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EROSIONAL MARGINS AND PATTERNS OF SUBSIDENCE IN THE LATE PALEOZOIC WEST TEXAS BASIN AND ADJOINING BASINS OF WEST TEXAS AND NEW MEXICO

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Abstract—The West Texas Basin is a complex late Paleozoic basin on the unstable craton. It is a composite of Early Pennsylvanian and Early Permian deformation and Early Pennsylvanian through late Permian subsidence. The postdeformational bowl of subsidence of the West Texas Basin is broadly similar to the subsidence of true intracratonic basins, such as the Michigan and Williston Basins. Unlike these basins, however, the present boundaries of the West Texas Basin do not follow or preserve the original limits of subsidence. The southern, western and to a lesser degree the eastern margins have been altered by pre-Albian uplift and erosion, assisted by Laramide and Tertiary uplift on the western margin. Only the northern margin is preserved, although it is complicated by the neighboring Anadarko Basin. The Pennsylvanian and Permian subsidence continued to the south and west of the preserved basin and probably connected with the Orogrande and Pedregosa Basins. This larger "Permian Basin" contains both the Central Basin axis and the Diablo-Pedernal axis as intrabasin tectonic belts. The post-Permian erosion was probably due to a combination of uplift on the flanks of the Triassic-Jurassic rift complex, which resulted in the opening of the Gulf of Mexico, and uplift on the flanks of the Early Cretaceous Bisbee–Chihuahua Trough. Reconnaissance subsidence analysis of the West Texas Basin discloses a complex pattern of subsidence rates through the Permian. The most rapid tectonic subsidence took place in the Wolfcampian of the southern Delaware Basin, between the Marathon thrust sheets and the Fort Stockton uplift. Flexural subsidence is probably responsible. Post-Wolfcampian (postdeformational) subsidence of unknown origin continued to be centered in the north-south Delaware Basin trough, but extended north and east over a broad area of the Central Basin axis, the Midland Basin and the Northwest shelf to form the "Permian Basin."

INTRODUCTION

The West Texas Basin (informally but widely known as the "Permian Basin") is a broad region, about 500 by 600 km, underlain by thick Paleozoic strata in most of west Texas and southeastern New Mexico. It is one of the leading oil-producing regions in the United States. Its present boundaries, as commonly taken (Fig. I), are the Guadalupian (Permian) outcrop on the west (in the Guadalupe Mountains, Delaware Mountains, Apache Mountains), the Pedernal uplift to the northwest, the exposed and subcropping Marathon thrust belt and the Devils River uplift on the south (thereby enclosing the Val Verde Basin), the Llano uplift and the Bend arch on the east, and a broad margin on the north, including the shallow Palo Duro and Tucumcari Basins and the Matador arch/Roosevelt uplift, Amarillo/Wichita uplift and Sierra Grande uplift separating the West Texas Basin from the related Anadarko Basin of Oklahoma and the northeastern Texas Panhandle.

The presence of significant Permian outcrops in the area was determined in the first exploratory surveys. The true significance of the basin, however, was left for the drill to discover, most notably after the discovery of Big Lake Field in 1921. Most of the principal structural and stratigraphic features of the basin were delineated in the 1940s and 1950s, when wells were drilled to the Ordovician in many areas. Subsequently, much improved seismic data and continued exploration in the deep basins has helped to refine geologic knowledge of the basin.

However, the development of a comprehensive tectonic understanding of the West Texas Basin has been slow. Early works by Harrington (1963), Galley (1968) and Hills (1970) were the only published overviews of basin tectonics. In the 1970s, broad-brush papers in plate-tectonic terms were largely unconstrained by detailed field relationships (e.g., Walper, 1977). Brief treatments and speculations by Bolden (1984), Elam (1984) and Hills (1984) are also significant. Recently, several more detailed syntheses based on regional mapping have been published, notably by Gardiner (1990), Ewing (1991) and Shumaker (1992). The significance of Permian basinal subsidence has not been treated separately by most authors; Yang and Dorobek (1991) reported a model for flexural subsidence of the southern Delaware and Midland Basins. The West Texas Basin can best be viewed as a combination of two processes. One, tectonic structuring, resulted in the development of a complex pattern of uplifted and depressed blocks bounded by thrust to normal and strike-slip faults. This deformation is part of the "Ancestral Rocky Mountains" deformation, which continues north through New Mexico and Colorado and northeast into Oklahoma (Kluth and Coney, 1981). The basin was deformed in both the Early Pennsylvanian and in the Early Permian; structuring was essentially completed by late Wolfcampian time. The other process, basin subsidence, began in the Middle Pennsylvanian (Desmoinesian), accelerated in the Early Permian (Wolfcampian) and continued, long after deformation ended, to the end of the Permian. This subsidence created a broad bowl of sedimentation (Fig. 2) and led to the development and maintenance of the deep-water Delaware and Midland Basins adjacent to high, reef-rimmed Permian carbonate platforms, most notably the Central Basin platform (between the Delaware and Midland Basins), the Northwest shelf (north of the two basins), and later in late Guadalupian time, the Capitan reef rimming the Delaware Basin. The processes of deformation and subsidence interacted and overlapped, creating deep basins with buried, oil-productive structures (Hills, 1984).

This paper originated as an overview of subsidence history in the West Texas Basin, to accompany the Tectonic Map of Texas (Ewing, 1991). During the compilation of the map, I realized that, although the postdeformational subsidence was apparently similar in gross form to that of intracratonic basins, the patterns of subsidence fit only poorly the ideal model of concentric intracratonic subsidence. This led to a consideration of the margins of the preserved Permian strata of the West Texas Basin—which are significantly affected by pre-Albian (mid-Cretaceous) erosion. The primary purpose of this paper is to document the regional character and significance of these erosional boundaries, so that they may be taken into account in constructing preliminary maps of the complex subsidence history of the basin.

ERODED MARGINS OF THE WEST TEXAS BASIN

The West Texas Basin differs in a major way from other large intracratonic basins of North America: its present boundaries do not reflect the original limits of Pennsylvanian-Permian basin subsidence. Hence, in interpreting the location of the center of subsidence or its causes, we may be substantially misled by the preserved thicknesses shown in Fig. 2. Three margins of the West Texas Basin have been altered by later uplift and erosion; only the northern boundary appears to be truly depositional.

