University of Nevada

Reno

Late Quaternary Tectonic Activity of the Meers Fault, Southwest Oklahoma

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Geology.

by

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April 1988

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April 1988

ABSTRACT

The Meers fault in southwestern Oklahoma is an active fault capable of producing large, damaging earthquakes. The most recent large event is late Holocene, occurring some 1,200 - 1,300 years ago, and it was preceded by one or more earlier Quaternary events. Few faults in stable continental interior (SCI) areas are known to be active, so this fault holds many implications for seismic hazards in these poorly understood regions.

Paleoseismic events probably had magnitudes of at least 6 3/4 to 7 1/4. Seismic events may be relatively larger in SCI regions and magnitudes of 7 1/2 or greater may be possible. The minimum scarp length is 37 km. Displacements have both left-lateral and high-angle reverse components. Vertical separation of the surface reaches about 5 m, while lateral separation exceeds vertical by a ratio of about 3:1 to 5:1, reaching approximately 20 m. Individual events appear to have had maximum surface displacements of several meters.

This fault may be part of a larger active zone. The Washita Valley and Potter County faults also have surface expressions believed to indicate recent surface faulting. No additional active surface faults have been recognized in the Meers fault area, but activity may be concealed by poor preservation or non-brittle surface deformation. Active faults are likely to be sparse and to rupture infrequently.

ACKNOWLEDGMENTS

I would like to extend thanks to my advisor, Burt Slemmons, for providing the opportunity for this project, for help and comments, and for patience in awaiting the completion of this thesis; to Nowell Donovan, Texas Christian University, Charlie Gilbert, Norm Tilford, and Diane Westen, Texas A&M University, Ken Luza, Oklahoma Geological Survey, Rich Madole and Tony Crone, U. S. Geological Survey, Steve Brocoum, U. S. Department of Energy, Ernst Zurflueh, U. S. Nuclear Regulatory Commission, Milton Stubbs and Curtis Ditzell, Oklahoma State University, Craig dePolo, Nevada Bureau of Mines and Geology, Mike Ellis and Paul Bodin, University of Nevada-Reno, participants in the Penrose Conference on the Southern Oklahoma Aulacogen, and others working in this area or on related topics for comments and discussions; to Mr. and Mrs. Charlie Bob Oliver, Mr. and Mrs. Lester Brown, Mrs. Hillary, and the many other landowners in the Meers area for hospitality and cooperation; to Bob Whitney, Kati Killeen, and Tom Redican for the help with the low-sun-angle photography; to Great Basin Aerial Surveys, Reno, Nevada, for all the help and excellent work on this and other projects; to Ken Smith for pinch-hitting in South Carolina; and to the U. S. Nuclear Regulatory Commission for financial support.

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INTRODUCTION

The Meers fault in southwestern Oklahoma (fig. 1) recently and unexpectedly was recognized as exhibiting a conspicuous fault scarp resulting from late Quaternary tectonic activity (Gilbert, 1983; Donovan and others, 1983). This potential source of moderate-to-large earthquakes adds another exception to the general rule that earthquakes of the central and eastern United States (CEUS) have magnitudes less than 5 1/2 to 6. This is a Paleozoic fault, has been historically aseismic (Lawson and others, 1979; Lawson and Luza, 1980-1987), and lies within a region thought to be tectonically stable, yet exhibits striking evidence of recent surface faulting. A prominent, linear scarp trends N60°W across an area of relatively flat terrain for a distance of almost 40 km. Features with such fresh and impressive geomorphic expression are usually restricted to more tectonically active regions. It is often stated that evidence of surface faulting is generally lacking in the CEUS and other stable continental interior (SCI) regions. The Meers fault scarp indicates that in at least some cases such evidence does exist, but it often may not be recognized.

The Meers fault scarp is extremely well expressed at the surface, but despite this fact, it escaped recognition as representing fault activity until very recently. Moody

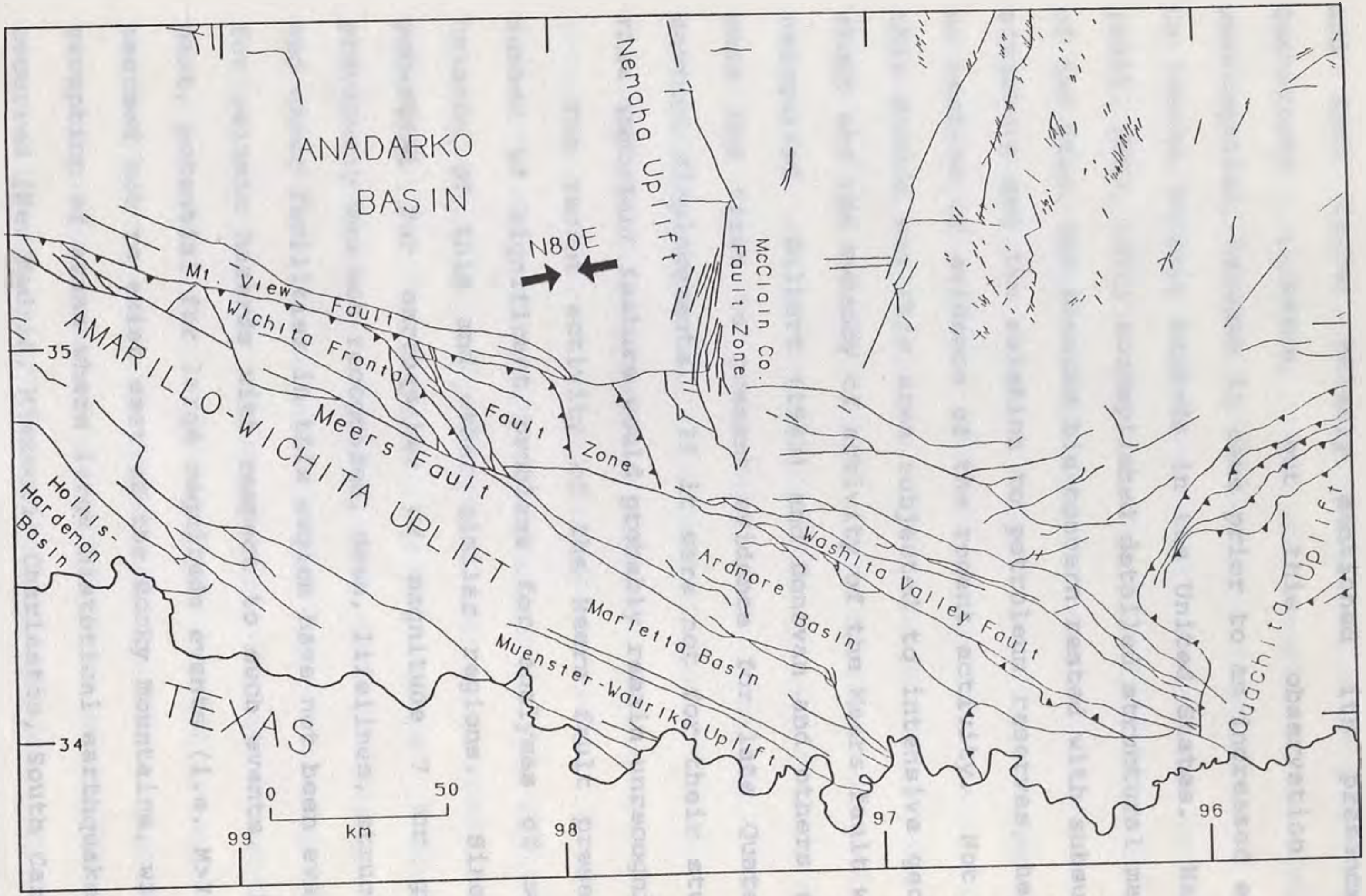


Figure 1: Principal structural features of southwestern Oklahoma. Direction of greatest horizontal stress from Dart (1987b). Fault mapping after Chenoweth (1983).

and Hill (1956) briefly mentioned its presence in Quaternary alluvium, but this observation went unrecognized, because it came prior to an increased effort to assess seismic hazards in the United States. Harlton (1951, 1963, 1972) accomplished detailed structural mapping of the area, but because his concern rested with subsurface structure and its relation to petroleum reserves, he made no mention of evidence of the recent activity. Not until this decade was this area subjected to intensive geologic study and the recency of activity of the Meers fault widely recognized. Gilbert (1983) and Donovan and others (1983) were the first to present evidence for late Quaternary surface displacements. If it were not for their studies, this important feature would probably remain unrecognized.

The recent activity of the Meers fault presents a number of significant problems for analyses of seismic hazards of this and other similar regions. Since the potential for earthquakes of magnitude 7 or greater previously was not recognized, dams, lifelines, structures, and other facilities in this region have not been evaluated for seismic hazards with respect to such events. In the past, potential for large magnitude events (i.e. M>7), was assumed not to exist east of the Rocky Mountains, with the exception of areas where large historical earthquakes have occurred (New Madrid, Missouri; Charleston, South Carolina; the St. Lawrence Seaway). Far-reaching possibilities are that the Meers fault is only part of a larger system of active faults and/or only one example of unrecognized active faults in the CEUS and other SCI regions.

This clearly expressed fault, almost 600 km east of the Rocky Mountains, is also important in that it is within a region where earthquake potential is determined principally by historical seismic activity, as opposed to incorporating active fault investigations as used near plate tectonic boundaries. The recognition of the Meers fault being an exception to this approach suggests that historical records often do not adequately reflect seismic potential in the CEUS and that accepted earthquake assessment methods need to be modified.

Several questions are apparent with respect to the recognition of activity of the Meers fault. What is the potential for a damaging earthquake along this fault? Have other potential sources of damaging earthquakes been overlooked or underestimated in SCI regions? What are the implications for areas where there have traditionally been no seismic design codes? Is this a one-of-a-kind feature, or does it indicate that traditional methods for evaluating seismic potential in SCI regions may be inadequate and misleading? An increased effort to identify and characterize paleoseismic surface ruptures throughout this region is warranted and could go a long way towards resolving these and other related issues.

Present evidence indicates that large seismic events in intraplate regions are generally confined to preexisting zones of weakness (i.e. ancient rifts, uplifts, and thrust belts). Such zones should be the primary focus of future investigations. The Meers fault is part of a large fault zone bordering the Amarillo-Wichita uplift, one such zone of weakness.

The Meers fault scarp has a very conspicuous surface expression resulting from late Holocene and Quaternary displacements. It owes this expression both to recency of offset and presence in Post Oak Conglomerate, a resistant rock unit (fig. 2). The two units that are faulted, Post Oak Conglomerate and Hennessey Shale, have highly contrasting surface expressions due to their differing resistance to erosion. The fault has a much more subdued scarp in the Hennessey Shale (fig. 3). At the present, this is the best known surface expression of a recent faulting event in the CEUS and illustrates well how active faults can escape recognition when they are unexpected. It is likely that there are other faults with unrecognized surface expression elsewhere in SCI regions.

It is commonly assumed that zones of active deformation exhibit some level of historical seismicity, but the Meers fault demonstrates this is not always the case. This region has had little historical seismicity, and none that can be specifically determined to be on the

Figure 2: Oblique aerial view (to NW) of Meers fault scarp in Post Oak Conglomerate, a resistant Permian carbonate boulder conglomerate. Contrast geomorphic expression with Figure 3. This results from both the greater resistance of the material and the greater degree of brittle displacement. (Photo by D. B. Slemmons)

Figure 3: Oblique aerial view (to NW) of Meers fault scarp in Hennessey Shale. This unit is which is much less resistant than Post Oak Conglomerate and has largely accommodated surface displacements by warping, rather than brittle offset. Accordingly, it has a more subdued geomorphic expression than that shown in Figure 2. (Photo by R. A. Whitney)

Meers fault. However, Gordon (in press) has relocated a number of historical earthquakes and has found several events to be associated with the Amarillo-Wichita uplift as a zone. He places the 6/17/59 (M-4.2) earthquake about on line of the Meers fault, about 24 km southeast of the end of the recent fault scarp (see section on historical seismicity). The record of felt earthquakes in southwestern Oklahoma dates back only about 85 years and the instrumental record only about 25 years (Lawson and others, 1979). This demonstrates very well that tectonically active faults can have temporal variations in seismicity, with quiescent intervals that are longer than historical records.

Various seismic zones have been interpreted to possess the potential for levels of seismicity exceeding that of surrounding regions. The Meers fault is considered to be part of the "Wichita-Ouachita" zone (e.g. Gordon and Dewey, 1985). Such interpretations do not address the seismic potential or size of events that could be expected for specific structures within a given zone. Seismicity data need to be supplemented by geological investigations, which should be concentrated along zones such as this, where fault orientations are favorable with respect to the stress field and/or a definable concentration of seismicity exists.

Implications of Activity

The recognition of recent activity on the Meers fault has several far-reaching implications for seismic hazards in stable continental interior (SCI) regions. This presents a potential source of strong ground motion in a region (south-central United States) generally thought to be tectonically stable and devoid of a potential for large magnitude earthquakes. A repeat of the last scarp-forming event would produce strong ground motion over vast parts of Oklahoma and Texas and be felt over most of the central United States. A recurrence interval for the Meers fault may be on the order of 10^4 years. This, combined with the recency of the last event, suggests that the probability of this occurring in the near future is quite low, although an accurate probability is not well constrained. Given the presumed low rate of deformation, a long period of stress accumulation is probably required before a similar sized event can be generated. It is less well understood to what extent recent activity increases the likelihood of events along adjacent parts of the Meers or other fault zones.

A second implication for seismic hazards is that the Meers fault scarp may be only one part of an actively deforming zone capable of producing large magnitude earthquakes along much of its length. Structures bounding the Amarillo - Wichita - Arbuckle uplift extend through most of southern Oklahoma and the Texas panhandle and may be linked with the Rio Grande rift system to the west, the Nemaha uplift to the north, the Matador arch to the south, and/or faults of the Ouachita fold and thrust belt to the east. An active zone would imply more frequent events over a larger area. Low-sun-angle aerial photographs obtained of faults along the northern edge of the Wichita uplift in southwestern Oklahoma show little indication of activity aside from the known rupture on the Meers fault (figs. 2 and 3), but various lines of evidence indicate a potential for activity does exist along other parts of the zone. Possible evidence of this situation includes numerous Pliocene to Holocene volcanics with WNW-trending source structures east of the Rio Grande rift valley in northeastern New Mexico (Dane and Bachman, 1965), historical seismicity and deformation of Pleistocene sediments in the Texas panhandle (McGookey and Budnik, 1983), probable late Quaternary tectonic activity of the Washita Valley fault in south-central Oklahoma (Cox and VanArsdale, 1986), and possible Holocene uplift of the Monroe uplift in northern Louisiana (Schumm, 1986).

