour surface (heavy contour lines) underlies the playa depression. (b) Structural cross section through the Lake McConnell playa derived from gamma-ray logs.

of sinkholes and open surface fractures is essentially restricted to areas of relatively rapid salt dissolution, the peripheral dissolution zone. Gustavson and others (1980a) previously thought that the lack of development of sinkholes and fractures on the Southern High Plains indicated that salt dissolution was inactive there. However, additional analyses of core from the SWEC Zeeck No. 1, SWEC Harman No. 1, SWEC J. Friemel No. 1, SWEC G. Friemel No. 1, and SWEC Detten No. 1 wells, all of which occur on the Southern High Plains, show clearly that dissolution of Upper Permian salt has been widespread. In core from these wells, salt of the Salado and Seven Rivers Formations are overlain by insoluble residues composed of anhydrite, dolomite, and mudstone, as well as complexly fractured rock with slickensides on fracture surfaces and extension fractures filled with satin spar. It seems more likely that the absence of sinkhole development suggests only that dissolution beneath the Southern High Plains has not developed the cavernous conditions necessary for catastrophic collapse.

Breccia Beds

Breccia beds are evidence of the former presence of caverns or systems of caverns in Upper Permian strata exposed along the Little Red River in Caprock Canyons State Park (Gustavson, 1986c). The coarse breccia consists of gypsum clasts as much as 12 in (30 cm) long in a mudstone or gypsum matrix (fig. 68). The breccia is as much as 3 ft (1 m) thick and one or more breccias are exposed along the Little Red River over a distance of at least 0.3 mi (0.5 km). Clasts of breccia have been rotated, indicating that the breccia was formed by gypsum or anhydrite blocks that fell into an open space or cavern. Undeformed gypsum beds both underlie and overlie the breccia bed. Neither gypsum clasts nor the underlying or overlying gypsum beds show any evidence of dissolution. The gypsum beds lie within

the Upper Permian Whitehorse Formation and are probably correlative with the Salado-Tansill Formation in the subsurface.

Locally, coarse gypsum (var. selenite) crystals are associated with the breccia bed. In certain areas the gypsum crystals are intergrown and have completely filled the original cavity. Pods of coarsely crystalline gypsum crystals are several tens of centimeters thick and several tens of meters long. In other areas selenite crystals occur in open cavities.

Selenite crystals apparently grew in areas of the cavity not filled with gypsum breccia. Areas where fully terminated selenite crystals occur are apparently remnants of the original cavity. The cavern probably resulted from dissolution of a salt bed that formerly lay between the two undisturbed gypsum beds. The lack of evidence of gypsum dissolution indicates that this process did not contribute significantly to cavern development.

Collapse Chimneys

Chimneys filled with collapse breccia (collapse chimneys) are common in exposures of Permian strata in the Canadian River valley northeast of Amarillo and in the valley of the Salt Fork of the Red River east of Amarillo. Collapse chimneys are the subsurface extension of surface sinkholes or dolines. Twenty-seven filled collapse chimneys were mapped during construction of the Sanford Dam on the Canadian River (fig. 81) northeast of Amarillo (Eck and Redfield, 1963; Collins, 1984). They are circular to elliptical in cross section and are filled with slumped and brecciated sediments from the overlying Ogallala Formation or from Canadian River alluvium. The largest chimney exposed is approximately 1,000 ft (305 m) in diameter. Characteristically chimneys consist of an outer zone of slumped or downwarped Permian sediments, a central collapse breccia of Permian sediments, followed by a core of