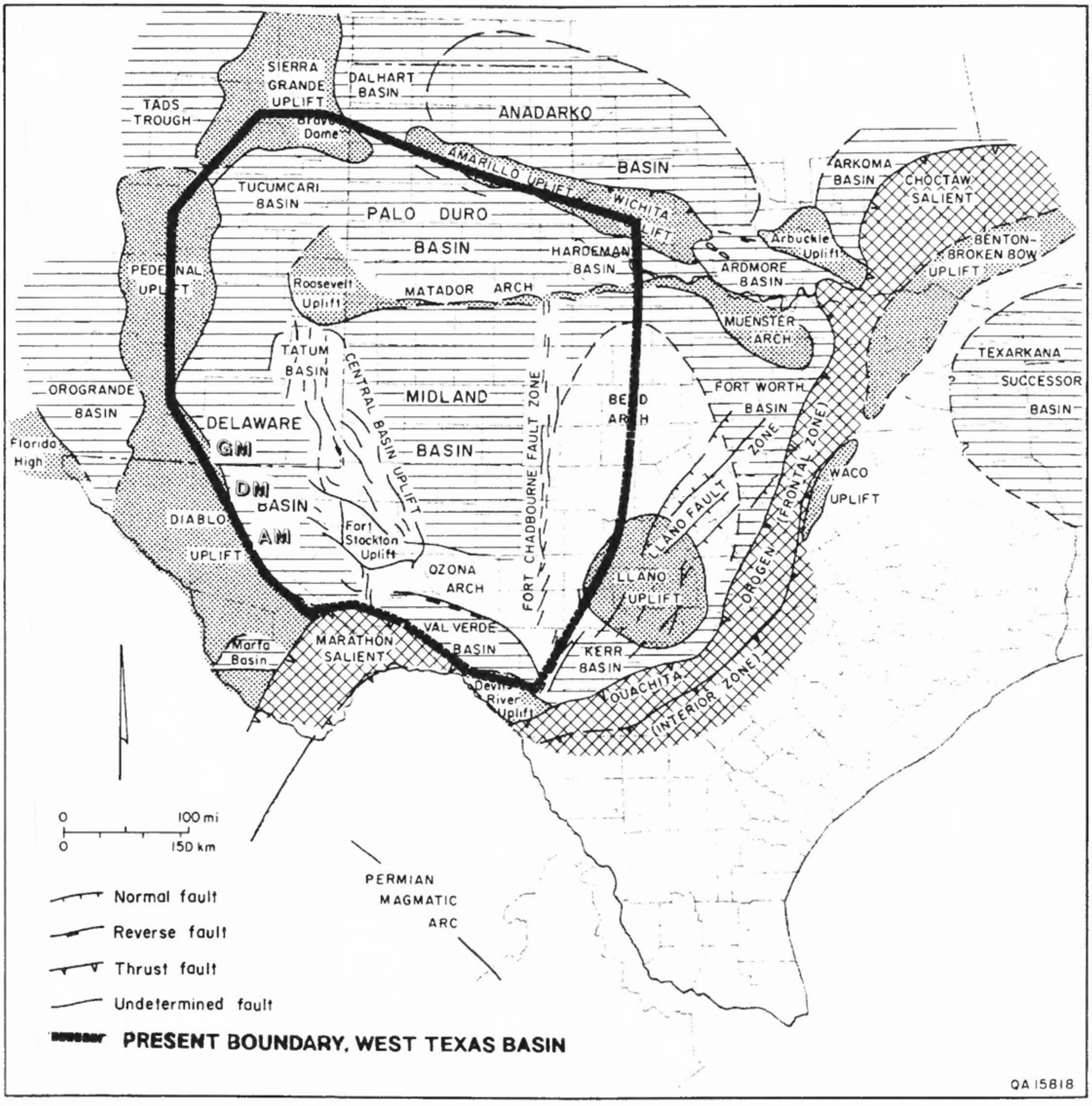

FIGURE 1. Present boundaries of the West Texas Basin and major late Paleozoic structural elements, modified from Ewing (1991). GM = Guadalupe Mountains; DM = Delaware Mountains; AM = Apache Mountains.

Northern boundary

The northern boundary of the West Texas Basin, extending from the north end of the Sacramento Mountains north and east to the Permian outcrop of Oklahoma, is the only one that can be confidently considered depositional (that is, with original thicknesses and facies essentially preserved). In this sector, sedimentary facies become steadily thinner (except for more rapid, tectonically influenced Pennsylvanian and Wolfcampian subsidence in northern New Mexico), more clastic and more continental away from the basin center (Oriel et al., 1977; Dutton et al., 1982). For example, the thin Bernal Formation of northern New Mexico is the equivalent of the thick, complex Artesia Group of the West Texas Basin.

The major anomaly of the northern boundary is the Permian depocenter in the Anadarko Basin. This additional depocenter is asymmetric, deepening against the upthrust Wichita uplift to the south. The Pennsylvanian and Wolfcampian subsidence of this basin is probably related to flexural loading by the Wichita uplift (Brewer et al., 1983). Post-Wolfcampian subsidence is also greater over the earlier foredeep (Fig. 2). The northern margins of the two basins merge in thickness and facies and later Permian deposition over the Amarillo-Wichita axis fits regional Permian subsidence trends.

FIGURE 2. Isopach of post-Wolfcampian Permian strata, showing the preserved outline of the West Texas Basin and the division into subbasins (from Oriel et al., 1977, as used in Ewing, 1991).

Southern boundary

The southern boundary of the West Texas Basin is traditionally taken to be the northern edge of the exposed Marathon orogenic belt, a thin-skinned thrust-and-fold belt of lower Paleozoic basinal strata and upper Paleozoic flysch and its buried extension to the southeast (Fig. 3). Upper Wolfcampian strata overlap the deformed Marathon rocks at the Glass Mountains on the northwest side of the exposed Marathon window (Ross, 1986). However, the thick sediments of the Delaware and Val Verde Basins continue beneath the Marathon thrust sheets for a substantial distance.

The southern margin is in reality two margins (Fig. 4). The syndeformational basins of the Val Verde Basin and the southern Delaware Basin occur in front of and beneath the Marathon thrust sheets and in front of the uplifted, basement-cored Devils River uplift, which may pass beneath much or all of the Marathon belt (Nicholas, 1983; Ewing, 1985). The Val Verde Basin is southward-thickening and similar to

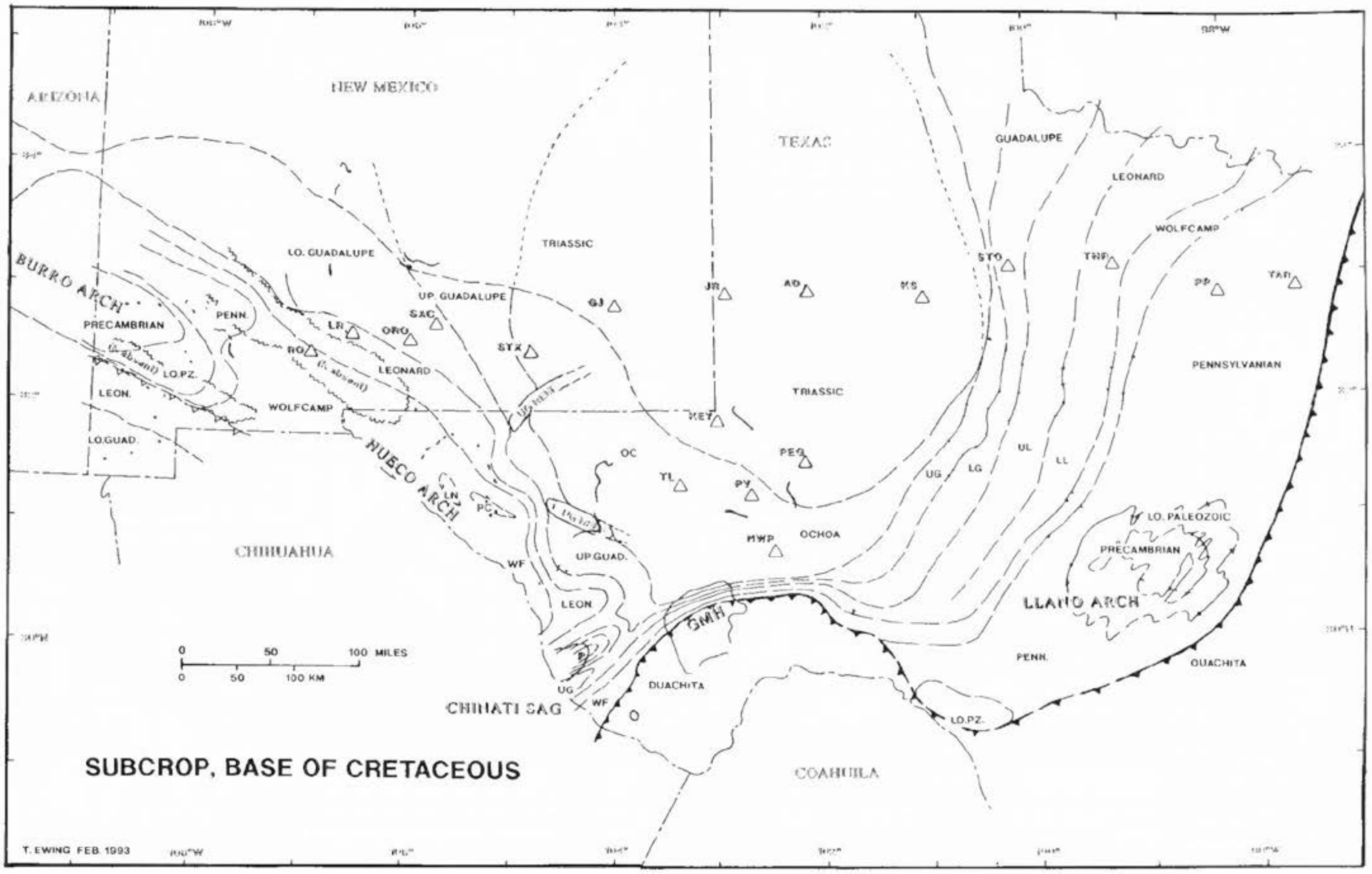

FIGURE 3. Sub-Cretaceous (mostly sub-Albian) subcrop map of the West Texas Basin, showing Mesozoic tectonic features: the northwest-trending Hueco and Burro arches, the east-trending Glass Mountains homocline (GMH) and the broad area of erosion west and north of the Llano Uplift (the Llano arch). Sources for Texas: the Geologic Atlas of Texas (summarized on the Geologic Map of Texas, Barnes, 1993) and Henry and Price (1985). Sources for New Mexico: Dane and Bachman (1961), Seager et al. (1982), Seager and Mack (1986), Ross and Ross (1986) and Seager et al. (1987). Triangles locate subsidence curves analyzed (for location abbreviations, see Table 1); barbed line is the front of the Ouachita-Marathon thrust belt.