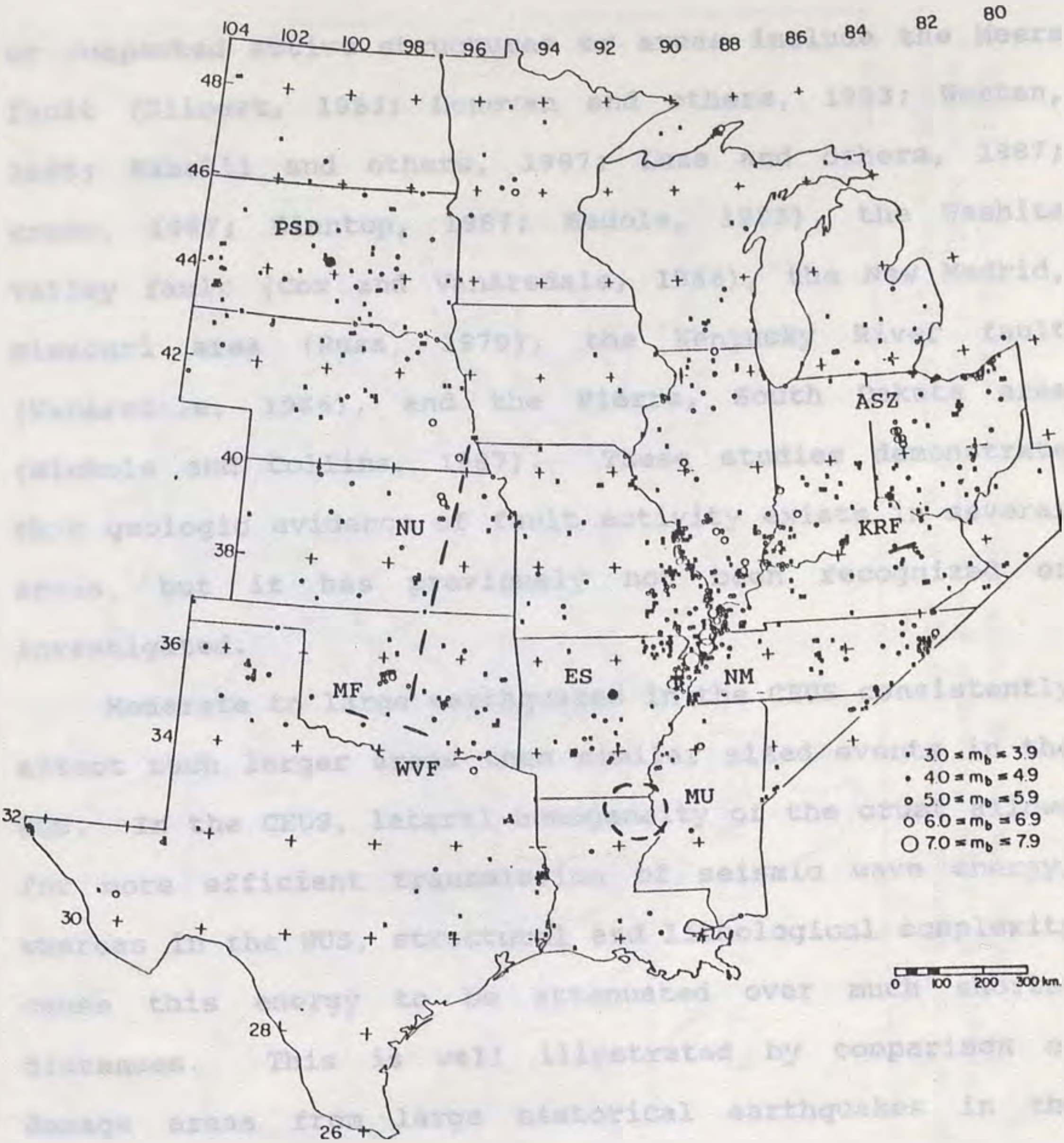
In tectonically active regions, most significant faults are interconnected or branch from major fault zones, suggesting a high degree of fault interaction. If this holds true for fault systems in SCI regions, activity along these structures may be linked to extensions of the New

Madrid seismic zone, and in turn to such faults as the Kentucky River fault system (VanArsdale, 1986). On the other hand, SCI earthquakes may typically result from failure along preexisting crustal flaws solely in response to far-field stresses and not require a great deal of fault interaction. Much more work throughout this region is required before such potential relations can be fully evaluated.

A more far-reaching implication lies in the possibility of additional active structures in other SCI regions and their impact on seismic hazard analyses. If a feature as prominent as the Meers fault scarp can escape recognition for so long, it seems likely that less obvious features have likewise been overlooked. Methods of seismic hazard analysis in the CEUS have greatly differed from those in the western United States (WUS), where the existence of numerous active structures has long been recognized. Relatively high rates of activity in the WUS have created distinctive landforms, which in many cases (e.g. the Basin and Range province) are well preserved in regions with arid climates. This, along with the occurrence of several large earthquakes over the last century, has led to a methodology for seismic hazard analyses whereby faults or fault zones are individually evaluated for their potential for producing large magnitude earthquakes. For design purposes, a magnitude value derived from determined fault parameters (i.e. fault or fault zone length, displacements, geologic or structural setting, historical seismicity, etc.) is considered for its ground motion potential at the site of concern.

In contrast, rates of tectonic activity in the CEUS are much lower. There have been no large earthquakes in this region over the last century, but we know that they do occur, as evidenced by the remarkable series of three great earthquakes near New Madrid, Missouri in 1811-1812. However, surface rupturing has rarely been documented during even large SCI earthquakes. Worldwide, only nine cases of historical surface rupture in SCI regions have been reliably documented (Coppersmith and others, 1987), and one of these, the 1819 Kutch, India earthquake should probably not be included. Seismic hazard analyses have thus not focused on fault specific studies, but rather are based on extrapolations of historical seismicity and/or general structural associations. Such methods can lead to a greatly underestimated seismic potential when the historical seismicity record is inadequate, as is often the case, or when structural associations are poorly understood.

It is often stated that evidence of large prehistorical earthquakes in the CEUS is generally lacking. However, several studies in this region have revealed evidence suggesting paleoseismic activity (fig. 4). Known



ASZ - ANNA SEISMIC ZONE

ES - ENOLA SWARM

KRF - KENTUCKY RIVER FAULT

MF - MEERS FAULT

MU - MONROE UPLIFT

NM - NEW MADRID

NU - NEMAHA UPLIFT

PSD - PIERRE, SOUTH DAKOTA

WVF - WASHITA VALLEY FAULT

Figure 4: Historical seismicity and areas of known or suspected late Quaternary tectonic activity in central United States (after Nuttli, 1979).

or suspected active structures or areas include the Meers fault (Gilbert, 1983; Donovan and others, 1983; Westen, 1985; Ramelli and others, 1987; Luza and others, 1987; Crone, 1987; Kientop, 1987; Madole, 1988), the Washita Valley fault (Cox and VanArsdale, 1986), the New Madrid, Missouri area (Russ, 1979), the Kentucky River fault (VanArsdale, 1986), and the Pierre, South Dakota area (Nichols and Collins, 1987). These studies demonstrate that geologic evidence of fault activity exists in several areas, but it has previously not been recognized or investigated.

Moderate to large earthquakes in the CEUS consistently affect much larger areas than similar sized events in the WUS. In the CEUS, lateral homogeneity of the crust allows for more efficient transmission of seismic wave energy, whereas in the WUS, structural and lithological complexity cause this energy to be attenuated over much shorter distances. This is well illustrated by comparison of damage areas from large historical earthquakes in the United States (fig. 5). The 1906 San Francisco earthquake is the largest earthquake to occur in the contiguous United States this century, but the area it affected is far less than the areas affected by large earthquakes in the eastern half of the country.

It is currently impossible to precisely estimate just how large the recent Meers fault earthquakes have been, but

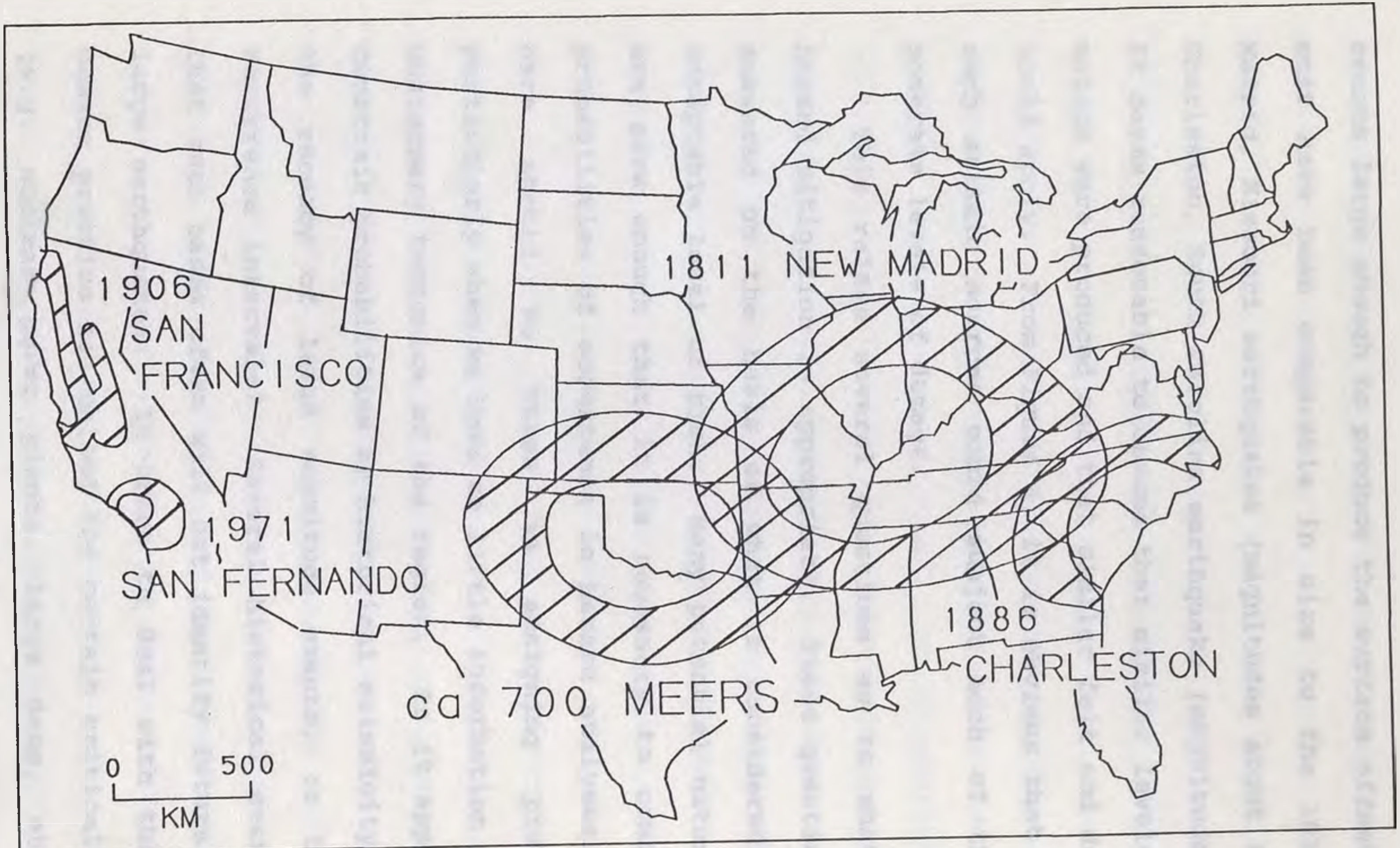


Figure 5: Comparison of damage areas during large historical earthquakes and hypothetical effect of late Holocene Meers fault event. Inner areas represent MMI $\geq$ VII, outer hachured areas represent MMI VI-VII (after Nuttli, 1979).

events large enough to produce the surface offsets observed must have been comparable in size to the 1811-1812 New Madrid, Missouri earthquakes (magnitudes about 8) and 1886 Charleston, South Carolina earthquake (magnitude about 7). It seems reasonable to assume that similar levels of ground motion were produced and that similar felt and damage areas would apply. From Figure 5, it is obvious that only a few such seismic sources could subject much of the CEUS to moderate levels of damage.

This raises several questions as to what level of hazard mitigation is appropriate. These questions must be answered on the basis of what is considered to be an acceptable level of risk. Many potential natural hazards are rare enough that it is necessary to consider their probabilities of occurrence in hazard analyses. However, care should be taken in assigning probabilities, particularly when we have so little information on the late Quaternary tectonics of the region. Is it appropriate to constrain probabilities by historical seismicity levels, by the recency of large magnitude events, or by apparent recurrence intervals? Several historical events suggest that such bases often will not identify future sources of large earthquakes. In order to deal with this problem, common practice has called for certain critical structures (e.g. nuclear power plants, large dams, etc.) to be designed for such events, with the low probability of occurrence taken into consideration for less critical structures.

Historical Seismicity

Historical seismic activity in the area around the Meers fault has been low, but the historical period in southern Oklahoma is very short. It is unlikely that any moderate to large events were undetected subsequent to the establishment of Fort Sill in 1869. However, from 1902 to the present, Fort Sill has been used extensively as a bombing range, so it is conceivable that small events could have been mistaken for artillery practice. Settlement in the area did not begin until about the turn of the century, when the Wichita Mountains were explored for their mineral potential (Ellenbrook, 1984).

Instrumental seismic coverage in Oklahoma began in 1961, with one seismographic station near Tulsa and another in the Wichita Mountains. Although the latter one was located only 7 km south of the Meers fault and was in operation from December, 1961 to June, 1971, it was operated by the United States Air Force for the purpose of distinguishing distant earthquakes and underground nuclear tests. Since high-frequency waves were partially filtered out, it was used little for local earthquake studies

(Lawson and others, 1979). No earthquakes were recorded within a 55 km radius of this station during this ten year period. The seismographic coverage of Oklahoma improved greatly when several stations were established from 1976 to 1978. One of these stations was opened in July, 1977 at Quartz Mountain State Park, about 60 km to the west of the Meers fault scarp. This station has operated to the present. As a result of the recognition of activity on the Meers fault, a station was installed on the Kimbell Ranch, 2.0 km to the north of the fault. This station was in operation from April 20, 1984 through May, 1985. On June 11, 1985, a replacement station was placed in the Meers Store, 3.8 km south of the fault (Lawson and Luza, 1986). These stations have continued to document the seismic quiescence of the Meers fault.

Historical earthquakes of magnitude 3 or greater and/or Modified Mercalli Intensity V or greater have been compiled for southern Oklahoma through 1986. These events are listed in Table 1 and represented in Figure 6. The early, pre-instrumental events have been assigned locations based on felt reports and therefore can be associated with structures only in a very general way. Those events referenced to Gordon (in press) have been relocated by the joint hypocenter determination method. These locations are surely much better than those previously available and show much better correlation with known structures.

TABLE 1: Historical Earthquakes in southern Oklahoma; M $\geq$ 3 or MMI $\geq$ 5

Date	Time	Lat.	Long.	Mag.	MMI	ref.
09/10/18	16 30	35.5	98.0		V	2
09/11/18	06 30	35.5	98.0		VI	2
09/11/18	09 00	35.5	98.0		VI	2
12/28/29	00 30	35.5	98.0	3.7ML	VI	2
08/19/33	19 30	35.5	98.0		V	2
04/11/34	17 40	33.9	95.5	3.2ML	V	2
03/14/36	17 20	34.0	95.0	2.4ML	V	2
10/18/41	07 48	35.4	99.0	4.2mb	V	3
03/20/50	13 24	33.5	97.1	3.8mb	IV	3
04/09/52	16 29 28.4	35.525	97.850	5.0 F		1
05/16/52	06 05	35.4	97.8		V	2
03/17/53	13 12	35.4	98.0		V	2
03/17/53	14 25	35.4	98.0		VI	2
06/06/53	17 40	34.7	96.7	3.8mb	IV	3
04/02/56	16 03 18	34.2	95.6		V	2
06/15/59	12 45	34.8	96.7	2.7ML	V	2
06/17/59	10 27 10.6	34.639	98.055	4.2 F	VI(3)	1
02/02/64	08 22 43.8	35.306	99.606	2.9mbLg	V(2)	1
10/14/68	14 42 54	34.0	96.4	3.5M3Hz	VI	2
04/13/69	06 27 51	34.2	96.3	3.5mb		3
09/13/75	01 25 05.6	34.131	97.221	3.2mbLg		1
11/29/75	14 29 44.9	34.681	97.421	3.5mbLg		1
03/14/79	04 37 15.27	35.519	97.781	2.2mbLg	V	4
06/07/79	07 39 36.3	35.216	99.759	3.0mbLg		1
07/25/79	03 15 37.27	33.967	97.549	2.7mbLg	V	4
09/13/79	00 49 21.5	35.193	99.473	3.4mbLg		1
11/27/79	09 10 36.79	35.630	98.408	3.3mbLg		4
05/30/80	07 44 02.72	35.512	99.390	3.0M3Hz		4
07/18/80	14 29 46.88	35.180	99.698	3.2mbLg		4
11/02/80	10 00 49.03	35.429	97.777	3.0mbLg	V	4
07/11/81	21 09 21.84	34.853	97.732	3.5mbLg	V	4
05/03/82	07 54 48.65	33.990	96.473	3.1mbLg	VI	4
12/19/82	05 15 42.94	34.891	97.584	3.1M3Hz		4
01/24/84	15 34 09.63	35.033	96.366	3.1M3Hz	V	4
02/03/84	04 38 28.04	34.665	97.356	3.2mbLg	V	4
11/20/84	10 57 31.98	34.707	97.410	3.1mbLg	IV	4
05/05/85	01 39 30.78	34.664	97.529	3.0M3Hz	felt	4
05/06/85	02 11 16.16	34.969	97.482	2.3mbLg	V	4
12/31/85	18 27 26.12	34.703	97.459	3.0M3Hz		4