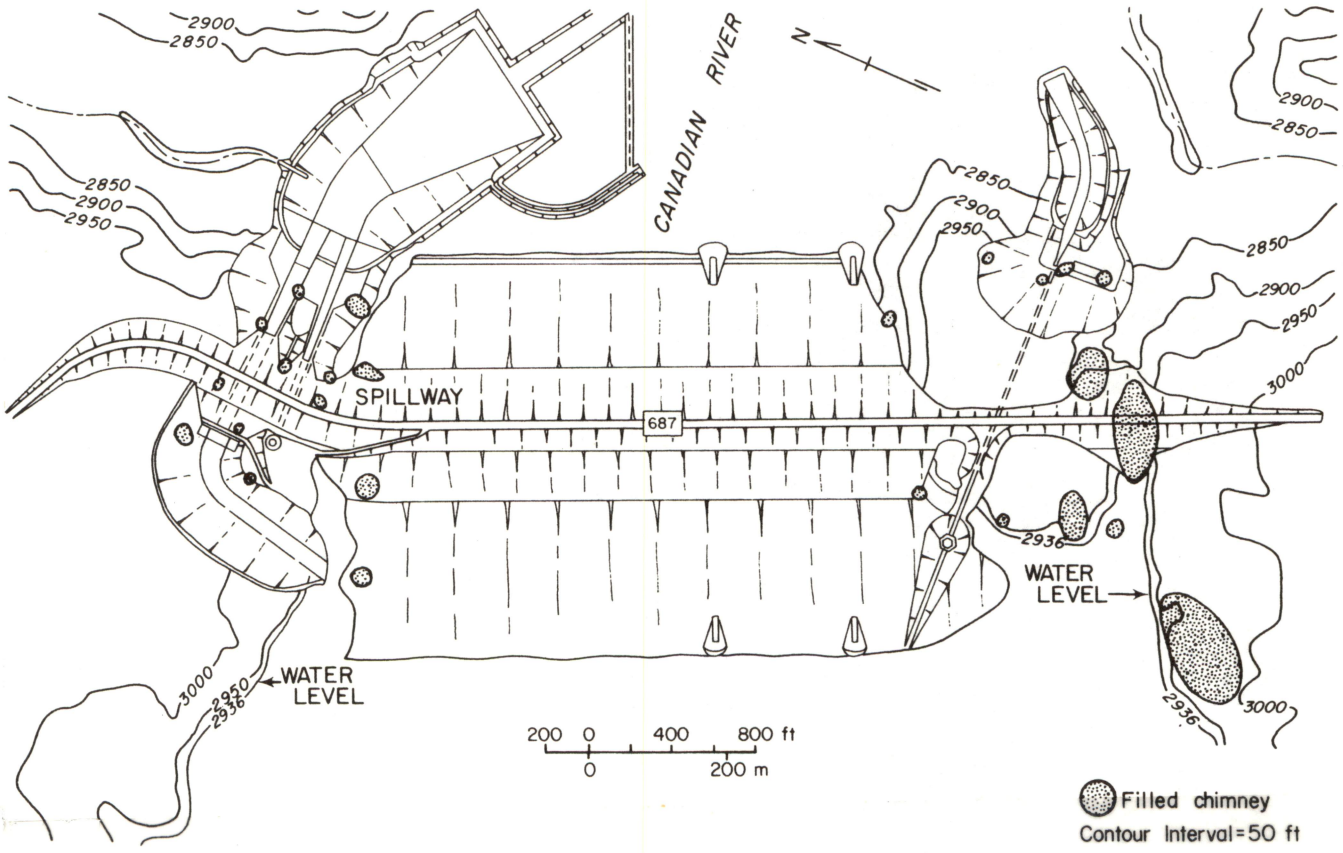


Figure 81. Distribution of chimneys in the vicinity of Sanford Dam on the Canadian River, Texas. From Eck and Redfield (1963).

Tertiary, or Quaternary sediments (fig. 67). Chimneys are the product of collapse of strata into caverns by a process of natural upward stoping or roof fall. Collapse in most cases is not a single event but rather a series of events (see Landes [1945] and Davies [1984] for a discussion of the process of chimney formation). If the natural stoping process continues to the surface a sinkhole or doline results.

Eck and Redfield (1963) attributed formation of collapse chimneys to subsurface dissolution of gypsum in the Blaine Formation. Examination of cross sections through the region reveals little evidence of gypsum dissolution while apparently several hundred feet of salt have been dissolved from beneath the Sanford Dam region (McGookey and others, 1988; Gustavson, in press). This suggests that dissolution of salt is the most likely cause of subsidence and collapse in the region.