other foredeep basins (Wuellner et al., 1986). The deep-water basin fill traditionally has been considered Wolfcampian, but much or all of it may be Late Pennsylvanian in age (Nicholas, 1983). The southern part of the Delaware Basin (which lies in front of the exposed Marathon thrusts) contains thick Wolfcampian (and Upper Pennsylvanian?) shales, but much of its subsidence may be due instead to the impinging Fort Stockton Uplift (Central Basin axis, Fig. I) to the northeast (Yang and Dorobek, 199 I). The north-northwest orientation of the Wolfcampian basin axis supports this idea.

The postdeformational (post-Wolfcampian) margin is, however, erosional. Strata from late Wolfcampian to Ochoan age are tilted 2°-3° to the north and beveled beneath Cretaceous strata of the Edwards Group (Albian). This relationship is exposed at both ends of the Glass Mountains (King, 1980; Ross, 1986), suggesting the name "Glass Mountains homocline" for this feature (Fig. 3). The homocline continues in the subsurface for 150 km eastward and is well imaged in seismic data in the Val Verde Basin (Fig. 4). This beveling appears to be entirely post-Permian, as no significant regional thinning or facies changes are evident; the post-Wolfcampian strata of the Glass Mountains are thick carbonates deposited near the platform margin and not thin, clastic-dominated, basin-edge deposits (Ross, 1986).

Western boundary

The southwestern boundary of the preserved West Texas Basin is typically defined by the exposures of the Guadalupe, Delaware and Apache Mountain ranges of Trans-Pecos Texas. These are ranges of mainly east-dipping Upper Permian strata, exposing the Guadalupian through Ochoan fill of the Delaware Basin in the Delaware Mountains and the imposing Guadalupian reef escarpments in the Guadalupe Mountains to the north and the Apache Mountains to the south. These ranges occur along the east margin of the Salt Basin, a Late Tertiary and Quaternary bolson. It is likely that they were tilted and eroded in late Tertiary time. However, an earlier (and more regionally significant) period of erosion also occurred.

Compilation of the available information on the sub-Cretaceous sub-crop in Trans-Pecos Texas and southern New Mexico discloses the size and the regional character of pre-Cretaceous uplift in Trans-Pecos Texas (Fig. 3). In the Sierra Diablo southwest of the Guadalupe Mountains, Lower Cretaceous rests on Leonardian rocks (Victorio Peak Formation); to the west and southwest, Lower Cretaceous nearly everywhere rests on Wolfcampian shelf carbonate strata (Hueco Limestone; King and Flawn, 1953; King, 1965). Locally, the Cretaceous rests on Precambrian, as at the Pump Station Hills and at places in the Van Horn area, where pre-Cretaceous north-down faulting is likely. This broad area of Wolfcampian subcrop continues northwest into the Cooks Peak area of southwestern New Mexico (northwest of RO in Fig. 3). Northwest of that location, Lower Cretaceous rests on Pennsylvanian through Precambrian rocks, part of a proto-Burro uplift, or "Burro arch," which extends into Arizona (Ross and Ross, 1986). To the southwest of the Burro arch, strata as young as early Guadalupian (part of the Pedregosa Basin succession) are preserved beneath the Lower Cretaceous rocks of the Bisbee trough. These subcrops have been displaced northeastward an unknown amount by Laramide thrusting.

Do these regional relationships mean that the post-Wolfcampian Per-

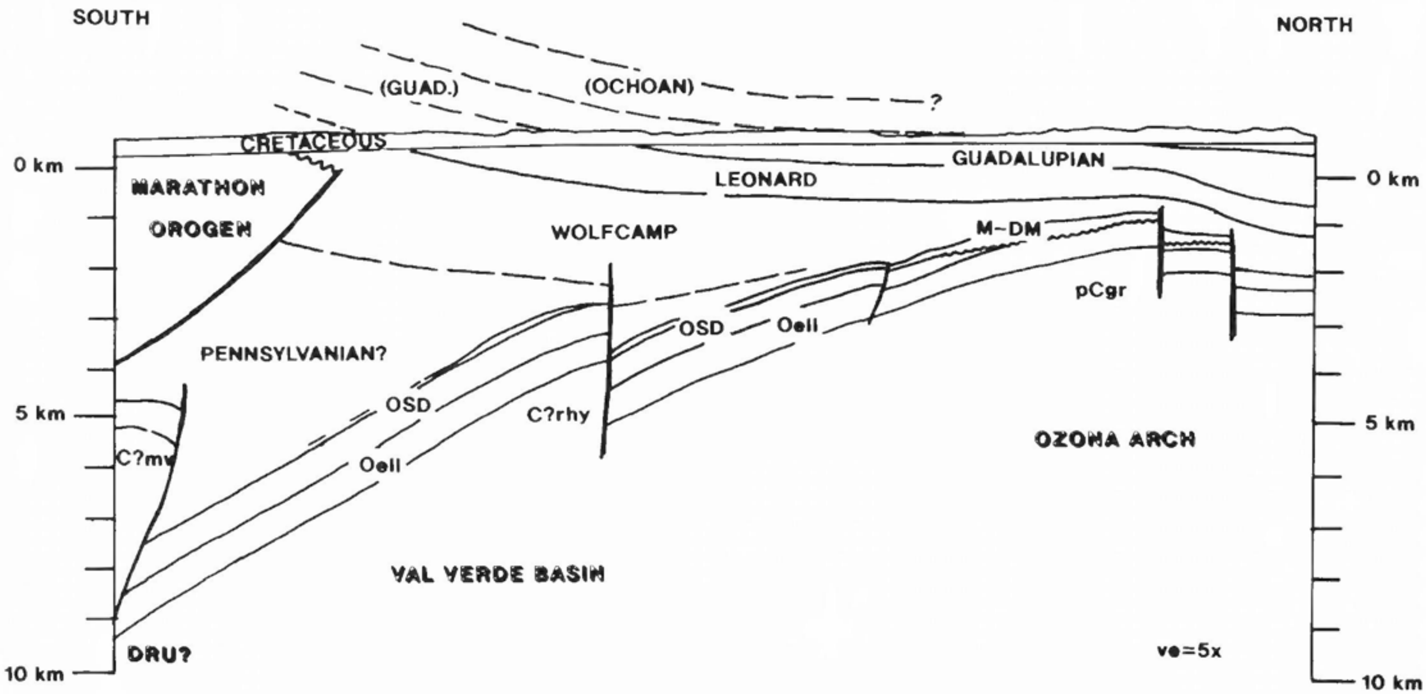

FIGURE 4. Cross section across the Val Verde Basin and Glass Mountains homocline, using information from seismic data and from Nicholas, 1983. DRU = Devils River Uplift (possible western extension); Oell = Ellenburger Group (Ordovician); M-DM = Mississippian and Desmoinesian limestones; C?mv = Cambrian? metavolcanics (Nicholas, 1983); C?rhy = Cambrian? rhyolite.

mian strata were never deposited across the broad area of Wolfcampian subcrop? No basin-edge mixed clastic-carbonate zones are reported in the westernmost exposures of the Permian rocks. It seems more reasonable to maintain that the younger Permian rocks continued an unknown distance westward, but were stripped off across a broad "Hueco arch" before late Early Cretaceous deposition. Additional uplift and stripping of Permian strata also occurred in New Mexico over Laramide uplifts, where the Cretaceous cover was removed (Seager and Mack, 1986). Still younger, later Tertiary tilting and erosion of the Guadalupe, Delaware and Apache Mountains fault blocks stripped their Cretaceous cover which, from the regional subcrop, must have lain just above the present high peaks of the Guadalupe Mountains.