References (unless otherwise noted):
1 = Gordon (in press) 2 = Lawson and others (1979)
3 = Nuttli (1979) 4 = Lawson and Luza (1980-1987)

Notes: F = felt area magnitude (Nuttli and Zollweg, 1974)

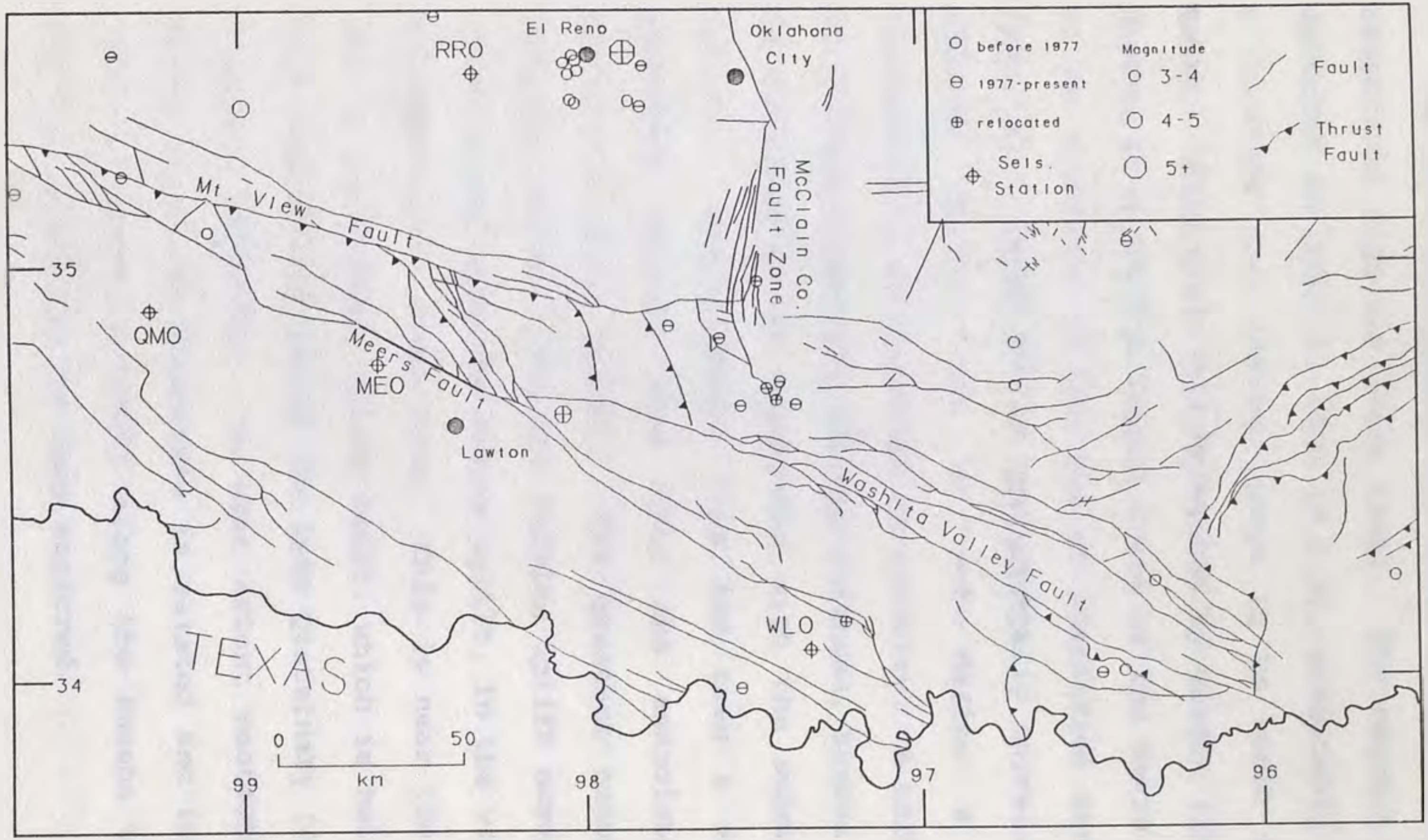


Figure 6: Historical earthquakes in southern Oklahoma through 1986 (magnitude ≥ 3 or Modified Mercalli Intensity ≥ V); see Table 1. Fault representation after Chenoweth (1983).

No felt earthquakes are known to have been definitely associated with the Meers fault. The nearest event, which occurred on June 17, 1959 (M 4.2), originally was assigned a location near Lawton, about 15 km south of the fault scarp. This event was relocated by Gordon (in press), who placed it along the mapped trace of the Meers fault, about 24 km southeast of the end of identified surface rupture (fig. 6). Other events are spatially correlated with the Wichita uplift, but do not define a conspicuous concentration or alignment of activity. A north-south band of diffuse seismicity through Oklahoma, Kansas and Nebraska is believed to be associated with the subsurface Nemaha uplift. Historically, this has been a somewhat more prominent seismic zone than one associated with the Amarillo-Wichita uplift. The greatest number of recent recorded events along the Wichita uplift occurred near its intersection with the Nemaha uplift, in the vicinity of the McClain County fault zone. This is near the northwestern end of the Washita Valley fault, which is believed to also have been active during the late Quaternary (see section on regional activity). To what extent neotectonic activity along these two structures is related and implications of current seismic activity along the Nemaha uplift on the Wichita uplift has not been explored.

STRUCTURAL AND REGIONAL RELATIONS OF THE MEERS FAULT

Geologic/Structural Setting of the Meers Fault

Even though the Meers fault has been recently active, considerable deformation has not occurred in this region since the late Paleozoic. This region is part of the craton, or stable continental interior (SCI), of the North American continent. The evidence of recent large earthquakes on the Meers fault is rather anomalous, since such regions are usually characterized by a relatively low seismic potential.

The Meers fault is generally considered to be the south-bounding fault of the Wichita frontal fault zone, which separates the Anadarko Basin, the deepest intracontinental basin in the United States, from the uplifted igneous complex of the Wichita Mountains, an exposed part of the mostly subsurface Amarillo-Wichita uplift. As much as 12 km of stratigraphic separation occurs across this zone (Donovan, 1986). The Anadarko Basin is asymmetric, with a depositional axis immediately adjacent and parallel to the Wichita frontal fault zone. Subsidence occurred throughout much of the Paleozoic, although most of it took place during late Cambrian and Ordovician time (Donovan, 1986).

During Pennsylvanian and Permian time, the closing of the proto-Atlantic ocean gave rise to the Ouachita orogen to the east and reactivated the Wichita frontal fault zone (McConnell, 1986). Latter stages (Permian) of this deformational event involved a component of down-to-the-south throw on the Meers fault, with Post Oak Conglomerate shed off the uplifting Wichita Mountains and Slick Hills into a narrow, elongate trough, which is roughly coincident with the present-day Meers Valley (Donovan, 1986).

In the Slick Hills, adjacent to the northwestern part of the Meers fault scarp, about one-fourth of the total width of the Wichita frontal fault zone (6 km out of a 24 km total) is exposed at the surface. Structures exposed in this area and in the adjacent Wichita Mountains are interpreted to have resulted from transpressive deformation, with a considerable amount of left-lateral deformation. Evidence for this is shown by large- and small-scale features such as wrench folds, fault displacements, fracture patterns, slickensides, vein systems, and Riedel shears (Butler, 1980; Donovan, 1982; Donovan and others, 1982; Beauchamp, 1983; McConnell, 1983; McLean and Stearns, 1983; Donovan, 1986; McCoss and Donovan, 1986). These studies have shown the Meers fault to be a major strike-slip fault associated with this deformation. Features observed along the present exposure of the Meers fault (e.g. horizontal slickensides, linear fault trace, joint pattern, Riedel shears, proximity to the

Post Oak Conglomerate/Arbuckle Group contact) strongly support this interpretation.

A lateral deformational phase, in which the Meers fault played a major role, apparently followed a shortening phase (Donovan, 1986). The recent displacements have clearly occurred along an extremely linear fault trace (fig. 7). Nowhere does the strike of the fault scarp vary by more than a few degrees from its overall N60^{o}W trend. This linearity persists through the Slick Hills, an area of slight relief (i.e. tens of meters), indicating that the fault is very nearly vertically oriented. Shallow seismic-reflection surveys (Harding, 1985; Myers and others, 1987) have suggested that the fault is steeply dipping throughout the shallow subsurface in this area. The preservation of this linearity suggests significant lateral displacement was occurring during late stages of the Paleozoic deformation. As much as 150 km of left-lateral displacement across the Amarillo-Wichita uplift as a zone has been proposed (Budnik, 1986a).

The Wichita frontal fault zone has long been depicted with a vertical orientation extending to depth (e.g. Harlton, 1951, 1963, 1972; Ham and others, 1964). However, recent studies, based principally on deep seismic profiling surveys and drill-hole data, have interpreted this fault system to be south-dipping, with the Wichita Mountains thrust over sediments of the Anadarko Basin (Brewer and

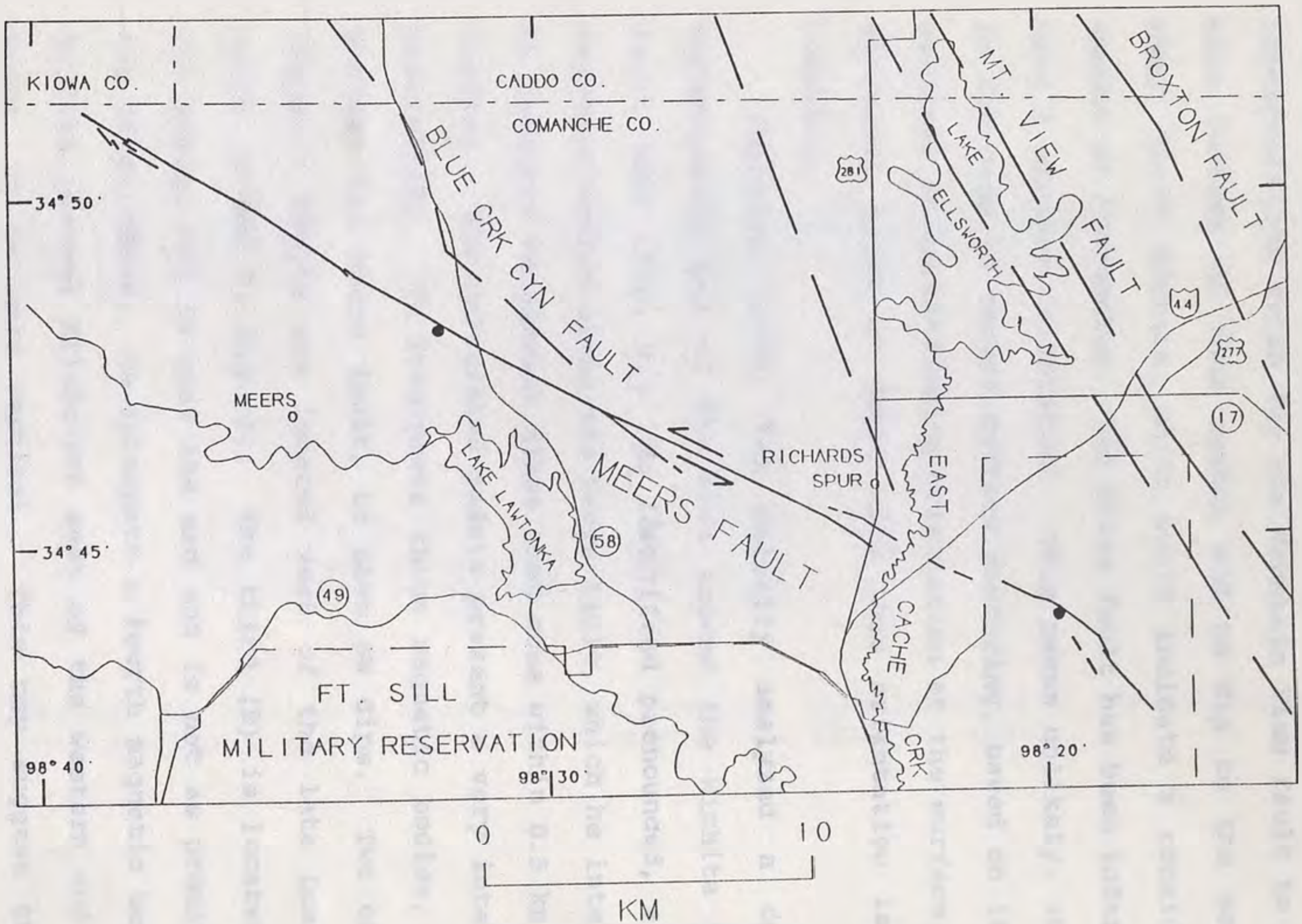


Figure 7: Surface rupture map of recent Meers fault surface displacements. Balls on downthrown side of fault; intersecting structures as mapped by Harlton (1972).

others, 1983 [fig. 8]; Brown, 1984; McConnell, 1986). This interpretation calls for the Mountain View fault to be the main feature of this system and to dip to the south at about 30-40 degrees, which would indicate a considerable amount of shortening. The Meers fault has been inferred to have a similar orientation. This seems unlikely, at least in the area of recent surface rupturing, based on its sub-vertical to north-dipping orientation at the surface, but a different fault at depth with this orientation is quite possible.

Purucker (1986) has recently analyzed a detailed aeromagnetic map of the area around the Wichita frontal fault zone (fig. 9). He identified pronounced, elongate magnetic bodies along the Meers fault, which he interpreted to possibly be diabase dikes that come within 0.5 km of the surface. His calculated models present a very interesting possibility. He interprets three magnetic bodies, and by analogy the Meers fault, to have SW dips. Two of these magnetic bodies are located west of the late Quaternary scarp (C and D; fig. 9). The third (B) is located along the scarp, but is near the end and is not as prominent as the other three. He interpets a fourth magnetic body (A), located several kilometers east of the western end of the scarp, to be near vertical. This may suggest that the recent activity has been confined to an area where the dip of the fault makes it favorably oriented for reactivation.

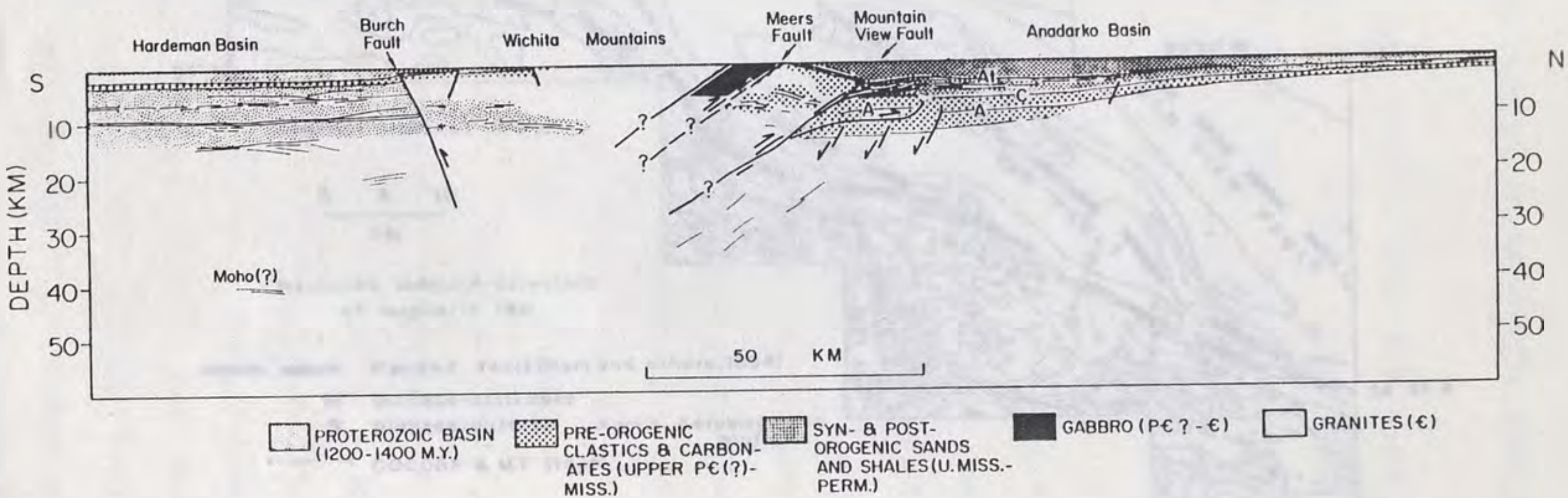


Figure 8: Geologic cross-section across Wichita Uplift, based upon interpretation of seismic-reflection profiles (Brewer, 1983).