Regional Subsidence Troughs

Regional subsidence troughs underlie segments of the three largest streams draining the Palo Duro Basin area; the Pecos River in eastern New Mexico, the Canadian River in eastern New Mexico and Texas Panhandle and the Prairie Dog Town Fork of the Red River in the Texas Panhandle (Gustavson and Budnik, 1985; Gustavson and Finley, 1985; Gustavson, 1986a). The stratigraphic sections in figures 62 to 64 illustrate the structural configuration of beds along the western, northern, and eastern margins of the Palo Duro Basin. In each case the loss of as much as 600 ft (180 m) of salt has resulted in subsidence troughs above the updip limits of salt beds. These subsidence troughs are not present in beds that underlie salt strata and consequently are not the result of tectonic movements.

Subsidence troughs have radically altered the physiography of the eastern New Mexico-Texas Panhandle region. Because of subsidence induced by salt dissolution along the western, northern, and eastern updip margin of the Palo Duro Basin, the

Pecos, Canadian, and Red Rivers flow at high angles to regional slope. Regional slope was initially to the east and southeast during the late Tertiary when fluvial facies of the Ogallala Formation were being deposited (Gustavson and Winkler, 1988). Elements of the southeasterly flowing Tertiary paleodrainage system, which are the paleovalleys that contain Ogallala fluvial sediments, are preserved on the mid-Tertiary erosional surface (Gustavson and Finley, 1985). Furthermore the gravels that make up Ogallala fluvial sediments exposed in the Caprock Escarpment contain metamorphic, volcanic, and igneous rocks that originated in the Southern Rocky Mountains of New Mexico. Ogallala exposures in the Caprock Escarpment are separated from the source beds of these sediments by the Canadian River and Pecos River valleys. The surface of the High Plains and all modern drainage elements except one slope to the southeast of east. The Pecos River flows south in the subsidence trough along the western margin of the Palo Duro Basin, the Canadian River flows east and northeast in the subsidence trough along the northern margin of the basin, and segments of the Prairie Dog Town Fork of The Red River flow northeast and then south along a subsidence trough along the northeast margin of the basin. These three drainage systems account for approximately two-thirds of the margin of the High Plains above the Palo Duro Basin.

Rates of Salt Dissolution

Rates of salt dissolution have been determined for the eastern margin of the Palo Duro Basin only. The average annual solute discharge for streams draining this area was 1.5 x 10^5 tons (1.4 x 10^5 t) between 1969 and 1983, including approximately 1.0 x 10^5 tons (0.9 x 10^5 t) of dissolved halite and 5.0 x 10^4 tons (4.6 x 10^4 t) of dissolved gypsum (U.S. Geological Survey, 1969-1983). Dissolved gypsum appears to be primarily derived from the hydration of anhydrite to gypsum without an increase in

volume and from near-surface dissolution along joint planes and outcrop faces. Dissolution of salt on the other hand is an interstratal process occurring at depths as great as 1,000 ft (300 m).

Previous discussions of salt dissolution rates were given by Swenson (1974) for a part of the headwaters of the Salt Fork of the Brazos River, and by Gustavson and other (1980) for all the gauged streams draining the eastern Texas Panhandle from the Canadian River on the north to the Brazos River on the south. Early estimates of dissolution rates were based on three to nine years of discharge data for each drainage basin. Horizontal dissolution rates for the 15 drainage basins in the eastern Texas Panhandle calculated by Gustavson and others (1980) varied over three orders of magnitude. To avoid some of the variability seen in earlier calculations Gustavson and Simpkins (1988) analyzed solute discharge data from down-stream gauging stations using 8 to 14 years of data. The estimated cross-sectional area of salt undergoing dissolution was divided into the volume of salt being transported by streams draining the region (see Gustavson and Simpkins [1988] for the method of calculating dissolution rates).

The rate of salt dissolution along the eastern margin of the Palo Duro Basin from 1969 to 1982 proceeded at average of 0.66 in/yr (0.68 cm/yr) based on salt-water discharge rates. However, among the drainage basins rates of dissolution varied from 0.32 in/yr (0.82 cm/yr) for the Salt Fork of the Red River to 29.6 in/yr (75.31 cm/yr) for the relatively small basin of the North Wichita River. These rates of dissolution are conservative estimates because they are based on solute loads carried by surface waters. No solute-load or discharge data are available for ground water discharging out of this area. Discharge rates are assumed to reflect a steady-state balance with long-term rates of dissolution of salt beds at depths of several hundred feet.