Between the uplifted Hueco arch and the Marathon-Glass Mountains is a complex area in the Marfa Basin and Chinati Mountains area of Trans-Pecos Texas where younger Permian rocks are preserved. Around the rim of the Chinati caldera, rocks ranging from Pennsylvanian (east margin) to lower Guadalupian (north margin) to upper Guadalupian (south margin) subcrop beneath the Lower Cretaceous (Ross, 1986). The inferred differential uplift is difficult to fit into a regional framework, but the entire area does form a preserved sag of later Permian strata between the Glass Mountains homocline and the Hueco arch. This is termed the "Chinati Sag" (Fig. 3), to differentiate it from the Paleozoic, structural "Marfa Basin." The sag is also the location of the inferred "Hovey channel," a Permian paleogeographic feature (Ross, 1986); it seems possible that the location of the channel could be an artifact of pre-Cretaceous preservation.

The western to northwestern boundary of the basin is formed by the extensive San Andres Formation and Artesia Group outcrop of the Pecos slope and Sacramento Mountains. As in the Guadalupe and Delaware Mountains, this area is today an east-dipping homocline formed by late Tertiary uplift of the Sacramento Mountains east of the Tularosa Basin. However, in the Sierra Blanca area, upper Guadalupian rocks (Artesia Group) are beveled by Triassic and Cretaceous strata and only lower Guadalupian rocks (San Andres Formation) subcrop to the northwest. Minor erosion before Late Triassic deposition (Chinle or Dockum Group) is widespread across the basin center (Fig. 3), but its relationship to the much larger pre-Cretaceous beveling to the south is problematic.

Eastern boundary

The eastern boundary of the West Texas Basin has elements of both erosional and depositional margins. It is marked by the Bend arch and Llano uplift, both high areas related to Pennsylvanian foredeep subsidence to the east and southeast and the Permian subsidence to the west. In north Texas, particularly, there is a continuous record of episodic infill of the basin by sediments derived from the Ouachita Mountains to the east and the Wichita uplift to the northeast. Permian sediments are generally clastic-evaporite deposits similar to the basin-margin facies of the northern boundary.

However, this boundary is also erosional. Substantial thicknesses of gently west-dipping Permian rocks are exposed in outcrop and are truncated beneath flat-lying Lower Cretaceous (Albian) rocks (Antlers Sandstone and Edwards Group; Barnes, 1993). This geometry suggests increasing pre-Cretaceous uplift and erosion to the east (Fig. 3); the name "Llano arch" is suggested. This very gentle homocline probably turns westward, steepens and joins with the Glass Mountains homocline, but the details are not known.

The "Llano uplift" of central Texas is an area of exposed Precambrian and lower Paleozoic rocks that lies within the broader Llano arch; it is not a structural uplift (a fault-bounded, high-standing basement block such as the Fort Stockton uplift). The exposure is due to the intersection of four epeirogenic arches. It lies at the end of the northwest-trending middle Paleozoic "Texas arch"; and at the junction of the north-trending Pennsylvanian "Bend arch" and the east-trending "Ozona arch" (both of which were probably flexural bulges related to the Fort Worth and Val Verde foredeep basins; Ewing, 1991). These previously generated gentle arches were scalped by Llano arch erosion to reveal Precambrian strata before Cretaceous deposition.

PATTERNS OF LATE PALEOZOIC SUBSIDENCE

To summarize the above discussion, the Permian subsidence that marked the West Texas Basin probably continued somewhat to the east and an indefinite distance to the west and south before erosion sometime during the Mesozoic. Thus, the apparently concentric nature of post-Wolfcampian subsidence (Fig. 2) is misleading and simple examination of such isopachs is probably insufficient in attempts to better understand the nature of the Permian subsidence. The only proper way to examine this history is through the use of subsidence curves, which show the effect of subsidence over time, explicitly show unconformities, and allow for water depth variations.

Methods used; the time scale question

Thickness and water depth estimates used for subsidence curves are general estimates based on publicly available information (Table 1). The cross sections of the West Texas Geological Society (1962, 1964, 1984) have been particularly useful for this purpose. In many areas, sedimentation took place for long periods far below sea level; therefore, some correction must be made for water depth. This has been done by estimating the height of the rimming reef margins near the analysis; between control, a reasonable "best-guess" water depth evolution was taken.

The absolute time scales for the late Paleozoic are needed for subsidence analysis. Unfortunately, there is substantial disagreement on the correct values. Three sets of published estimates are shown at the head of Table 2 (I have subtracted the absolute value given for the end of the Permian, to give more "intuitive" age numbers): COSUNA time chart (Salvador, 1985), Harland et al. (1982) and Harland et al. (1989). The last two are based on worldwide correlations of the Carboniferous and Permian. The Harland et al. (1989) scale approximates the Salvador (1985) time scale in the Pennsylvanian and the base of Permian is fairly uniform between scales.

None of the published scales gives good results when applied to West Texas Basin subsidence curves. This is shown on Fig. 5, where the values for Kelly Snyder (Locality KS, Fig. 3) are plotted with the time scales. The Kelly Snyder field is located on the Horseshoe Atoll, a long-lived carbonate reef complex distant from any significant deformation-related anomalies. The major problems are the short duration of the upper and lower Guadalupian in both the Harland et al. (1982) and the Harland et al. (1989) scale and the short duration of the Leonardian and Desmoinesian in the Salvador (1985) time scale.

For the purposes of this work, I adopted a modified time scale based on the three published efforts. This time scale follows Harland et al. (1989) in the Pennsylvanian, uses the 285 Ma Pennsylvanian-Permian boundary and a 245 Ma Permo-Triassic boundary, and expands the Guadalupian substages by 1 Ma. This scale gives the smoothest curves for total subsidence at Kelly Snyder and gives the most reasonable

TABLE 1. Basic data for subsidence estimates at various sites in the West Texas Basin area, shown on Fig. 5. All depths are in feet.