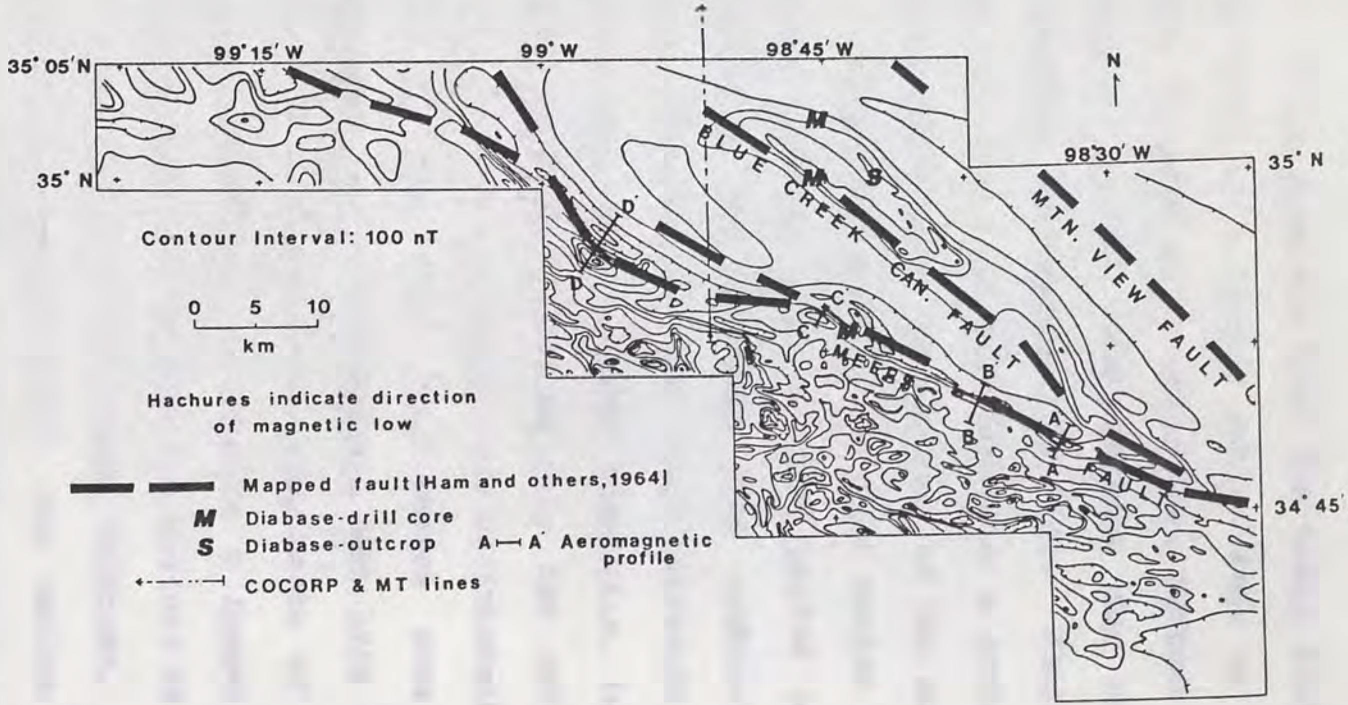


Figure 9: Aeromagnetic map of Meers fault area (from Purucker, 1986).

Regional Extent of Activity: the Amarillo-Wichita Uplift

It is possible that the Meers fault scarp represents only part of a larger active zone of deformation. This fault is part of a structural feature (Amarillo-Wichita-Arbuckle uplift) that extends for several hundreds of kilometers in length (fig. 10). Fault activity along a part of the zone may indicate a potential for activity exists along much, if not all, of the zone.

Stress conditions in this region have probably been fairly uniform over a long period of time. Activity apparently has resulted from combinations of stresses, fault orientation, and fault strength that have existed much longer than "recent" time (i.e. late Quaternary). It seems possible that activity has continued over a long time, at very low rates, and at intermittent intervals. If this is the case, the entire zone could potentially generate large earthquakes, but late Quaternary activity might be present only along parts of the zone. It thus becomes necessary to examine a longer period of geologic time in order to unravel the history of tectonic activity.

For most of southern Oklahoma, including the area around the Meers fault, the entire sedimentary record between Permian and recent times has been removed by erosion. This makes examination of fault history extremely difficult. This is not as large of a problem in the Texas

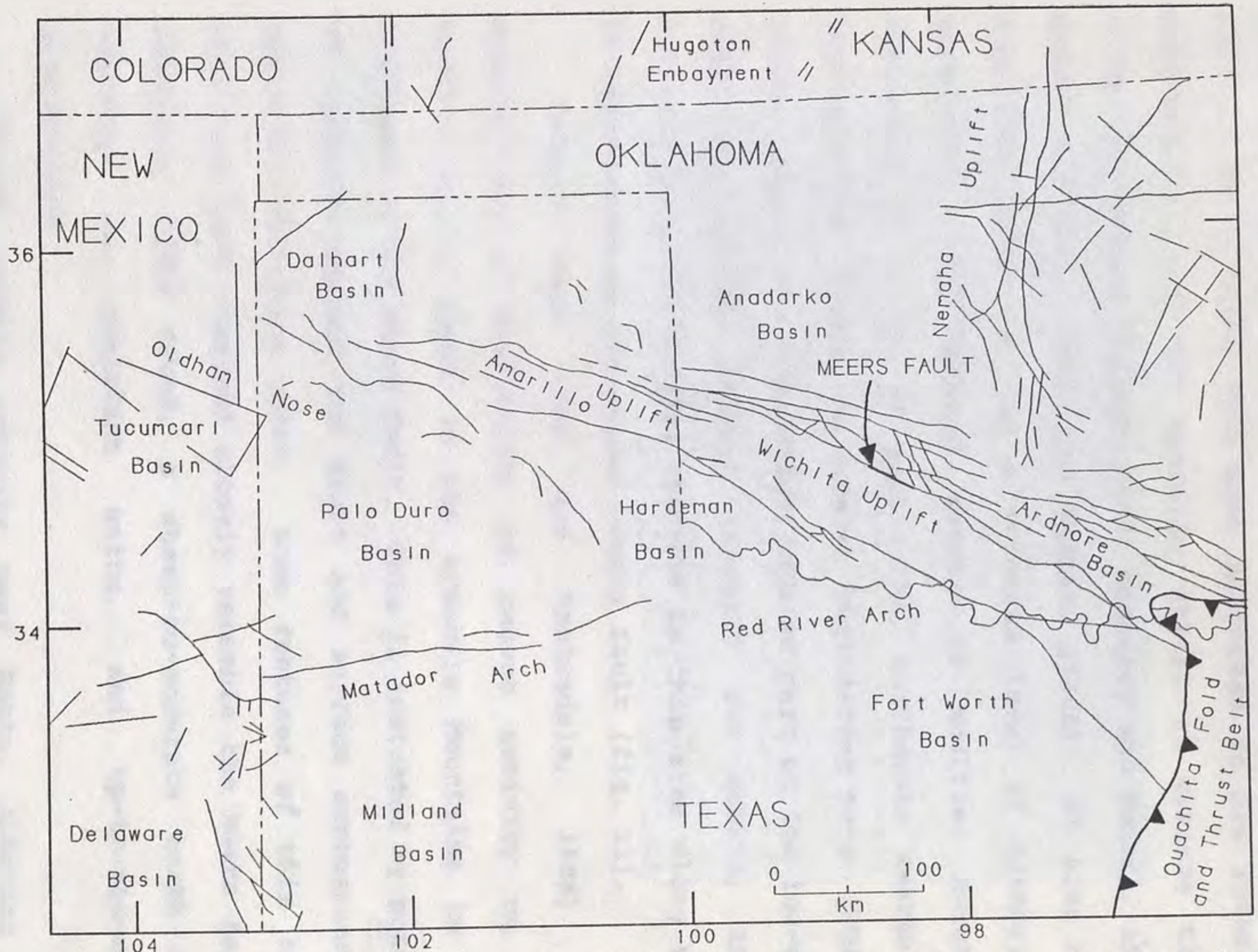


Figure 10: Faults and principal structural features of Meers fault region (after U.S. Department of Energy, 1986).

Panhandle, where there is a record of various Mesozoic and Tertiary units (Texas Bureau of Economic Geology [TBEG], 1969). These units have been subjected to low rates of deformation along the Amarillo uplift at various times since the Permian (Budnik, 1983; McGookey and Budnik, 1983; Budnik and Davis, 1985; and Budnik, 1986b). An area near Amarillo, Texas has had a moderate level of historical seismicity, with several events of Modified Mercalli Intensity IV to VI, in particular, earthquake swarms in 1917 and 1925 (Stone and Webster Engineering Corp. [SWEC], 1983). Aerial reconnaissance done as part of the low-sun-angle photography project (Ramelli and others, 1987) revealed possible surface rupture in this area along what is believed to be the Potter County fault (fig. 11).

Recent work (Cox and VanArsdale, 1986) has demonstrated a possibility of recent activity on the Washita Valley fault in the Arbuckle Mountains to the southeast of the Meers fault. This is indicated by ponding of alluvium against the fault and surface expression in resistant carbonate rocks. Some features of this fault that have been observed closely resemble the Meers fault, including fault trend, a steep-to-moderate north dip, exposure in resistant units, and up-to-the-north displacement.

Recent seismic activity near Enola, Arkansas has occurred (Haar and others, 1983) along the northern edge of

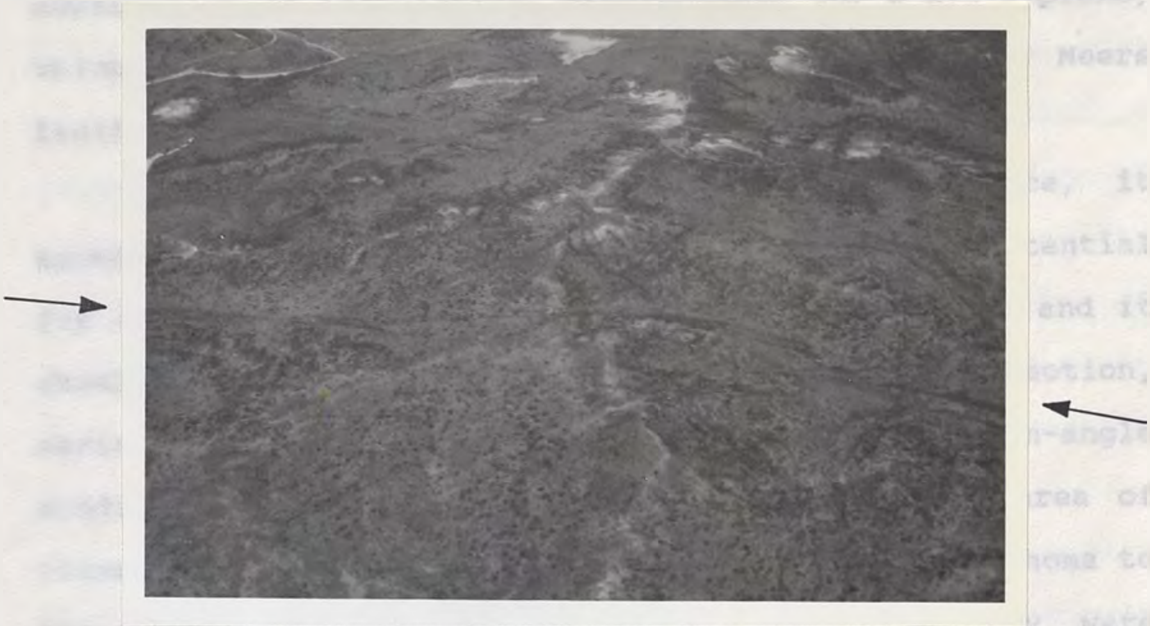


Figure 11: Aerial view (to NNW) of possible Quaternary surface rupture along Potter County fault, Texas panhandle.

the Ouachita fold and thrust belt, which truncates the Amarillo-Wichita-Arbuckle uplift. A composite focal mechanism derived from this activity is compatible with the observation of left-lateral displacement for a N70^{o}W plane, which is very near the N60^{o}W orientation of the Meers fault.

Considering these various pieces of evidence, it appears likely that this zone may have a finite potential for activity along much, if not all, of its length and it should be studied in detail. As a step in this direction, aerial photographs were flown under low-sun-angle conditions in conjunction with this study. The area of coverage extended for 265 km, from near Retrop, Oklahoma to the Arbuckle Mountains. Results of this study were presented by Ramelli and others (1987). Aside from the Meers fault scarp, no obvious indications of tectonic activity were revealed. It appears either that conditions such as climate, surface materials, and recurrence intervals do not allow good preservation of geomorphic features indicative of neotectonic activity, or that activity is not evenly distributed along the zone, with only a few areas of recent activity. Both possibilities are likely to hold true to certain extents. Much more study is needed to determine the extent and nature of neotectonic activity in this region.

QUATERNARY FAULT DISPLACEMENT

Surface Rupture Length

Fault rupture during large magnitude events (i.e. M>6) generally propagates to the surface, with the exception of subduction zone events. The surface manifestation of earthquakes can tell us a great deal about the size of the event, structural relations of the fault, style of faulting, and stresses in the crust. Surface rupture along the Meers fault, as determined from low-sun-angle photography and fieldwork, is depicted in Figure 7.

The prominent expression along almost the entire length of the Meers fault scarp allows for relatively simple length determination. Gilbert's (1983) initial reported length of 26 km was essentially unchanged until low-sun-angle aerial photographs revealed a significant extension to the southeast (Ramelli and others, 1987). The main fault scarp forms a prominent, linear topographic break extending from the Kiowa/Comanche County line to Highway 281, near Richard's Spur, where it intersects the very active floodplain of East Cache Creek (fig. 7). The northwestern end is well defined on standard aerial photography and while subsequent work has determined rupture at this end to be slightly more complex than originally recognized, no significant extension was found.

However, to the southeast, the recognition of a fairly distinct scarp trending across the northeastern part of Fort Sill Military Reservation (fig. 12) increased the minimum surface rupture length to 37 km.

A number of aspects about the Fort Sill scarp provide evidence that it is fault controlled. It is located on-line with, and in close proximity to, the main fault scarp; it has a WNW-orientation and is down on the south side, similar to the main fault scarp; and it is oriented transverse to local drainages. This scarp is clearly discernable beginning at about 4 km to the east-southeast of East Cache Creek and trending across most of the northeastern part of Fort Sill Military Reservation. Its location across an "impact area" for artillery practice will unfortunately limit access and study of this part of the fault. The eastern end of this scarp lies about 1 km from the eastern border of Fort Sill. Connection of the main fault trace and the Fort Sill scarp is not completely clear-cut. There is a moderate (1/3 km) left-step from the main trace to the trace of the Fort Sill scarp in the vicinity of East Cache Creek. Most, if not all, surficial evidence of faulting in this area is concealed by recent floodplain deposition.