Timing of Dissolution

Dissolution of bedded Permian salt has occurred at least periodically since the Permian. Solution pits that developed shortly after deposition, and clearly before the following cycle of deposition, are present in most if not all of the evaporite depositional cycles described by Fracasso and Hovorka (1986). Granada (1981) attributed thick linear accumulations of Triassic Dockum Group sediments near the southeastern margin of the present Southern High Plains to sedimentation into troughs that formed as a result of salt dissolution. Jurassic rocks are missing from all but a very small part of the Palo Duro Basin area and display no evidence of dissolution-induced subsidence. Cretaceous marine rocks also occur only in a small area of the Palo Duro Basin and show no recognized evidence of dissolution. By late Tertiary time, however, salt dissolution had become an important process in the Texas Panhandle.

Lacustrine sediments of the Ogallala Formation containing Clarendonian land mammal age fossils fill a series of sinkholes attributed to salt dissolution north of Clarendon, Texas (Schultz, 1977). Ogallala fluvial sediments carrying Clarendonian fossils also occupy a series of paleo-subsidence valleys along the southern side of the Canadian River valley (Gustavson, 1986a; G. A. Wilson, personal communication 1988). Lacustrine sediments that include Blancan land mammal age fossils occupy a large subsidence basin that developed along Rita Blanca Creek north of the Canadian River in Hartley County, Texas (J. Schmude, personal communication, 1987; Gustavson, 1986a). Terrace deposits with the Lava Creek B volcanic ash (approximately 620,000 yr BP) and in close proximity to collapse chimneys have been tilted as much as 30 degrees along the southern margin of the Canadian River valley. Dissolution and subsidence during the early Quaternary also was responsible for the development of a series of lacustrine basins including the Tule Basin in Swisher and Briscoe Counties (Gustavson, 1986b; Schultz, 1986), a basin along Mulberry Creek in

Armstrong County (Evans and Meade, 1945; Gustavson and Simpkins, 1988), a series of basins along Frio Draw, and Tierra Blanca Creek in Deaf Smith and Randall Counties (Gustavson and Budnik 1985; Gustavson, in press). Other lacustrine basins in the High Plains region that resulted from subsidence due to salt dissolution were described recently by Reeves and Temple (1986), Johnson (1988), and Caran (1988). Brines from hydrologic test wells beneath the southern margin of the Canadian River valley and from beneath the Rolling Plains have been interpreted to be no older than 23,500 and 16,200 yr respectively (Dutton, 1987a).

Recent examples of natural subsidence have occurred all along the margins of Southern High Plains (fig. 79) and at least 36 sinkholes have formed in eastern Hall County since 1940 (Gustavson and others, 1982). Collapse sinkholes formed earlier this century in the Cuneva Basin in eastern New Mexico and in Meade County, Kansas (Judson, 1950; Johnson, 1901). The most convincing evidence that salt dissolution is active along the margins of the Palo Duro Basin is the high solute loads carried by streams that drain the Great Plains of western Texas and eastern New Mexico: more than 2.5 x 10^5 tons (>2.3 x 10^5 t) of salt in the Pecos River annually, more than 5 x 10^4 tons (>4.5 x 10^4 t) of salt in the Canadian River annually, and more than 1 x 10^6 tons (>9.1 x 10^5 t) of salt in the Red River annually (Gustavson and Finley, 1985).

Summary

A broad area of active dissolution affects salts of the Permian Salado, Seven Rivers, San Andres, Glorieta and Clear Fork Formations in the Palo Duro Basin of northwestern Texas and eastern New Mexico. Dissolution and subsidence have been relatively rapid along the structural margins of the basin resulting in large-scale subsidence features including synclinal folds, dolines, sinkholes and fractures. This area of

dissolution-induced subsidence along the western, northern, and eastern structural margins of the Palo Duro is termed the peripheral dissolution zone, and is reflected at the surface in part by the Pecos River valley, the Canadian River valley, and the western Rolling Plains. In the interior of the Palo Duro Basin dissolution has been relatively slow and has not resulted in large-scale collapse features with the exception of certain lacustrine basins. Breccia beds and extension fractures filled with fibrous gypsum are characteristic of both the interior and peripheral dissolution zones.