LOCATION	COUNTY	PRESENT DEPTH TO BASE (feet) Ochoa	Up.Guad.	Lo.Guad.	Leonard	Wolfcamp	Virgil	Missouri	Desmoines	Atoka	Morrow	Precambrian	SOURCE
Grayburg-Jackson	Eddy, NM	1100	3190	4660	8200	9800		10700	10950	11900	11900	14300	WTGS, 1964
Water Depth, period start	T17S R30E	0	0	0	0	0	0	0	0	0	0		
Jones Ranch	Gaines, TX		0	1580	4230	4950		5360	5560	5760	5760	8130	WTGS, 1984
Water depth, period start			0	0	0	0		0	0	0	0		
Adair	Terry, TX	935	2515	3805	6485	7310	8000	8595	9310	9885	9885	10775	WTGS, 1984
Water depth, period start		0	0	0	350	0	0	0	0	0	0		
Pegasus	Upton, TX	1240	3550	4590	7580	8505	8505	8505	8700	9050	9050	11990	WTGS, 1962
Water depth, period start		0	0	100	650	650			0	0	0		
Kelly-Snyder	Scurry, TX	160	1185	2155	3270	6200	6280	6580	7310	7360	7360	8000	WTGS, 1984
Water depth, period start		0	0	0	0	300	0	0	0	0	0		
Keystone	Winkler, TX	0	1360	2710	4240	5210					5210	8420	WTGS, 1964, 1984
Water depth, period start		0	0	0	0	0					0		
Toyah Lake	Reeves, TX	4700	5720	7880	10300	14700					15100	20500	WTGS, 1964
Water depth, period start		1400	1400	1400	1300	1100					0		
Pecos Valley	Pecos, TX	1010	2040	3090	4680	5440					5440	9120	WTGS, 1962, 1964
Water depth, period start		0	0	0	0	0					0		
NW Puckett (Gulf 1-Winfield)	Pecos, TX	830	2900	4600	6700	16100		17450			18100	26000	WTGS, 1964
Water depth, period start		0	1600	1900	1700	1000		0			0		
Stonewall Co. composite				0	2500	3040	4860	5250	6300	8900	6900	7390	BEG Data
				0	0	0	700	100	0	0	0		
Throckmorton Co. composite					0	570	1600	2800	4500	4800	4800	5490	BEG Data
					0	0	0	0	0	0	0		BEG Data
Palo Pinto Co. composite						0	150	1530	3650	4550	4550	7500	BEG Data
						0	0	0	0	0	0		
Tarrant Co. composite								0	1400	5000	8000	11940	BEG Data
								0	0	0	0		
E. Side Pedernal Uplift	Eddy, NM			3500	4450	5100	6180	6850	7050	7480		9580	Meyer, 1966
(SoTex #1 State E-6584)	T21S R22E			0	0	0	0	0	0	0			
Sacramento Mtns.	Otero, NM			2500	3040	3350	3860	4780	4950	4970		6220	Kottlowski, 1963
(Nigger Ed Canyon sec'n)	T19S R11E			0	0	0	0	0	0	0			(in Meyer, 1966)
Orogrande Basin	Otero, NM			0	1850	5650	6000	6575	7170	7500	7630	8880	Meyer, 1966
	T20S R9E			0	0	500	250	0	0	0	0		
Love Ranch	Dona Ana, NM			0	650	2310	4645	4830	5020	5120	5120	6370	Kottlowski et al. 1956
	T19S R4E			0	0	0	300	0	0	0	0		(in Meyer, 1968)
Robledo Mtns.	Dona Ana, NM				0	1850	2090	2300	2510	2540	2540	3800	Kottlowski, 1963
	T21S R1W				0	0	0	0	0	0	0		(in Meyer, 1966)

TABLE 2. Results of subsidence modeling. Time scales shown are in Ma, counting back from the end of the Permian (245 Ma according to Harland et al., 1989). Tectonic subsidence rates are in meters per million years (m/Ma), calculated from data of Table 1; decompaction and correction for isostatic effects of sediment load have been applied.

	Ochoa	Up.Guad.	Lo.Guad.	Leonard	Wolfcamp	Virgil	Missouri	Desmoines	Atoka	Morrow
--AGE ESTIMATES (Ma; End Permian=0)--										
Age at base--Salvador, 1985	5	12	20	23	40	52	60	62	64.5	80
Age at base--Harland et al., 1982	8	9	10	17	40	44	47	52	57	72
Age at base--Harland et al., 1989	5	7.5	10	24	45	51	58	62	66	78
Age at base -- optimized	5	11	18	24	40	51	58	62	66	78
Absolute Age -- optimized	250	256	263	269	285	296	303	307	311	323
RATES (m/Ma)										
Grayburg-Jackson (GJ)	10.1	12.3	15?	22?	16?	7.1	7.1	9.7	41.0	
Jones Ranch (JR)			14.9	22.3	6.1	3.5	3.5	8.0	8.2	
Adair (AD)	13.0	18.3	12.1	10.3	11.3	9.8	15.1	36.3	34.0	
Pegasus (PEG)	16.9	25.9	6.0?	6.0?	12.2	6.8	6.8	6.8	14.0	
Kelly-Snyder (KS)	1.9	9.5	6.5	6.0	26.8	1.4	8.6	42.6	3.2	
Keystone (KEY)		10.3	5.3	28.6	8.3					
Toyah Lake (TL)	10.0	13.8	27.1	37.0	40.0	7.6	7.6	7.6	7.6	
Pecos Valley (PV)	6.5	2.9	13.8	29.4	6.2					
NW of Puckett (NWP)	0.0	19.6	6.8	24.0	64.0	22.0	22.0	3.5	3.5	3.5
Stonewall Co. (STO)				11.5	2.7	17.9	25.0	57.0	37.0	
Throckmorton Co. (THR)					3.1	9.5	23.6	64.0	14.0	
Palo Pinto Co. (PP)						1.1	19.0	59.0	36.0	
Tarrant Co. (TAR)								21.0	89.0	36.0
SoTex #1 State E-6584 (STX)			?27	4.5	5.5	13.3	14.9	8.3	19.1	
Sacramento Mtns. (SAC)			?20	4.6	2.5	4.1	28.0	8.6	1.0	
Orogrande Basin (ORO)				7.1	20.8	8.9	19.1	28.0	17.2	2.4
Love Ranch (LR)				2.6	10.3	26.0	12.8	9.3	5.0	
Robledo Mtns. (ROB)					14.4	3.5	5.0	9.3	1.4	

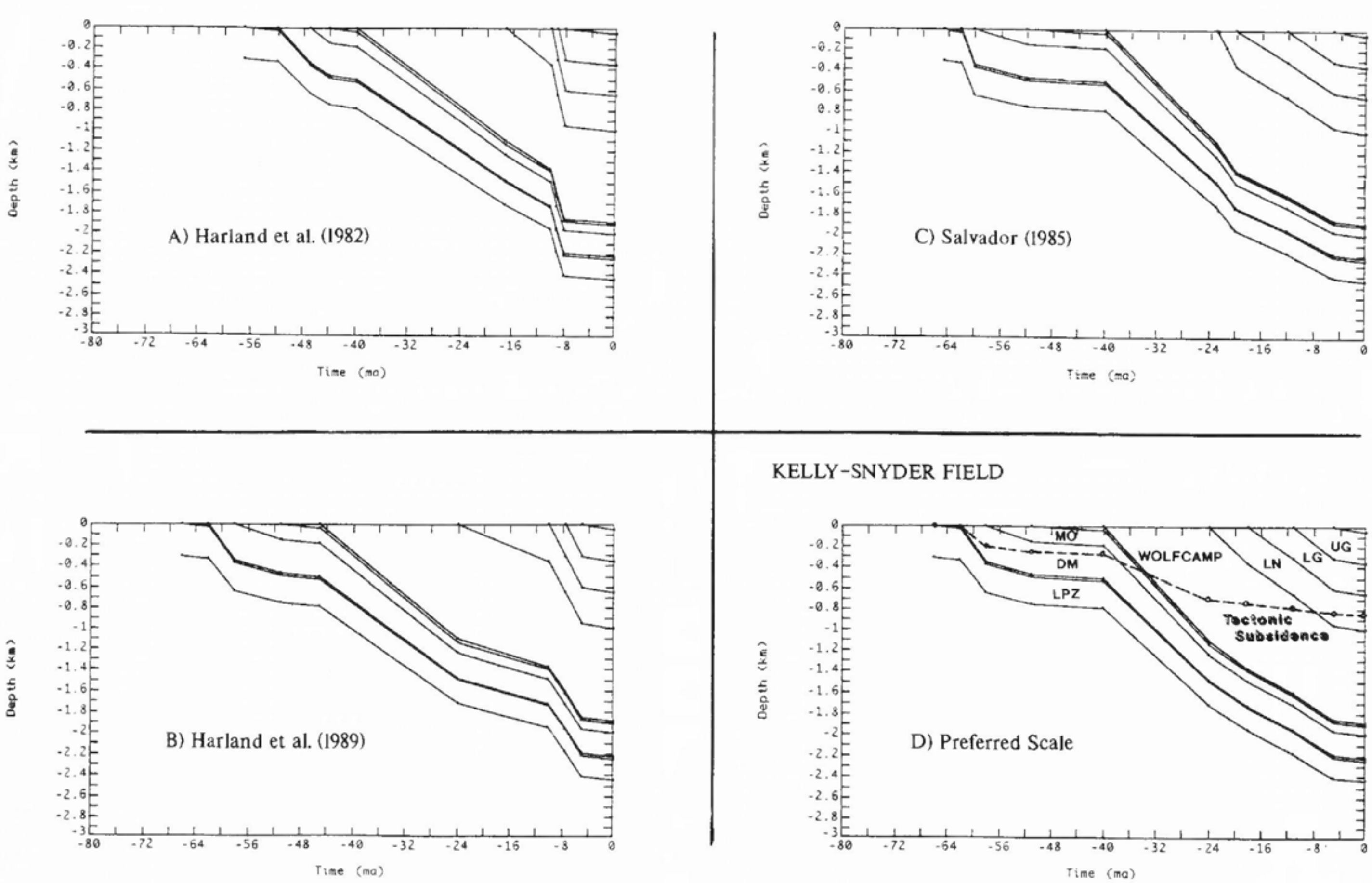

FIGURE 5. Subsidence curve from Kelly Snyder field, Scurry Co., Texas, using A) time scale of Harland et al. (1982); B) time scale of Harland et al. (1989); C) time scale of Salvador (1985); D) time scale as optimized and used in the present study.