Subdued lineaments evident on the low-sun-angle aerial photographs cross much of the East Cache Creek floodplain, but fault control of these is problematic without

Figure 12: Aerial view (to ENE) of fault scarp trending across the northeastern part of Fort Sill Military Reservation. (Photo by R. A. Whitney)

additional exploration. Field observations in this area revealed a probable, subdued scarp at U. S. Hwy. 281 and a wide fracture zone with nondefinitive offsets on the east side of Interstate 44. No definitive evidence of surface faulting has been observed along this line across the East Cache Creek floodplain, but this is a very active floodplain and small scarps could easily be erased in a short time. Along extension of and parallel to the Fort Sill scarp and the East Cache Creek floodplain lineaments is a subdued, linear south-facing scarp. It is presently not clear to what extent this scarp is fault controlled, but its location and linearity suggest that it is tectonic.

As shown in Figure 7, the three largest discontinuities in the surface fault trace lie along the eastern half of the scarp. The western-most of these is a splay from the main fault. While not obvious on aerial photographs, this splay was mapped by Harlton (e.g. 1972) and forms a slight topographic break. The middle discontinuity, the step-over at East Cache Creek, is the largest of the three. The eastern-most discontinuity is a bend near the end of the fault. These irregularities lie across an area where several NW- to NNW-trending faults (Mountain View, Blue Creek Canyon, Broxton, Apache, etc.) are mapped as intersecting the Meers fault (see fig. 7). It is probable that interaction with these faults is responsible for these discontinuities.

Recently, "segmentation" of fault zones has been increasingly used to better characterize paleoseismic events (e.g. Wheeler, 1987). This has grown out of recognition that in many cases fault slip at a given location is fairly consistent from event to event (i.e. characteristic earthquakes; Schwarz and Coppersmith, 1984) and that discontinuities within a fault zone often mark initiation and termination points of individual ruptures. In defining rupture segments, it is attempted to delineate those features that reflect the starting or stopping points of individual ruptures. These are commonly significant bends, step-overs, or other fault zone discontinuities (e.g. Knuepfer and Coppersmith, 1987). Fault zones have many discontinuities that do not act as ends of ruptures, however. It is rarely obvious which are boundaries (or "barriers") between more or less independent ruptures. Segments are sometimes described based solely on a geometric or structural basis, without regard to whether they separate discrete fault ruptures. This has lead to some confusion regarding the term, so it should be explained how the term is meant when it is used.

Features along the southeastern part of the Meers fault scarp possibly reflect boundaries between separate rupture segments (the step-over at East Cache Creek, in particular), but this seems unlikely for the following reasons: 1) this would require the Fort Sill scarp to be formed with a displacement of at least a few meters, but a length of only about 10 km; 2) vertical displacement curves (fig. 13) suggest that displacements do not die out at East Cache Creek; and 3) this would further complicate the displacement vs. length problem discussed in Ramelli and others (1987) and in the section on seismic potential in this paper.

Sense of Displacement

The Meers fault scarp has conspicuous surface expression with the north side consistently upthrown. This led to some early speculation that displacement may be normal, based on a south-dipping fault as postulated by Brewer and others (1983). However, subsequent study showed that, at least at the near surface, the dip of the fault plane ranges from vertical to moderately northward dipping. The combination of fault orientation and existing compressional stress seems to preclude normal faulting.

Throughout that section of the fault that offsets Post Oak Conglomerate at the surface, the fault is vertically oriented. To the east, two trenches dug by the Oklahoma Geological Survey showed a 20 to 50 cm wide fault zone dipping moderately (about 50° to 60°) to the north (Crone and Luza, 1986a), but this dip is likely a shallow, near surface effect from behavior of the surficial alluvium and Hennessey Shale. The fault likely steepens within a few to several meters below the surface. This phenomenon has been shown in experimental and theoretical work by Lade and Cole (1984). At "Brown's Creek," further to the east, the fault dips about 80° to the north.

Displacement measurements considered to represent the cumulative late Quaternary offsets (figs. 13 and 14) have been previously presented (Ramelli and Slemmons, 1986; Ramelli and others; 1987). These measurements are taken from mismatches of surficial features across the scarp, primarily offset ridgelines. Relatively accurate displacements can be determined due to the discrete, narrow zone of deformation, the recency of scarp formation, and, in the case of Post Oak Conglomerate, the resistant nature of the faulted material. The scarp reaches a height of about 5 meters, with a 2-3 meter scarp height being common. The amount of lateral displacement, however, has been a point of contention.

Tilford and Westen (1985a) suggested that lateral deflections of surficial topography could be relicts of older displacements now evident in an exhumed topography. However, scarp profiles (see appendix A) illustrate that the Meers fault scarp has an expression similar to late Quaternary fault scarps in other regions (e.g. Basin and Range Province). It is unlikely that such a feature could

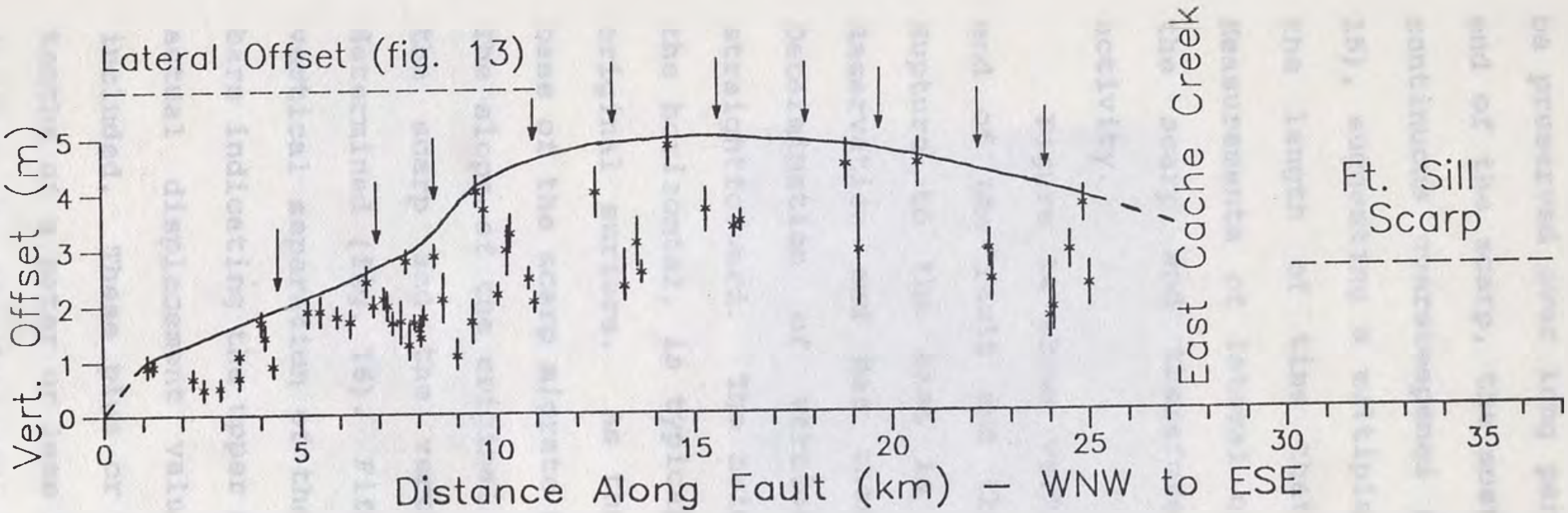


Figure 13: Graph of vertical displacement along Meers fault scarp from NW to SE. Note range and scale difference from fig. 14. Arrows indicate locations of streams. Post Oak/Hennessey contact at about 12 km.

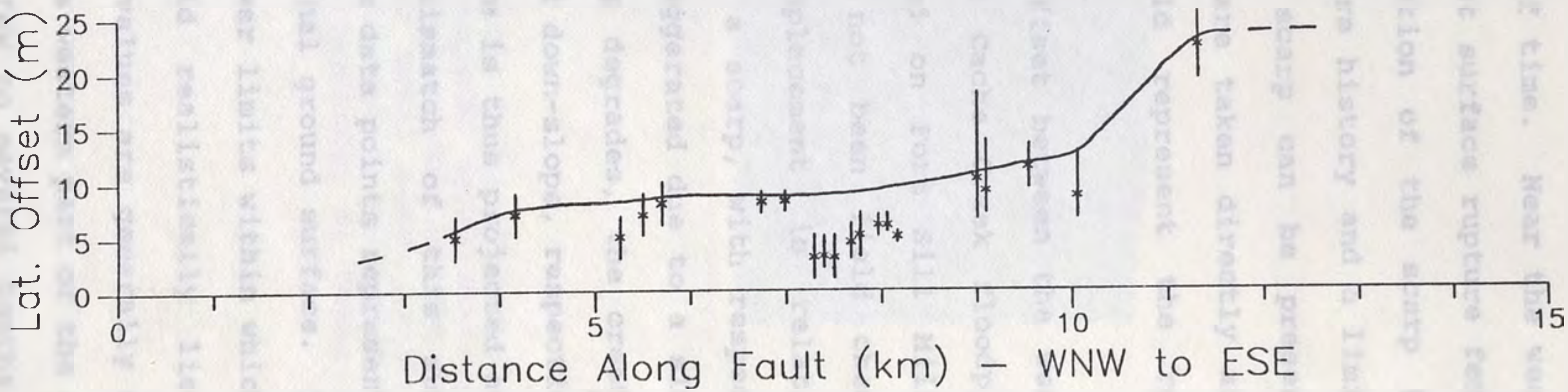


Figure 14: Graph of horizontal displacement along Meers fault scarp from NW to SE. Note scale difference from fig. 13. Limited to Post Oak conglom.

be preserved over long periods of time. Near the western end of the scarp, the most recent surface rupture forms a continuous oversteepened mid-section of the scarp (fig. 15), suggesting a multiple rupture history and a limit to the length of time that the scarp can be preserved. Measurements of lateral offset are taken directly across the scarp and therefore should represent the recent activity.

Figure 14 shows vertical offset between the western end of the fault and the East Cache Creek floodplain. Rupture to the east is located on Fort Sill Military Reservation and has therefore not been field checked. Determination of vertical displacement is relatively straightforward. The height of a scarp, with respect to the horizontal, is typically exaggerated due to a sloping original surface. As the scarp degrades, the crest and base of the scarp migrate up- and down-slope, respectively. The slope of the original surface is thus projected across the scarp and the vertical mismatch of this surface determined (fig. 16). Fifty-five data points represent the vertical separation of the original ground surface. Error bars indicating the upper and lower limits within which the actual displacement value could realistically lie are included. These plus or minus values are generally a few tenths of a meter or less for the western part of the scarp in Post Oak Conglomerate and a few to several tenths of a

Figure 15: Photograph of Meers fault scarp with parallel surface traces forming an oversteepened upper mid-slope, near western end of scarp; displacement across these traces accounts for about 1/3 of total scarp height at this location; traces are accentuated by vegetation growing along fault. Profile T1 (Appendix A) taken near ridge crest on right side of photo.

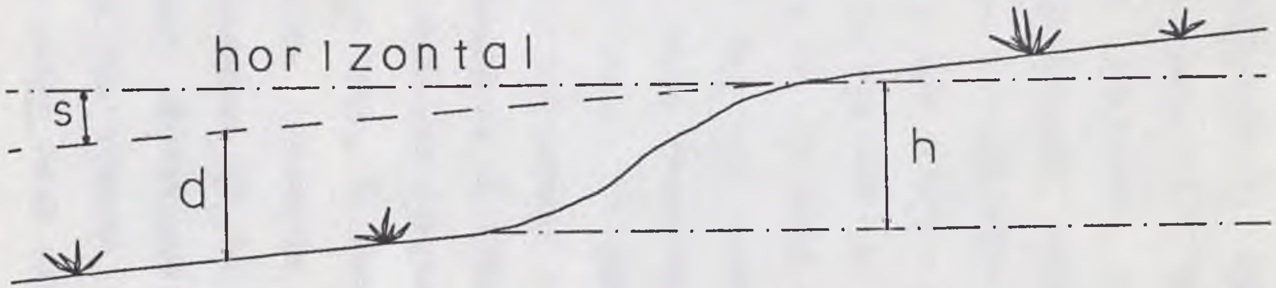


Figure 16: Method for determining vertical displacement; s = original surface slope, d = vertical displacement, h = scarp height. Note that h>d; for backfacing scarps, h<d.

meter to the east in Hennessey Shale.

Determination of lateral displacement is usually much more ambiguous. Even in cases of historical events, it is often difficult or impossible to accurately determine lateral offsets more than a few years after the event, based on surficial expression alone. For example, documented lateral offsets from the 1932 Cedar Mountain earthquake in west-central Nevada (Gianella and Callahan, 1934) are now either erased or very subdued (Molinari, 1984). In the case of the Meers fault, however, part of the scarp is in Post Oak Conglomerate, which is quite resistant and very closely retains the topography present prior to scarp formation. The other unit the fault cuts, Hennessey Shale, is apparently too easily eroded and does not retain indicators of lateral offset for very long, so the measurements of lateral displacement are limited to the scarp in Post Oak Conglomerate.

Ridge lines in Post Oak Conglomerate have been little eroded since formation of the scarp, and thus exhibit the best expression of lateral offsets. Lateral offsets of ridges were determined by flagging the ridge crests and measuring the lateral offset of projections of the crest into the scarp from either side (fig. 17). The locations of ridge crests were determined by sighting along the ridge and, using a Brunton compass as a level, locating a series of horizontal tangents along the crest. Locations of

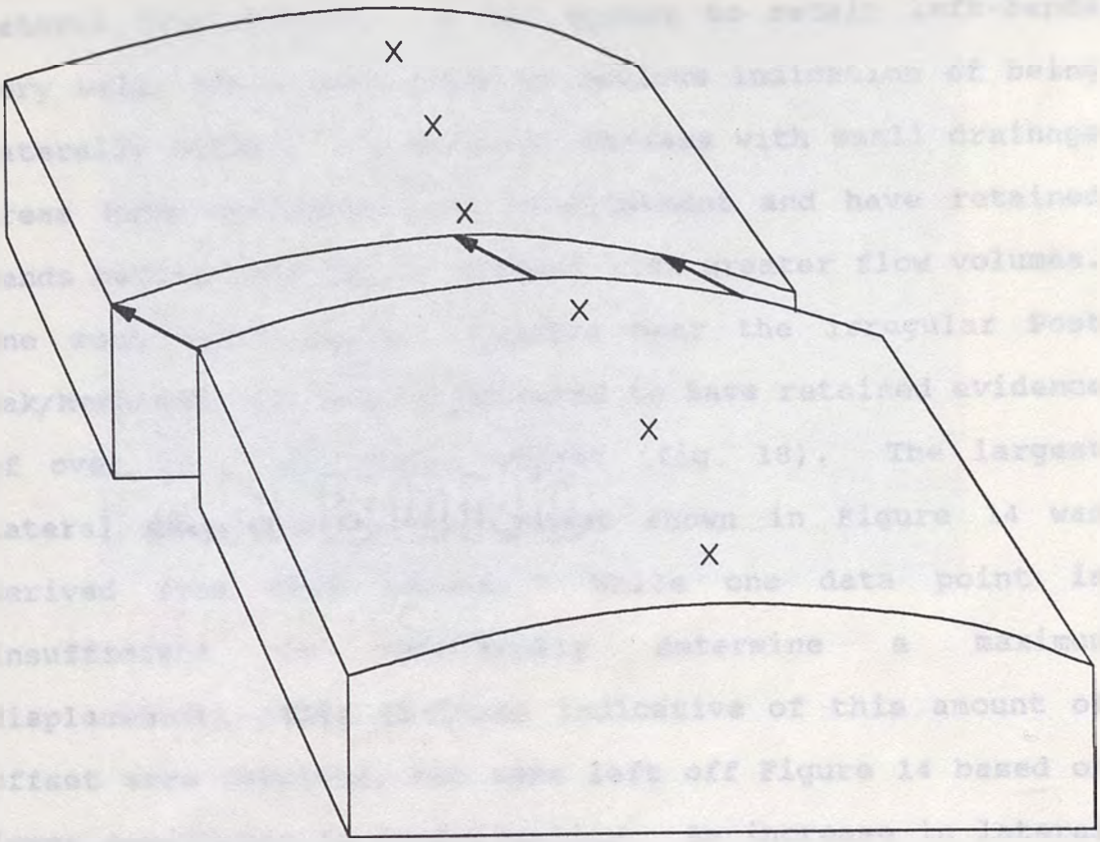


Figure 17: Schematic block diagram illustrating method for determining lateral offset of ridge crests and accentuation of scarp on west sides of ridges. Left-oblique displacement direction indicated by arrows. X's indicate location of ridge crest axis.

crests could usually be determined within a confidence range of plus or minus 1-2 meters. Displacement amounts are fairly consistent from location to location.