Dissolution has occurred in this area since salts were deposited in the late Permian. However, regional uplift during the late Cretaceous and Tertiary initiated the hydrogeologic conditions that prevail today. Streams draining sediments exposed at the margins of the Palo Duro Basin carry a combined solute load of approximately 1.24×10^5 tons (1.1×10^5 t) of halite per year; strong evidence that dissolution is an ongoing process.

Unresolved Issues

The circulation of ground water and the distribution and rate of salt dissolution beneath the Southern High Plains remain poorly understood, although salt-dissolution processes beneath the Pecos and Canadian River valleys and the Rolling Plains are better known. Data from hydrogeologic and geochemical studies (Dutton, 1987a, in press) and geomorphic studies (Gustavson, 1986, in press; Gustavson and Budnik, 1985; Gustavson and Finley, 1986; Gustavson and Simpkins, in press; Reeves and Temple, 1986) suggest that dissolution of Permian bedded salt has occurred locally during the Quaternary beneath the Southern High Plains. However, the extent of Quaternary dissolution is not known, nor is it known unequivocally whether dissolution beneath all parts of the Southern High Plains is presently active. The presence of large lacustrine basins and other subsidence features on the Southern High Plains are

strong evidence that subsidence has resulted from dissolution of Permian bedded salt beneath the Southern High Plains. Although it is obvious that ground water, which is unsaturated with respect to sodium chloride, dissolved salt, it is not obvious why many subsidence basins on the Southern High Plains developed where they did.

CONCLUSIONS

Major uncertainties in stratigraphy and basin history remain in three broad and overlapping categories: (1) depositional processes, (2) diagenetic processes, and (3) tectonic processes. General depositional models have not been worked out for the Wichita Group and the Seven Rivers Formation, which make up significant parts of the evaporite section. Although the San Andres Formation probably serves as valid model for these and other Leonardian to Ochoan deposits, careful examination of these other units would help improve understanding of the processes and controls of cyclic evaporite sedimentation on stable platforms. Knowledge of the Wolfcamp Series carbonates, which comprise the main permeable strata in the deep-basin aquifer system, is still rudimentary. Basic models for the depositional setting of these complex and diverse deposits have been presented but detailed studies of paleoenvironments, lithologic and mineralogic variability, and their effect on the distribution of porosity and permeability are incomplete. Exceptionally detailed studies of the diagenesis of evaporites in the San Andres have documented a very complex depositional and diagenetic history for these deposits. In comparison, studies of effects of diagenesis on other parts of the Paleozoic sequence have been neglected. Analysis of the diagenetic alteration of lower Permian carbonate (for example, the Wolfcamp and Wichita) is especially needed in developing predictive models of ground-water flow based on accurate porosity and permeability relationships.

There is also a great need for studies directed toward establishing a chronostratigraphic framework for the Paleozoic of the Palo Duro Basin, which would help resolve issues such as exact timing of tectonic events and their effect on depositional patterns and rock diagenesis. The relationship between deep-seated tectonic features and salt dissolution at the top of the Permian section also is not fully understood.

Stratigraphy of Triassic sandstone and mudstone and the degree of hydrologic connection between the Triassic and Permian deposits should be better defined to help determine the amount of downward cross-formational flow of ground water. The lithologic transition between Permian and Triassic rocks in the basin appears gradational and may include Early Triassic deposits. Only Late Triassic deposits, however, occur around the margins of the basin, suggesting a significant unconformity. If Early Triassic deposits are not present within the basin, the existing depositional and tectonic model for the Dockum Group requires reevaluation.