values elsewhere. The largest remaining anomaly is unusually rapid subsidence in the Desmoinesian, possibly indicating that too short a time interval has been allotted.

There is certainly no universal validity to this time scale, as it is based on reconnaissance work and is subject to regional stage miscorrelations and unresolved tectonic activity. However, this sort of approach in various basins worldwide could help tie down some of the more poorly calibrated parts of the time scale. In the present context, any inaccuracies in these ages affects the comparison of subsidence rates between stages. Map patterns within any one stage are unaffected by these errors, although regional miscorrelations can still have a substantial effect.

A commercially available subsidence program (Subside! from Rockware) was used for converting thickness and time estimates into subsidence rates. Compaction corrections from Sclater and Christie (1980) and Schmoker and Halley (1982) were applied, using general lithology estimated in the various areas. A tectonic subsidence component was thereupon calculated, allowing for the isostatic effects of sediment deposition. The commercial program does not explicitly allow for water depth variations; the effect of these was estimated by adding changes in water depth and correcting for the reduced isostatic effect of water vs. sediment.

Two corrections made to three curves are due to the reconnaissance nature of the study, relying on regional and published correlations. Two localities, STX and SAC, were not correlated above the Leonardian in the original source; hence, the total thickness overlying the Leonardian, assigned to the lower Guadalupian, yields too much subsidence. These are adjusted downward somewhat arbitrarily; the numbers are queried on Table 2. Locality GJ yielded erratic, uncontourable data, probably due to the difficulty of correlating the Leonardian stage boundaries in the Abo Reef area. The tectonic subsidence curve at this locality was smoothed between the base Wolfcampian and the mid-Guadalupian; numbers are queried on Table 2.

Results

The subsidence history of the West Teas Basin, examined away from the effects of local uplifts or flexural subsidence related to those uplifts, is generally similar to that in other intracratonic basins such as the Michigan or Illinois Basins (Quinlan, 1987). Major tectonic subsidence began in the Early Pennsylvanian, accelerated to a maximum during Wolfcampian time and continued throughout the Permian, as shown by the subsidence history curve of Kelly Snyder field (KS; Fig. 5D). When the subsidence curve is corrected for the effects of sediment loading, the "tectonic subsidence" curve shows nearly constant, nonzero subsidence rates after the Wolfcampian.

In order to understand the history of subsidence during and after the main period of deformation in the basin, tectonic subsidence rates were mapped for the Wolfcampian, Leonardian, lower Guadalupian and upper Guadalupian (Fig. 6). Ochoan rates were not mapped, as the top is usually eroded.

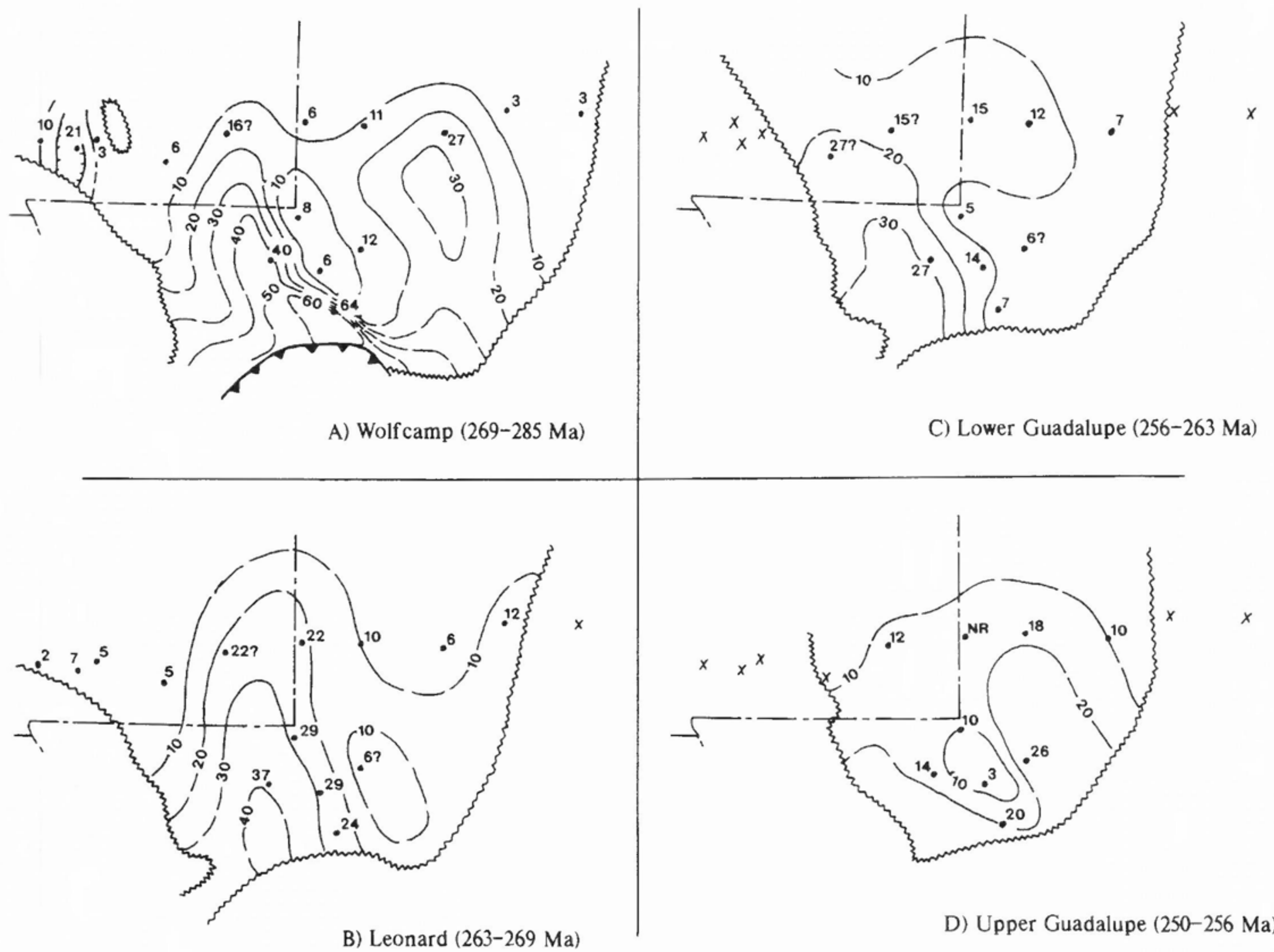

FIGURE 6. Spatial variation in tectonic subsidence rates for A) Wolfcampian, B) Leonardian, C) lower Guadalupian and D) upper Guadalupian strata across the West Texas Basin. See Fig. 5 and Table 2 for identity of localities. All values are in meters/million years (m/Ma), using the adjusted time scale in Table 2.