Streams exhibit varying degrees of adjustment to the lateral displacement. A few appear to retain left-bends very well, while most show no obvious indication of being laterally offset. In general, streams with small drainage areas have undergone less readjustment and have retained bends better than larger streams with greater flow volumes. One such small stream, located near the irregular Post Oak/Hennessey contact is believed to have retained evidence of over 20 m of lateral offset (fig. 18). The largest lateral displacement measurement shown in Figure 14 was derived from this stream. While one data point is insufficient to confidently determine a maximum displacement, other features indicative of this amount of offset were observed, but were left off Figure 14 based on lower confidence in their quality. An increase in lateral offset at this location, relative to the Post Oak section of the scarp, is reasonably consistent with a similar increase in vertical offset, which is more well constrained, so it is believed that this is a real value.

Due to the normal lack of good markers of lateral offset, confidence levels in lateral displacement measurements range from a few to several meters (fig. 14). This is small enough with respect to the amount of offset

Figure 18: Oblique aerial view (to SSW) of small stream with a sharp left bend at the fault. This is located about at the Post Oak/Hennessey contact and suggests more than 20 m of left-lateral displacement. (Photo by A. R. Ramelli)

for measurements to be fairly well constrained. A check on measuring offset of ridges is afforded by scarp heights being accentuated on the west sides of ridges and subdued on the east sides. By measuring scarp heights on both sides of the ridge and the side slope angles, one can calculate trigonometrically the amount of lateral offset required. Results from this type of analysis agree remarkably well with direct measurements.

A left-lateral component of displacement was recognized by Donovan and others (1983). The displacement measurements made for this study indicate that, on the average, left-lateral displacements exceed up-to-the-north vertical displacements by about 3:1 to 5:1 (figs. 13 and 14). Left-lateral displacement is compatible with an ENE-WSW oriented greatest horizontal principal stress direction in the south-central United States (see fig. 1) as determined principally from hydraulic fracturing and borehole breakouts (Dart, 1987a, 1987b; Zoback, 1987) and discussed in the section on state of stress.

In Figures 13 and 14, an enveloping surface is shown enclosing the data points and is assumed to be an approximation of the cumulative component of displacement across the zone represented by the existing scarp. Comparison of measured surface slip and geodetic surveys for a number of earthquakes (Thatcher and Bonilla, 1988) shows that maximum surface slip generally approaches, but rarely exceeds the geodetic estimates (fig. 19). In Figure 13, points that lie beneath the enveloping curve may represent areas where either an older scarp was completely or partially eroded (as discussed later in this section) or where additional deformation occurred by slip on Riedel and other shears, warping, or other forms of strain.

By comparing the enveloping curves in Figures 13 and 14, a ratio of the lateral to vertical components of surface separation can be estimated. This suggests for the Poat Oak Conglomerate section of the fault, the lateral component exceeds the vertical by a ratio of about 4:1 with values that range from about 3:1 to 5:1.

There is a decrease in vertical offset from east to west at about the 9 km point (fig. 13). This coincides with a slight bend (less than 5^{o}), with the fault trending slightly more northerly to the 7 km point where it straightens to its original orientation (fig. 20). From the east, the lateral displacement appears to decrease rapidly approaching this bend, be suppressed through this area, and increase slightly to the west. With left-lateral displacement, this would be a restraining bend. A decrease in net slip across this feature should be reflected in both components of displacement, so the decrease in the vertical component is not surprising. Figures 13 and 14 suggest that this had a somewhat more pronounced effect on the lateral component.

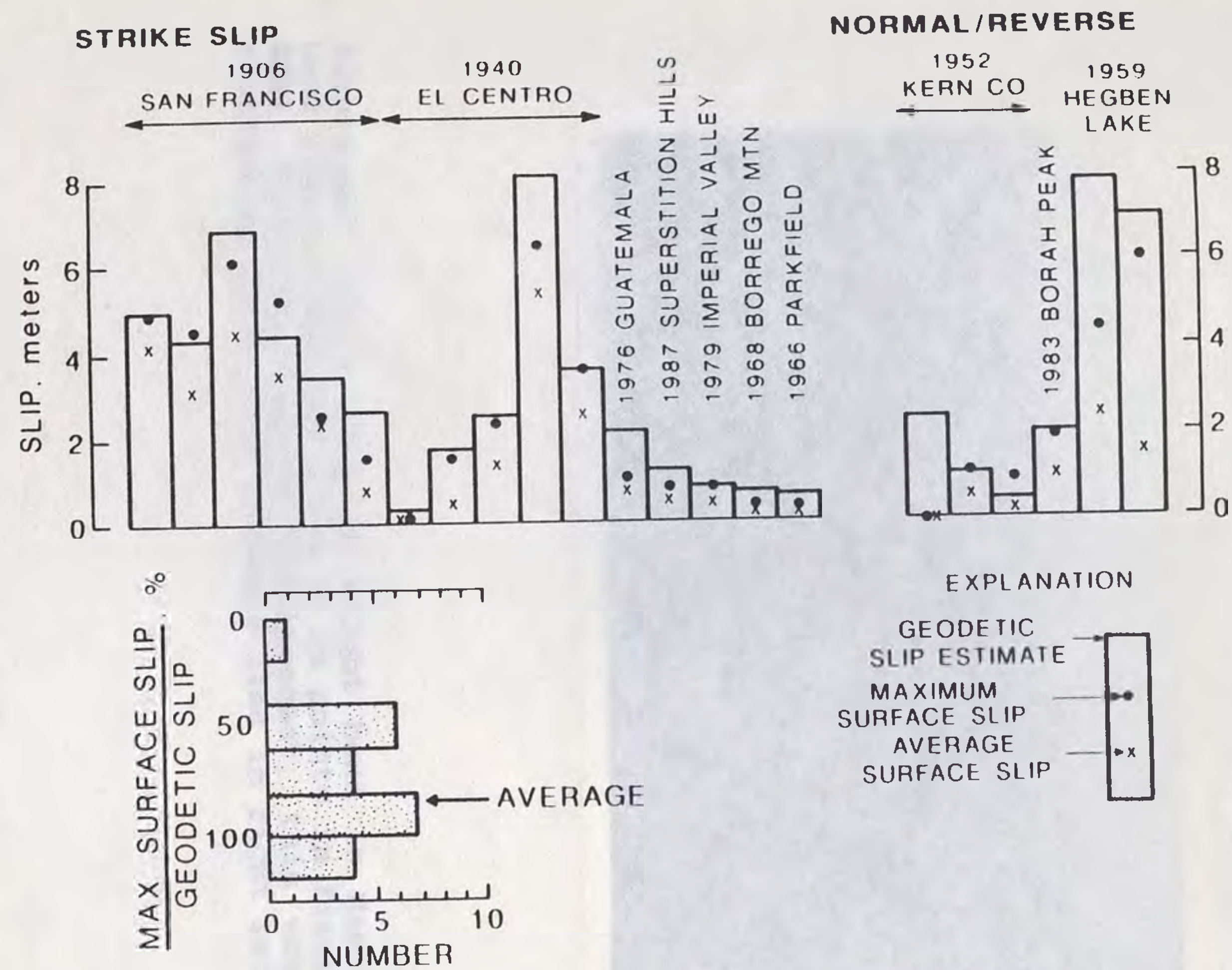


Figure 19: Comparison of measured surface slip and geodetic estimates for a number of historical surface faulting events. Histogram at bottom shows that maximum surface slip reasonably approximates geodetic estimates (Thatcher and Bonilla, 1988).

Figure 20: Aerial view of slight bend in Meers fault, bend lies between the 7 and 10 km points in Figure 13 (to ESE). With left-lateral displacement, this would be a restraining bend. Lake Lawtonka lies in right background.

A number of features can be observed that clearly show the fault has a definite strike-slip history, although to what extent the various evidence is a result of the recent activity or of past activity is not well known. Features probably resulting from past activity include the linear fault trace and a conjugate joint pattern dominated by the fault trace and consistent with east-west compression. Other features that could result from either past or recent activity include numerous Reidel shears that, even if preexisting, experienced slip during recent event(s) and tensional (dilational) features near the bend that was discussed previously. Although this is a restraining bend, tension would be expected on either side, as demonstrated in Figure 21. Horizontal slickensides are present along

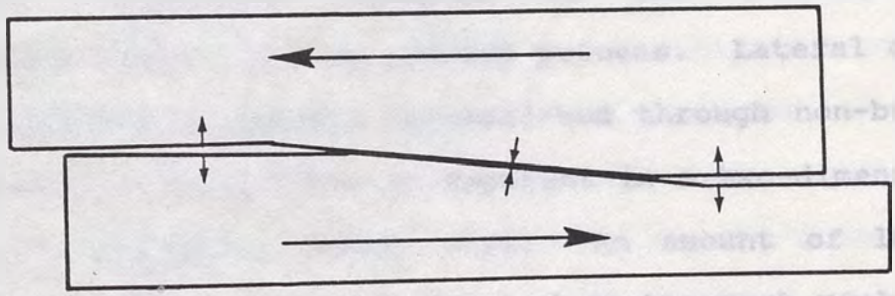


Figure 21: Schematic model predicting dilation along a strike-slip fault on either side of a restraining bend.

the fault in Post Oak Conglomerate. These may not have formed during the most recent event, but Crone and Luza (1986b) feel that it is unlikely that the fine striae would be preserved in soluble carbonate materials for hundreds of thousands of years and thus were probably formed during the late Quaternary.

Two separate studies based on information derived from areas in Hennessey Shale have argued against a dominant lateral component of displacement, at least for the most recent event. The first is based on exploratory trenching just east of the Post Oak/Hennessey contact (Luza and others, 1987). One of the trench logs shows fairly good continuity of thin units across a zone of brittle failure, arguing against significant lateral offset along this plane. However, at this location, brittle failure accounts for only a small part (about 30 cm or 10%) of the total vertical deformation observed in the trench, with monoclinal flexure as the primary process. Lateral offset could likewise be largely accomplished through non-brittle deformation and would not be apparent in a two-dimensional plane afforded by a trench wall. An amount of lateral offset consistent with that observed to the west would call for only one meter of brittle strike-slip, assuming that the brittle failure was accomplished equally for the vertical and lateral components. The other trench showed more brittle failure and less continuity across the fault plane, but in less distinct stratigraphy. The trench work was done primarily by Tony Crone, who has since reinterpreted displacement to be strike-slip, based on trench exposures to the west in Post Oak Conglomerate (Crone and Luza, 1986b). In another study conducted further to the east (9 1/4 km), samples of Hennessey Shale were recovered from the fault zone. These are reported to have near vertical striations (Tilford and Westen, 1985b). Further work of this type is warranted to determine if these are in fact fault striae, or striae produced by expansive clays.

The number of events accounting for the cumulative displacement is not well known, but several factors indicate multiple events, including bevels along the crest of the scarp, multiple flights of stream terraces on the upthrown side of the fault, and heights of scarps formed in stream terraces. Along much of the fault, in particular where faulting was accommodated by warping of surficial materials, the existence of bevels is problematic, but in some places it is quite clear that the most recent event formed an oversteepened mid-section of an older, more degraded scarp (e.g. fig. 15). The terrace scarps and sharp knickpoints in stream channels cut into Post Oak Conglomerate indicate that the most recent event accounts for roughly one-half the total scarp height. Whether this indicates that the scarp was primarily formed by two large events, or whether earlier events were smaller or had lesser vertical displacements, remains to be determined. More than a few events would require greatly varying amounts and/or senses of displacement. This seems unlikely, given the uniform tectonic stress believed responsible for driving fault displacement. Since the most recent event accounts for as much as one-half the total scarp height, it is believed that surface displacement reached several meters (i.e. in the range of 5 to 10 meters).

Evidence for Coseismic Scarp Formation.

Tilford (1987) has suggested that it is premature to designate the Meers fault as a possible source of large magnitude earthquakes. However, given our current understanding of faulting processes and observations of the Meers fault scarp, we believe that such events are required.

The principal aspect that indicates that a fault is "creeping," is slow, but persistent surface displacements. This results in cracking and misalignments of paved surfaces, buildings, gutters, underground lines, etc. No such damage was observed during field studies, nor is known to have been reported, along the trace of the Meers fault scarp.

Additionally, the presence of a continuous scarp with steep slope angles requires rapid scarp formation. Slow displacement of the surface would allow erosional processes to degrade surface offsets before a sharp topographic break could be created. Near the western end of the scarp, a small (about 1/4 m), continuous break in the mid-section of the scarp has a slope angles of 25^{o}-30^{o} (fig. 15). This could not be preserved if it were formed by slow, continual displacement. Even if conditions were such that a scarp could be formed (i.e. high slip rates and low erosion rates), it would be highly uneven, due to local variations in erosion; this is not the case.

The lack of microseismicity further suggests that the Meers fault is currently "locked", and that stress accumulation is released during sudden, stick-slip seismic events. Fault sections observed to be creeping are typically characterized by high levels of microseismicity. The creeping section of the San Andreas fault, approximately from Corralitos to Cholame, and parts of the Hayward and Calaveras faults, are clearly delineated by abundant microseismicity (fig. 22). The non-creeping sections of the San Andreas fault, which ruptured during the 1857 and 1906 earthquakes (magnitudes about 8), show sparse seismic activity. Fault creep is presumed to represent fault zones that lack strong enough

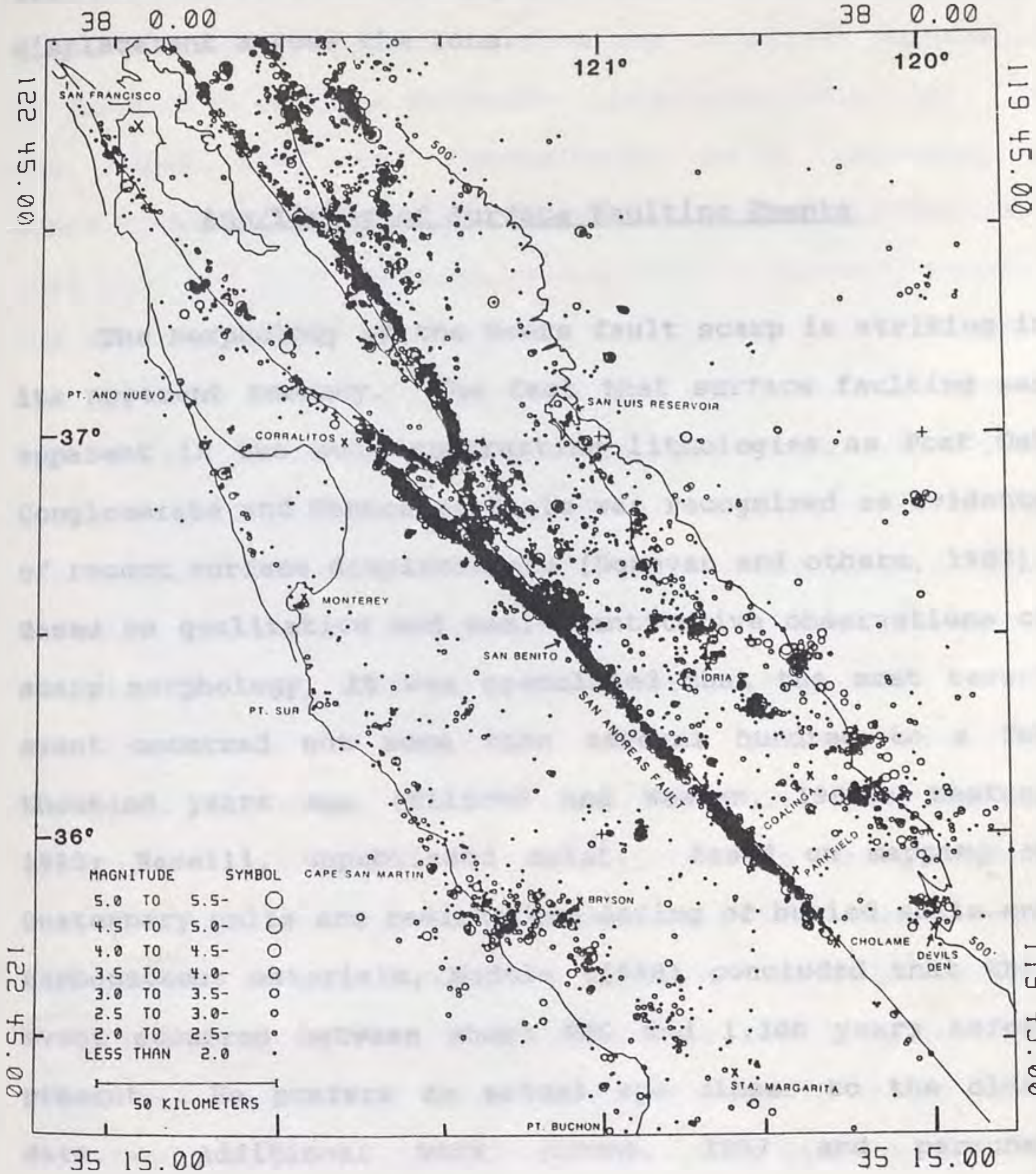


Figure 22: Seismicity map of central California Coast Ranges area, illustrating high level of microseismicity along creeping sections of San Andreas, Calaveras, and Hayward faults. $M_L \geq 1.5$ for January, 1972 through April, 1983 (Dehlinger and Bolt, 1987).

discontinuities (i.e. asperities) to lock the differential displacement across the zone.