The Ogallala Formation constitutes the main part of the region's significant High Plains aquifer, from which drinking water and irrigation water for agriculture are obtained. Because Ogallala deposits of this region represent a long interval of geologic time and underlie a vast area, the depositional and tectonic history is undoubtedly much more complex than has been generally perceived. Interpretation of the stratigraphy of the Ogallala, Blackwater Draw, Blanco, and Tule Formations has been evolving in recent years with new studies of outcrops and application of depositional models. The stratigraphic relationships between these formations are still incompletely known. The effect of paleoclimatic variation on surficial sediments and geomorphic processes during the Quaternary has not been completely investigated. The origin of playa lakes perched above the water table, saline lakes that intersect the water table, and of Quaternary lake basins is controversial. Only a sedimentary record remains of Quaternary lake basins. The origin of lake basins on the Southern High Plains needs to be further studied to document (1) the relationship between lacustrine basins and underlying salt dissolution, (2) paleoclimatic variation, and (3) the nature of focused recharge of ground water throughout the Southern High Plains.

Foremost among the several remaining unresolved issues regarding the petroleum potential of the Palo Duro Basin is the question of whether or not source rocks in the basin are sufficiently mature to have generated significant quantities of

hydrocarbons. Perhaps the most compelling evidence that parts of the basin have become mature enough to generate hydrocarbons is the discovery of several new, small oil and gas fields. Although most discoveries have been along the margins of the basin, the recent discovery of what appears to be a sizeable oil reservoir in the center of the basin suggests that there could be significant hydrocarbon accumulations throughout the basin. This discovery will no doubt rekindle interest in further hydrocarbon exploration in the basin. Additional studies of the organic geochemistry of oil and source rock in the basin must be undertaken to refine our knowledge of the hydrocarbon potential in the basin.

Uncertainties in hydrogeologic conceptual models include (1) amount and mechanism of leakage through the evaporite confining system, (2) stratigraphic, structural, and hydrologic boundary conditions for the deep-basin brine aquifers, (3) mechanism and rate of recharge to the High Plains aquifer, including the role of playa lakes in focusing recharge and the nature of transport processes in the thick vadose zone overlying the water table, (4) age and residence time of ground water in the Dockum Group and salt-dissolution zones, (5) extent and effect of stratigraphic heterogeneities on uniformity of flow and flushing of connate water in all hydrogeologic units, and (6) analysis of flow potential where water density varies significantly because of salinity. These uncertainties result partly from scarcity of geohydrologic data and the spatial variability of hydrogeologic parameters, but also from the incomplete description of the stratigraphic and diagenetic framework of the lithologic sequence. Data are least abundant and uncertainty in hydrologic properties and continuity of hydrogeologic units is greatest for hydrogeologic units in the evaporite confining system. Confidence in results of ground-water modeling is limited by these uncertainties and by inadequate testing of assumptions. Regional models developed to date ignore effects of variations of fluid density on flow velocity and assume that the present ground-water flow system is essentially at steady state. Continued growth in the stratigraphic and

hydrogeologic data base, refinement of hydrogeologic conceptual models, and improvement in modeling techniques dictate that numerical models be updated.

A broad area of dissolution affects salts in Permian Salado, Seven Rivers, San Andres, Glorieta and Clear Fork formations in the Palo Duro Basin of northwestern Texas and eastern New Mexico. The distribution, timing, and processes related to dissolution beneath the Southern High Plains remain poorly understood, although data from so-called peripheral salt-dissolution zones are only a little more abundant. Extent of Quaternary dissolution beneath the Southern High Plains is not known, nor is it known unequivocally whether dissolution is currently occurring throughout that area. Dissolution has been progressing downdip in the formations since salts were deposited in the late Permian. However, regional uplift during the late Cretaceous and Tertiary initiated the hydrogeologic conditions that prevail today. Dissolution and subsidence have been relatively rapid along the structural margins of the basin (the so-called peripheral dissolution zone) resulting in large-scale subsidence features including synclinal folds, dolines, sinkholes and fractures. The peripheral dissolution zone is reflected at the surface in part by the Pecos River valley, the Canadian River valley, and the western Rolling Plains. In the interior of the Palo Duro Basin dissolution has been relatively slow and has not resulted in large-scale collapse features, with the exception of certain lacustrine basins. It is not obvious why many subsidence basins on the Southern High Plains have locally developed and why subsidence in other areas has not occurred.

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