Wolfcampian subsidence rates show the effect of uplift of the Central Basin axis (KEY, PV on Fig. 3) and the Diablo-Pedernal axis. Between the axes, the deep Delaware Basin subsided rapidly, with greater rates to the south (NWP) but substantial rates at least to the state line (TV, GJ). Slower but still significant subsidence characterized the Midland Basin to the east (PEG, KS). The subsidence of the Delaware and Midland Basins during this time was modeled by Yang and Dorobek (1992) as a flexural response to the load of the southwest-overthrust Fort Stockton uplift. In south-central New Mexico, the Orogrande Basin west of the Pedernal uplift subsided significantly.

During Leonardian time, the most rapid subsidence occurred in the Delaware Basin. Sites on the Central Basin axis and the Northwest shelf (GJ, JR, KEY, PV) shared in this subsidence, while the Midland Basin subsided more slowly.

During early Guadalupian time, the most rapid subsidence was still in the Delaware Basin (TV); parts of the Central Basin axis showed low subsidence (KEY) or intermediate subsidence rates. During late Guadalupian time, the most rapid observed subsidence was in the Midland Basin (PEG; anomalous?) and lower subsidence was observed over the buried Central Basin axis, but the main effect may be an overall southward increase in subsidence rates.

More points need to be added to this mapping to give a satisfactory picture of the Permian subsidence. However, these preliminary maps are enough to discuss mechanisms of subsidence. The most significant and long-lived center of subsidence was the Delaware Basin, especially the north-northwest trending trough extending from NWP towards GJ. Although the Wolfcampian can be explained by the subsidence and filling of a foredeep basin, the post-Wolfcampian results need further explanation. It may be possible that there is a time-delayed flexural-isostatic response or a time-delayed compaction response (as tectonic subsidence calculations assume effectively instantaneous isostatic and compaction responses). The broader Leonardian anomaly could result from such a deep adjustment. Alternatively, it is possible that there is
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a deeper coupling of Central Basin axis uplift and Delaware Basin subsidence that continued until late in the Permian, superposed on a general subsidence of the region. The low values of Guadalupian tectonic subsidence at KEY and PV would be consistent with this model. Enhanced postdeformational subsidence over an earlier foredeep basin is also evident in the Anadarko Basin (Fig. 2).

A useful synoptic picture of the relationships of the late Paleozoic basins of Texas and New Mexico can be obtained by extending this preliminary subsidence analysis eastward and westward. An east-west cross section from Fort Worth, Texas (TAR) to the Robledo Mountains near Las Cruces, New Mexico (ROB; Fig. 7) shows the relationships of the Fort Worth, West Texas ("Permian") and Orogrande Basins, as reduced to relative tectonic subsidence rates. This northern line of section was chosen to minimize the effects of Pennsylvanian and Wolfcampian structuring of the Central Basin axis. Despite the remaining uncertainty on the time axis and hence on time variation of the calculated subsidence, several points are immediately evident:

1. The Fort Worth Basin, as a foredeep in front of the Ouachita thrust belt, shows high subsidence rates (locally over 100 m/Ma) from the earliest Pennsylvanian through the Missourian. The easternmost preserved Virgilian values suggest a slowing of subsidence, possibly marking the cessation of Ouachita thrusting in the area.

2. In the northern part of the West Texas Basin, significant (over 50 m/Ma) early (Desmoinesian) subsidence was followed by local Wolfcampian subsidence and broader post-Wolfcampian "Permian Basin" subsidence, both over 20 m/Ma. Regional "background" subsidence rates of 5-10 m/Ma prevail in the rest of the basin.

3. The Orogrande Basin shows two periods of subsidence of over 20 m/Ma, one Desmoinesian-Missourian and one Virgilian-Wolfcampian, with varying centers of subsidence.

Again, these results need to be refined using more control, especially coupled with closer evaluation of the paleontologic dating and correlation of the sections.

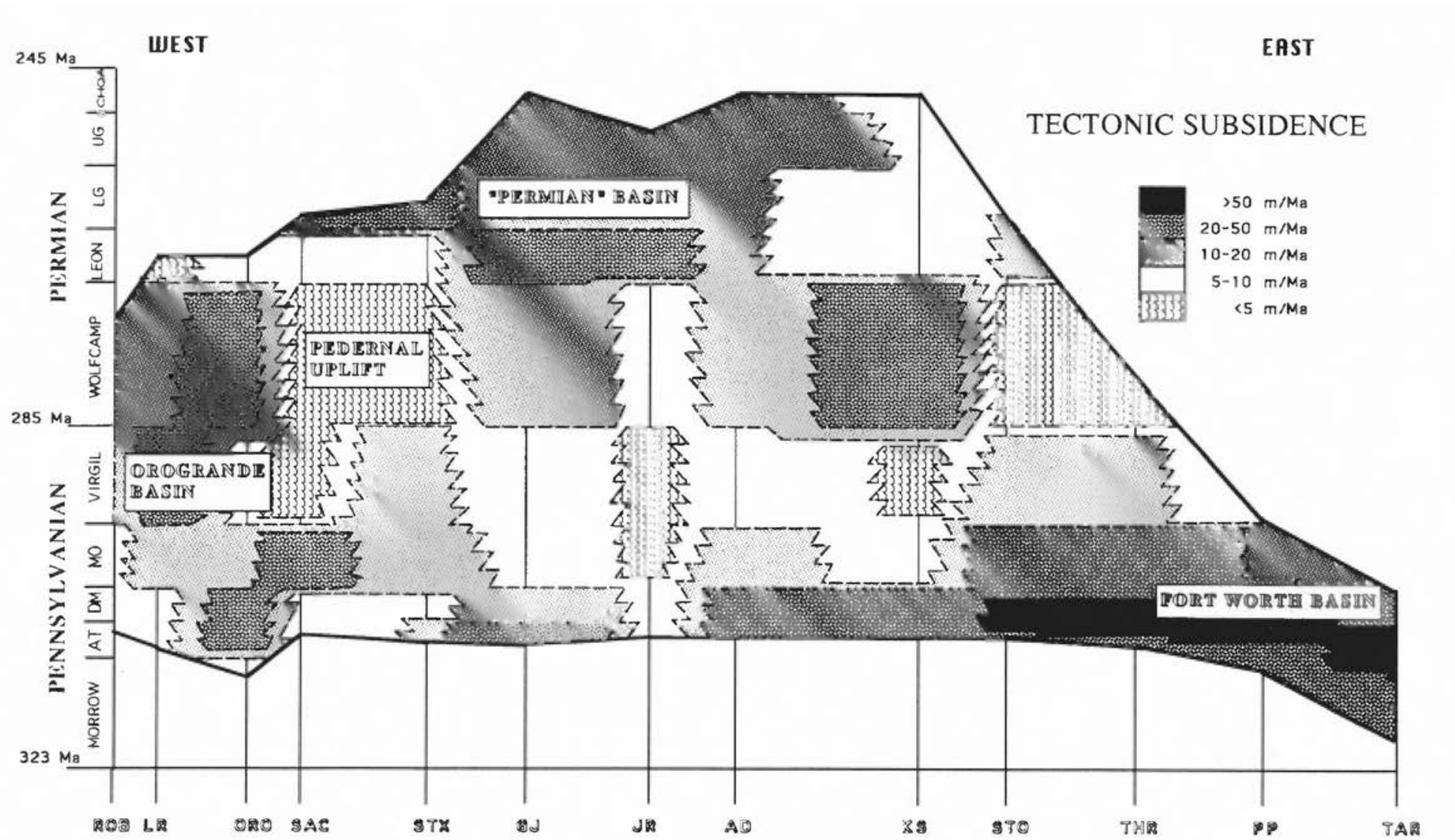

FIGURE 7. West-east time section from the Robledo Mountains, New Mexico to Fort Worth, Texas; locations shown in Fig. 5. Values are tectonic subsidence rate (m/Ma) as calculated from site estimates in Table 2.

IMPLICATIONS FOR TECTONICS

Episodes of Mesozoic erosion

So far, we have examined evidence that the eastern, southern and western boundaries of the Permian Basin are erosional and that basinal subsidence was complex and centered to the west of the Central Basin axis, in the Delaware Basin near the present outcrop. What were the natures of the erosional events that so skewed the present outcrop or subcrop pattern?