Age/Timing of Surface Faulting Events

The morphology of the Meers fault scarp is striking in its apparent recency. The fact that surface faulting was apparent in two such contrasting lithologies as Post Oak Conglomerate and Hennessey Shale was recognized as evidence of recent surface displacements (Donovan and others, 1983). Based on qualitative and semi-quantitative observations of scarp morphology, it was speculated that the most recent event occurred not more than several hundred to a few thousand years ago (Tilford and Westen, 1985a; Westen, 1985; Ramelli, unpublished data). Based on mapping of Quaternary units and radiocarbon dating of buried soils and carbonaceous materials, Madole (1988) concluded that this event occurred between about 600 and 1,300 years before present. He prefers an actual age closer to the older date. Additional work (Crone, 1987 and personal communication) supports these dates.

One sample was collected and submitted for radiocarbon dating for this study. This sample consisted of fragmented, reworked shell material in a thin (about 5 cm) near-surface sandy alluvial deposit. This deposit is located on the upthrown side of the fault, about 30 cm below a stranded terrace surface and is offset vertically by about 3 cm along a secondary fault trace (fig. 23). It was hoped that the age-estimate would represent a depositional event closely preceding the stream incision that resulted from the most recent event. However, because the shell material was reworked, it yielded only a maximum age for this event. The age-estimate obtained, for which a ^{13}C correction was done, is 13,330$\pm$190 years B.P. This material must have been reworked out of the basal part of the Browns Creek Alluvium (Madole, 1988), for which an age-estimate of 13,670$\pm$120 years B.P. was obtained from shell material (Madole, 1988; Luza and others, 1987).

Comparison of vertical offset measurements and locations of major streamlines (fig. 13) suggests the following correlation. Scarp heights near these streams are, in general, roughly 1/2 that of the total scarp height represented by the enveloping surface. This holds true more in Hennessey Shale than in Post Oak Conglomerate, which was apparently less affected due to being more resistant. Scarp heights formed in stream terraces in Post Oak Conglomerate are also generally about one-half of the total scarp height at these locations. At about 13 km from the western end of the scarp, just west of U. S. Highway 58, the Oklahoma Geological Survey dug two exploratory trenches. Stratigraphic units, which range in age from

Figure 23: Young deposit showing small offset along secondary fault trace. Snail shell material from this deposit was submitted for radiocarbon dating. The age-estimate obtained indicates this material was reworked from basal part of "Brown's Creek Alluvium" (Madole, 1988). Diane Westen for scale. (Photo by A. R. Ramelli)

about 1,500 to 13,000 years B.P. (Madole, 1988), revealed about 2 1/2 to 3 m of vertical displacement (fig. 24). Although complicated by the large amounts of warping, these studies have argued for only a single event in the last 13,000 to 14,000 years, that being the most recent one (Luza and others, 1987). This suggests a possible recurrence interval of greater than 10^4 years, but timing of events and interseismic intervals need to be better constrained.

The evidence for the most recent event accounting for about one-half of the total scarp height, along with locations where the most recent event can be observed to form an oversteepened midsection on an older, more degraded scarp, indicates there have been multiple late Quaternary events. However, numbers and ages of any large events prior to the most recent one are not well documented.

Slip Rates

Accurate determinations of slip-rates could provide insight into the cause of the recent faulting, probability of renewed activity in the near future, and probability of the presence of active faults in other intraplate areas. Not enough data on displacement amounts and recurrence

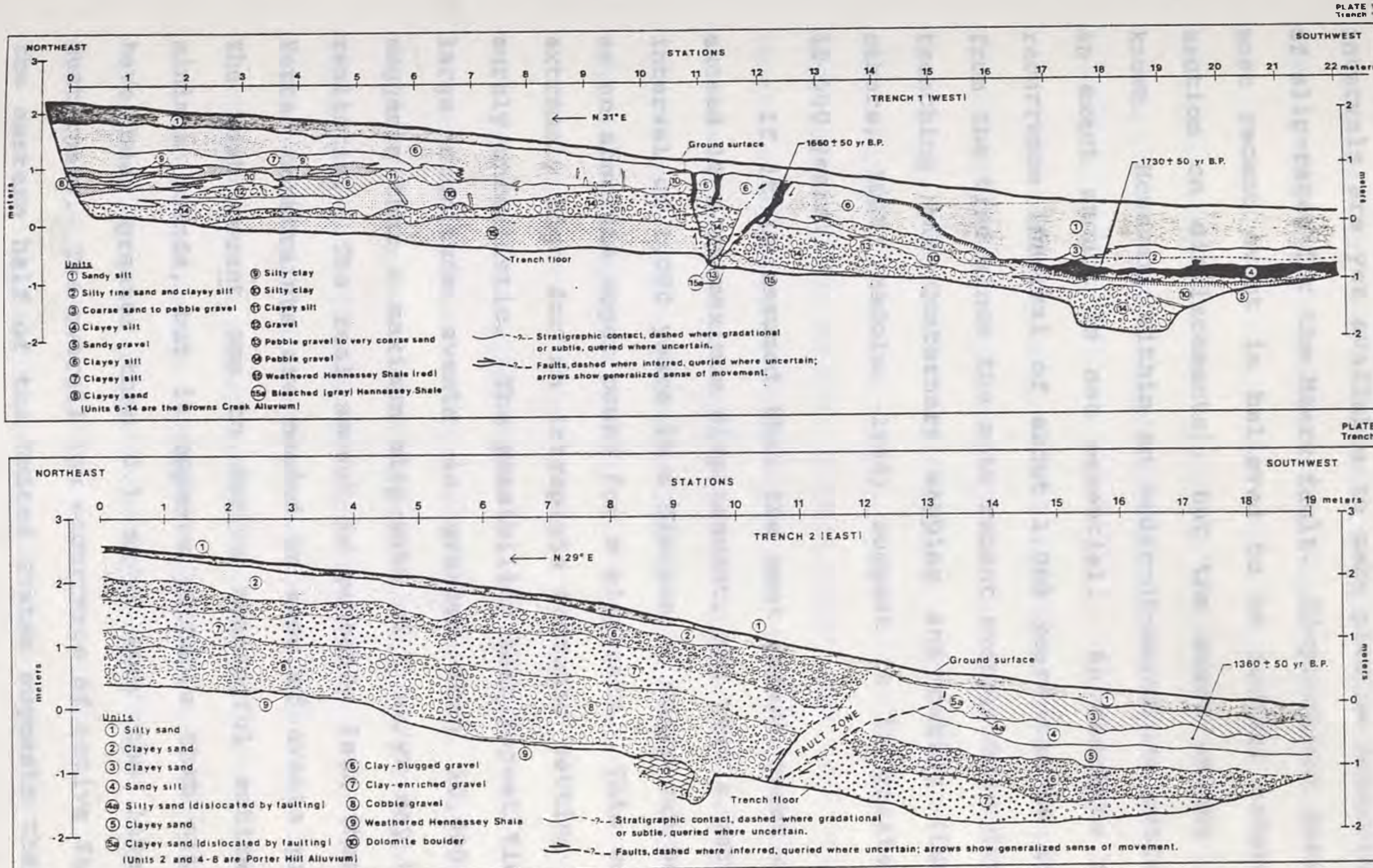


Figure 24: Logs of trenches across Meers fault scarp (Luza and others, 1987). Trenches located just west of Highway 58. Both logs of east walls. Work by Oklahoma and U.S. Geol. Surveys; logging by A. J. Crone (USGS).

intervals are yet available to make precise determinations of slip-rates for the Meers fault. Displacement during the most recent event is believed to be several meters (see section on displacements), but the exact amount is not known. However, within an order-of-magnitude estimation, an exact amount is not essential. An absolute minimum recurrence interval of about 1,000 years can be inferred from the time since the most recent event, while data from trenching and Quaternary mapping and dating (Luza and others, 1987; Madole, 1988) suggest it is greater than 10,000 years.

If it is assumed that the most recent event did not exceed 10 m in maximum displacement, and that a recurrence interval of 1,000 years is a minimum, 1 cm/yr can be taken as an absolute upper bound for a slip-rate. This would be extremely high for an intraplate tectonic setting and is surely unrealistic. The possibility that repeat times for large magnitude events is greater than 10,000 years suggests that a maximum slip-rate of 1 mm/yr may be more realistic. The real amount is probably less than this. Better constraints are needed on ages of events prior to the most recent one to derive meaningful estimates of minimum bounds, but it appears probable that slip-rates have been greater than 0.1 mm/yr over the latest-most Quaternary. The overall low occurrence of active faults in the eastern half of the United States suggests that rates as high as this are unlikely to persist for long periods of time. Long-term rates are surely much lower.

Scarp Morphology

Over recent years, a number of workers have attempted to quantify the morphology of fault scarps, with the hope that such applications could yield estimations of ages of prehistorical seismic events. Early work was aimed at describing fault scarps, those processes responsible for their degradation, and deposits resulting from such processes (e.g. Wallace, 1977). While such studies are invaluable for understanding the nature of scarp degradation, they yield only crude age estimations.

Bucknam and Anderson (1979) developed a technique of relative age dating for fault scarps of differing ages over an area or region relatively consistent in materials and climate. For three sets of fault scarps with order-of-magnitude age differences in south-central Utah, they found very good correlation between maximum scarp slope angle and the logarithm of scarp height. This technique provided more quantitative age estimations than previously available ones, but only in a relative sense and still only to an order-of-magnitude.

A more sophisticated approach is currently available through the use of diffusion mathematics. This technique has recently been applied in numerous studies (e.g. Nash, 1980, Hanks and others, 1984; Pierce and Colman, 1986). It has the advantage over previous techniques in providing absolute age determinations, with a claimed accuracy of about plus or minus 50% (Hanks and Wallace, 1985).

The vast majority of scarp morphology studies have been carried out in the Basin and Range Province, where abundant normal faults and a dry climate provide numerous preserved fault scarps, and large prehistoric inland lakes, primarily Lahontan and Bonneville, provide shoreline scarps of known age for cross-cutting and comparative relations. Correlation of these studies to other regions, such as the south-central United States, require consideration of differing factors, such as materials, climate, and vegetation.

The surface expression of the Meers fault scarp is indeed remarkable, making it an attractive target for scarp morphology studies. It would be advantageous if detailed scarp morphology techniques, such as diffusion modelling, could be applied. There is no reason such techniques can not be of use in stable continental interior (SCI) regions, as long as applicable degradation rates can be determined. However, in the case of the Meers fault scarp, the materials factor presents an imposing stumbling block.

Where the fault cuts Post Oak Conglomerate, which lies at or very near an erosional lag surface, it forms a bedrock scarp. An unconsolidated cover, primarily composed of weathered bedrock, and probably containing some wind-blown material, is generally a few tenths of a meter thick. Along much of this part of the fault, bedrock free-face makes up a significant part of the scarp. Slopewash and eolian input are likely to cause more material to be deposited at the base of the scarp than is eroded from the crest, violating the assumption of conservation of material.

On the other hand, where the fault cuts Hennessey Shale, much of the near surface deformation is commonly taken up by warping, as observed in trenches and natural exposures. For example, in trench 1 (Luza and others, 1987), warping accounts for as much as 90% of the vertical surficial deformation (fig. 24). The apparent scarp age would be highly dependent on the amount of warping. Areas of predominantly brittle offset do occur, as evidenced by the Brown's ranch streamcut exposure, but the consistency of expression may suggest this is not common. This can only be determined through subsurface investigations (e.g. trenching and/or geophysics).

Scarps in stream terraces may be the best for scarp morphology analyses, although they also present a couple of problems. They typically have a dense vegetative cover, which makes their accessibility very poor, and they are subjected to periodic flooding, likely violating the assumption of conservation of material. Some data was collected on scarp heights and slope angles for these scarps, but no profiling was done. This would be a worthwhile endeavor for future studies.

These problems, along with the poorly understood fault history, limit the application of such methods at the present time. Although the apparent problems in applying scarp morphology analyses to the Meers fault scarp raise many more questions than can be answered here, the data collected is presented in order that it be accesible (Appendix A). Any similarities with data from other study areas should be carefully scrutinized before they are taken to indicate the validity of correlations.

<u>Scarp Height vs. Slope Angle Analysis</u>

The Meers fault data on scarp height versus maximum scarp slope angle are broken into three groups, based on material type: Post Oak Conglomerate, Hennessey Shale, and stream terrace deposits. To provide a frame of reference, these data are compared to that of Bucknam and Anderson (1979). The age of the most recent event on the Meers fault (i.e. about 1,000 years B.P.) is roughly comparable to that on the Fish Springs fault. The Post Oak and terrace data agree fairly well with the Fish Springs data, although there is much more scatter. The Hennessey data set is significantly different, approximating the Panguitch set. The Fish Springs (about 10^3 years) and Panguitch (about 10^5 years) fault scarps vary in age by two orders of magnitude. The fact that the morphology of the Meers fault varys by this amount for scarps of similar ages illustrates some of the problems in haphazardly applying rigorous scarp morphology analyses. For two principal reasons, it is felt that extreme caution should be used in evaluating Hennessey Shale scarp morphology. Primarily is the aforementioned non-brittle warping of the surface. Secondly, much of the area in which this part of the fault scarp lies has been extensively cultivated.

Two differences are apparent between the Meers and Bucknam and Anderson (1979) data sets. Perhaps the most significant of these is that all three of the Meers data sets have slightly better correlation when scarp height, rather than the logarithm of scarp height, is used. The second is that the linear regression equations have lower slopes. The reasons for these differences are problematical and require a more thorough treatment than can be afforded here.

Diffusion Analysis

The diffusion equation has been utilized in quantification of landform evolution for some time (e.g. Culling, 1960). The first known application of this method to fault scarps was done by Hirano (1972), working with fault scarps of the Hira Mountains in central Japan. Nash (1980) introduced the method to this country and initiated its current widespread use. Several reseachers are now working to refine this method and establish the validity of its inherent assumptions and its problems (Nash, 1984; Hanks and others, 1984; Mayer, 1984; Andrews and Hanks, 1985; Pierce and Colman, 1986; Andrews and Bucknam, 1987).