Except for the late Tertiary uplift of the Sacramento, Guadalupe and Delaware Mountains, most of the erosional trimming of the Permian Basin was pre-Edwards in age (i.e., pre-Albian). There are two possibilities for uplift and erosion between the Permian and the Edwards.

1. Late Triassic (+Jurassic?) uplift and erosion could have taken place on rift shoulders surrounding the initial (Eagle Mills) rifting that resulted in the Gulf of Mexico Basin (Fig. 8a). Erosion of such a rift shoulder could have led to the deposition of the thick, coarse elastics of the Late Jurassic Cotton Valley Group and earliest Cretaceous Travis Peak Formation in the Gulf Coast Basin. Pre-Late Jurassic grabens also developed in northeastern Mexico (Sabinas Basin), but not to the northwest.

2. Earliest Cretaceous uplift and erosion could have accompanied the formation of the northwest-trending Sabinas and Chihuahua troughs and the Bisbee Basin in southeast Arizona (Fig. 8b). Pre-Albian erosion has been identified in a wide area of central Arizona and southwestern New Mexico (Fig. 3; Dickinson, 1981; Ross and Ross, 1986) and related to a rift shoulder synchronous with the beginning of Bisbee Group deposition in fault bounded troughs (Bilodeau and Lindberg, 1983). Uplift also produced Lower Cretaceous and Upper Jurassic coarse elastics in the Sabinas Basin, but not to the southeast (Alfonso Zwanziger, 1978; Jones et al., 1984; McKee et al., 1984).

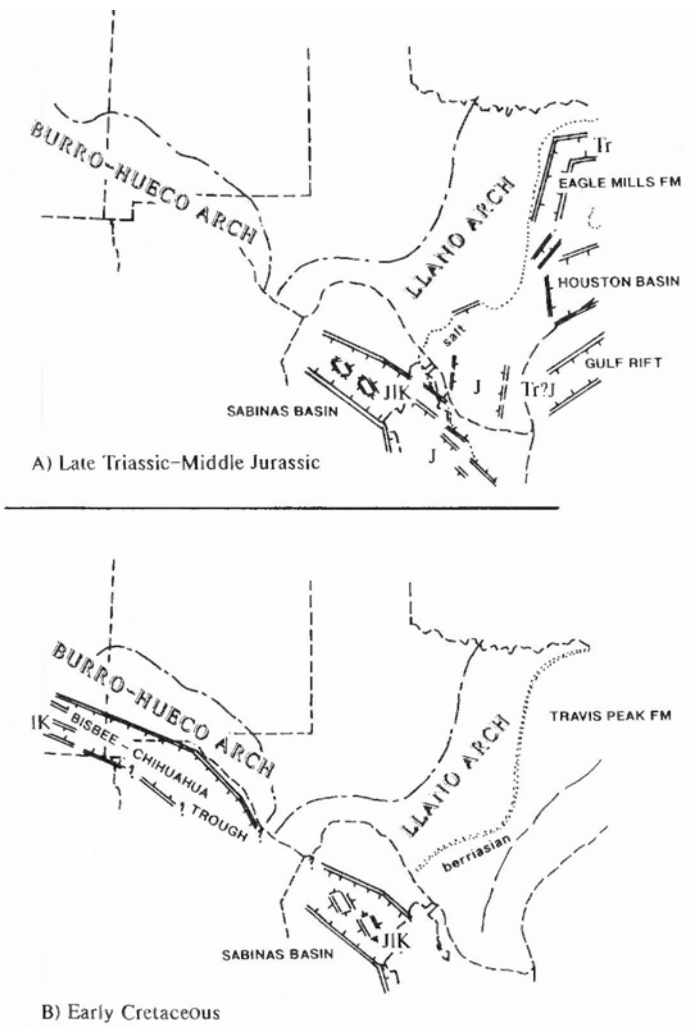

FIGURE 8. Possibilities for epeirogeny: A) Late Triassic–Middle Jurassic rifting of the Gulf of Mexico and Sabinas trough; B) Early Cretaceous rifting of the Bisbee-Chihuahua-Sabinas trough.

My best guess is that the pre-Cretaceous uplift in Trans-Pecos Texas is dominantly Early Cretaceous, due to uplift adjacent to the rifting Chihuahua and Bisbee Troughs. The less dramatic erosion in north and central Texas is dominantly Triassic-Jurassic, related to the Eagle Mills grabens. The tilting and erosion of the southern margin could be a combination of these two events; the Sabinas Basin to the south has evidence for both Jurassic sedimentation and Early Cretaceous block-faulting (Alfonso Zwanziger, 1978). The Chinati Sag suggests a break possibly indicating the boundary between the Bisbee-related and Sabinas-related flank uplifts.

Finding testable ways to more accurately locate, distinguish and date these regional erosive episodes will remain one of the most challenging tasks in the geology of the southwestern borderlands.

Complex intrabasin uplifts, known and unknown

Accepting these larger borders for the West Texas Basin before Mesozoic erosion, it is significant that the Diablo uplift and at least part of the Pedernal uplift (or Diablo-Pedernal axis) is intrabasin, just as is the Central Basin axis, which runs parallel from southeast New Mexico nearly to the Marathon front. The Central Basin axis is complex. Although it has been densely drilled, conflicting interpretations have resulted from the data. An interpretation involving a combination of northwest-directed left-lateral strike-slip faulting, east-west compression on north-trending axes and substantial (clockwise?) block rotation has been reached independently by three authors in recent years (Gardiner, 1990; Ewing, 1991; Shumaker, 1992).

The most important observation for the present argument, however, is the small size of the "average" Central Basin axis structural block. The most frequent style, that of asymmetric reverse-faulted ridges, involves blocks only 30 km long by 15 km wide. Trap-door structures and fold-thrust structures of the oblique slip zones are smaller, only 10-15 km across. Only a few larger blocks are known, notably the Fort Stockton uplift. All of these are smaller than the classic Laramide uplifts, which are more on the order of 100 by 60 km or larger. They are more similar in size to rift-valley blocks, such as the Franklin Mountains near El Paso, in the Rio Grande rift.

If both the Central Basin axis and the Diablo-Pedernal axis are intrabasin axes of deformation, could similarly complex styles be expected in the Diablo-Pedernal axis, which are now hidden because of poor exposure and lack of drilling?

The answer to this is unknown, but some factors lead me to be suspicious of internal complexity. The basement in the area is complex (see Soegaard, 1993, for the most current synthesis). A very extensive 1.35—Ga rhyolite province (with sediments), overlying unknown lower crust, was deformed and thrust northward in the Grenville orogeny, affecting the south half of the future West Texas Basin, generating foredeep sediments in some areas, with younger sequences and crustal fragments also involved. A regional 1.1-Ga igneous province includes rhyolites and granites in the El Paso area, which probably extended to the south. A 1 .1 Ga layered mafic complex also has been described from part of the Central Basin axis (Keller et al., 1989) and mafic complexes and granites are probably present in the Sacramento Mountains area (Soegaard, 1993). The widespread plutonism of this age suggests extensional tectonics, related to the "Keweenawan" of the Midcontinent, which could have pervasively fragmented the basement. The basement geology of at least the southern part of the Diablo-Pedernal axis is probably quite similar to the Central Basin axis.

The extent of the Diablo and Pedernal uplifts as usually drawn is larger even than the block-faulted uplifts usually encountered in basement-involved "Laramide" structural styles. Significant internal faulting, folding and basin development is nearly inevitable, especially if preexisting basement fracturing was pervasive. The complex small-block rotations and rapid variation of structural style documented in the Central Basin axis should cause any worker in the Diablo-Pedernal area to hesitate before generalizing the observations of a local area into a large regional picture!

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Guano bucket elevator near the entrance of Bat Cave (Carlsbad Cavern). Photograph by W. T. Lee, August 21, 1924. Courtesy of Southeastern New Mexico Historical Society of Carlsbad.