The principal assumptions of this method of analysis are that material is moved downslope at a rate proportional to the slope angle and that mass is locally conserved. For a topographic profile taken across strike of a fault scarp, this situation can be expressed as:

$$\frac{dy}{dt} = k\frac{d^2y}{dx^2} \qquad (1)$$

where: y = relative elevation within the profile

x = horizontal position within the profile

t = time

k = a constant incorporating all factors, except time, affecting scarp degradation (climate, materials, etc.)

For a step introduced into an originally planar sloping surface (e.g. a fault scarp in an alluvial fan surface), this formula has the solution:

$$y(x,t) = a\ \mathrm{erf}(x/2(kt)^{-2}) + bx \qquad (2)$$

where: a = one-half the offset

b = tangent of the surface slope

erf = the error function.

This equation can thus be solved for the value of k times t. If a value for k is known (from fault scarps of known age in similar materials), then it can be solved for t. Likewise, if t is known (from dated scarps created by the same event in differing materials), then it can be solved for k.

In diffusion modeling, "time" does not begin at the time of faulting, but rather at the time the scarp has degraded to a debris slope, presumably at the angle of repose of the faulted material. This is assumed to occur over a short time period (tens of years) relative to the age of the scarp. The appearance of historical fault scarps suggests this is usually a valid assumption.

For the following reasons, it seems logical that the behavior of scarps in Post Oak Conglomerate would violate the assumptions inherent in degradation by diffusivity: 1) it can not be assumed that the scarp degrades to the angle of repose in a short time. A bedrock free-face is commonly still preserved, and locally accounts for nearly the entire scarp height; 2) the resistant nature of material at the scarp crest should greatly decrease the rate of degradation of the upper part of the scarp, possibly allowing a greater amount of material accumulation in the lower part of the profile through slope wash and eolian influx; and 3) the surface is erosional, indicating possible removal of material from parts of the profile. Although some material may be removed from scarp profiles, this is unlikely to be significant when profiles are carefully located.

Warping of the surface over Hennessey Shale violates the assumption of an angle-of-repose starting point. This may not be a damning problem, however. If the initial scarp profile can be inferred from the configuration of subsurface units, then the diffusion "clock" can be started at this point. This would require close coordination between profiling and subsurface investigations (i.e. trenching).

STATE OF STRESS IN THE CRUST

Although the overall level of seismicity in the central and eastern United States (CEUS) is generally low, large magnitude historical earthquakes, such as the 1811-1812 New Madrid, Missouri series of three great earthquakes (magnitudes about 8) or the 1886 Charleston, South Carolina earthquake (magnitude about 7 1/2), and paleoearthquakes such as those on the Meers fault, serve to remind us that "mid-plate" areas are subjected to crustal stresses of great enough magnitude to drive fault displacement. Given the right combinations of fault orientation, fault strength, time, stress level, and stress orientation, large earthquakes can and do occur in stable continental interior (SCI) regions. Due to lower attenuation of seismic energy, these can be even more destructive than those at plate boundaries. Recent work by Musman and Schmidt (1986) has shown a convergence rate of 1.2 cm/yr between Westford, Massachusetts and Fort Davis, Texas. This is a higher rate than is usually assumed for SCI regions, but the presence of features such as the Meers fault scarp suggests that this is may be real.

The causes of stress in the lithosphere can not be directly observed, so while a number of sources have been postulated, the interplay between various sources is poorly understood. Postulated sources include, but are not limited to, ridge-push from oceanic spreading centers, slab-pull from subducting oceanic lithosphere, crust/mantle interactions, crustal contrasts (density, thermal, and lithological), volcanism, basinal loading, and change in curvature from plate drift to differing latitudes (i.e. membrane tectonics).

A number of methods have been developed or invoked for indirect determinations of state-of-stress of the crust, most of these aimed at the contemporary (or present day) state-of-stress. Several of these principally rely on observations of bore-hole behavior, including hydraulic fracturing, overcoring, strain gauges, and bore-hole breakout. Another major group of methods aimed at the contemporary state-of-stress utilize observations of displacements during earthquakes, including focal mechanisms, measurements of surface displacements, and geodetic surveys. Methods of determining paleostress are also available in analysis of geologic features such as fracture patterns (e.g. Reches, 1987), dikes, and folds.

As methods of measuring stress in the crust have been refined and stress data amassed, it has become increasingly obvious that almost the entire CEUS is subjected to a nearly uniform, horizontal compressive stress oriented ENE-WSW to NE-SW (Sbar and Sykes, 1973; Zoback and Zoback, 1980; 1985; Zoback, 1987) and that the region around the Meers fault fits this pattern very well (Dart, 1987a;

1987b). Deformation of Post Oak Conglomerate is limited to relatively simple jointing (fig. 25) and minor pressure solution features. Joint density progressively decreases away from the fault. Stereoplots of joint measurements (figs. 26 and 27) show a conjugate pattern consistent with E-W compression and dominated by the main fault trace. This suggests that the stress conditions responsible for the recent faulting have been present for some time.

Fault-stress relationship models developed by various workers constrain the relative displacement that can be expected based on fault plane orientation (e.g. Harding, 1974). The Meers fault and the Amarillo-Wichita uplift have WNW trends. Given this orientation and the existing ENE-WSW compression, left-lateral displacement would be expected. This situation is represented by laterally displaced geomorphic features along the Meers fault scarp.

The large component of vertical, up-to-the-north displacement suggests that, at least locally, the principal stress axes may be somewhat inclined. A case of pure strike-slip displacement would suggest that the intermediate stress axis is oriented vertically, whereas pure reverse displacement would suggest it is horizontal. With the Anadarko Basin being the deepest sedimentary basin on the North American continent and having a sizable gravity gradient across the Wichita frontal fault system, it seems plausible, and has been postulated (Tilford and

Figure 25: Photograph of joint sets creating vegetation lineaments in Post Oak Conglomerate; at this location joints lie oblique to main fault trace, which crosses the foreground. Joint measurements plotted in Figures 26 and 27 were taken principally near the main fault trace and are dominated by fractures paralleling the fault. (Photo by D. B. Slemmons)

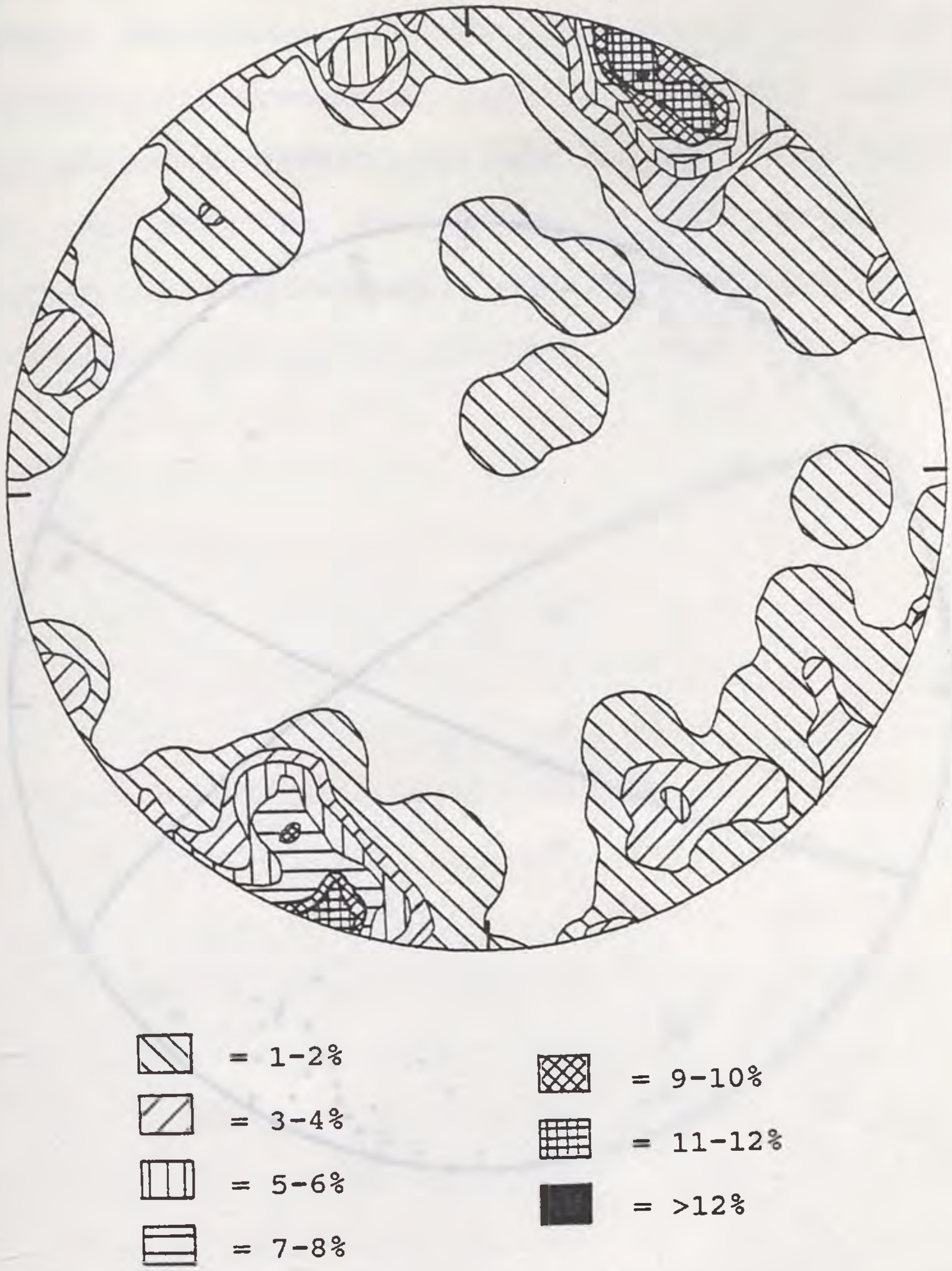


Figure 26: Contoured stereoplot of poles to joint planes data (Figure 27) taken from Post Oak Conglomerate adjacent to Meers fault scarp. Contouring done by floating circle counting method as described by Hoek and Bray (1981).

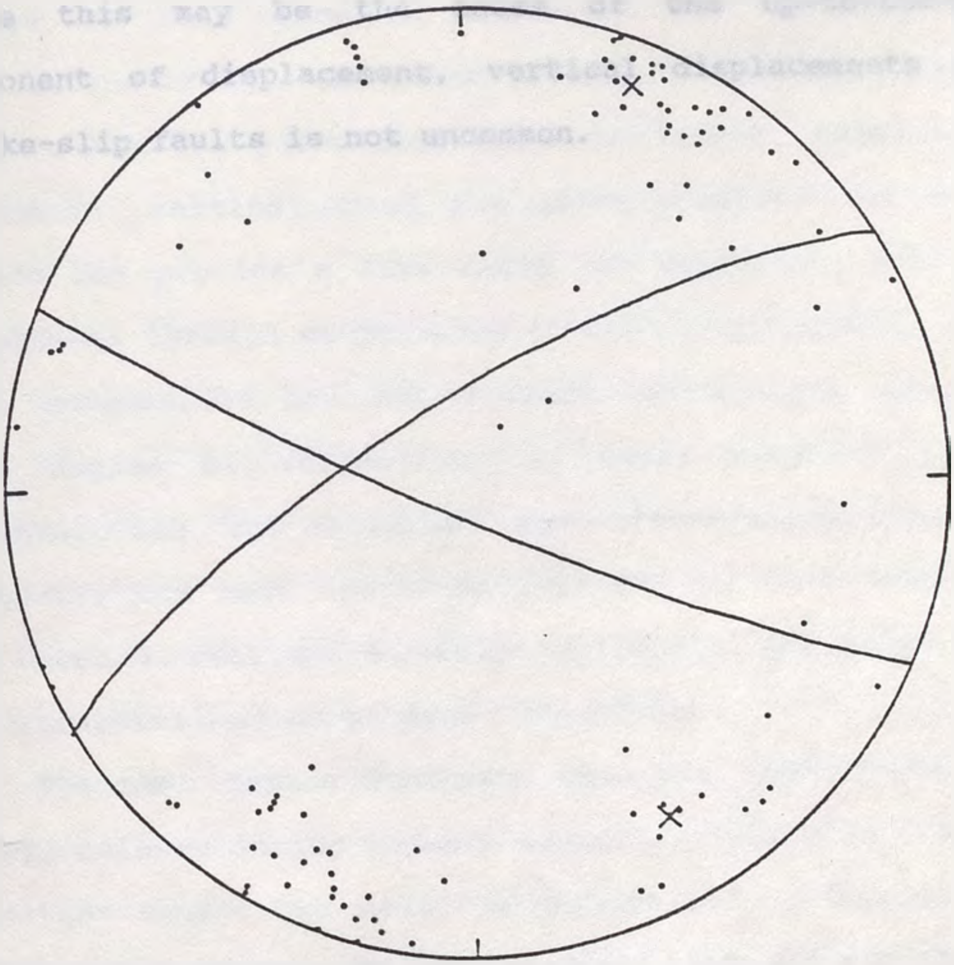


Figure 27: Orientations of poles to measured joint surfaces and mean orientations of joint surfaces as inferred from Figure 26. Two planes shown are oriented N67W, 84SW and N59E, 76NW.

Westen, 1985b), that vertical stresses may be caused by isostatic imbalances. However, the absence of significant topographic differences in this area indicate that large-scale isostatic adjustments are probably not occurring. While this may be the cause of the up-to-the-north component of displacement, vertical displacements along strike-slip faults is not uncommon.

SEISMIC POTENTIAL

Estimation of the maximum size that can reasonably be expected for future earthquakes on the Meers fault or other faults of this zone is important for seismic hazard evaluations of this and other tectonically similar regions. Surface rupture parameters (rupture length, displacement, structural setting, etc.) for large prehistorical seismic events can provide a firm basis for estimating earthquake magnitudes through comparisons to historical events. While such estimations are not without limitations, given the high degree of variability in fault behavior and the uncertainties in obtaining such information, they are generally the best estimates that can be made, since they are based on real and directly applicable data, rather than on extrapolations or general comparisons.

The most common indicator used for quantification of energy release during seismic events is magnitude. Various magnitude scales are based on measurement of seismic waves of varying periods. No scale can be used for all sizes of events. Magnitude scales commonly used include Richter (M_L or local magnitude), surface wave (M_s), and body-wave (m_b) magnitude. A good discussion on magnitude scales is given by Kanamori (1983).

Several studies have related surface rupture parameters to earthquake magnitude for large historical earthquakes (e.g. Slemmons, 1982; Bonilla and others, 1984). Magnitudes of prehistorical events can thus be estimated if rupture lengths and displacement amounts can be accurately determined. The prominent surface expression of the Meers fault scarp gives hope that this type of analysis can be applied in stable continental interior (SCI) regions. This has rarely been the case, since recognized evidence of surface faulting is generally not available. The possibility that paleoseismic studies can be applied in some SCI settings calls for an examination of the uncertainties involved in such applications.

Data used in studies relating magnitude to rupture length and/or displacement are typically grouped by fault type (e.g. strike-slip versus reverse) and region (e.g. interplate versus intraplate). Comparisons by fault type are relatively straightforward. However, there have not been enough large historical SCI earthquakes to allow regional comparisons to be made with much confidence. If it is assumed that no significant differences exist, regression analyses (Slemmons, 1982; Bonilla and others, 1984) give magnitude estimates for the Meers fault of between about 6 3/4 and 7 1/2, based on rupture length and maximum displacement, respectively. The wide range in these values (about 3/4 of a magnitude) result from the unusually large displacements with respect to surface rupture length. The exact amount of displacement per event is not well known, but it is at least a few, and probably several, meters. This is well above the average for historical events (fig. 28).

Historical surface rupture data are too sparse to evaluate with a high degree of confidence whether this is typical for SCI regions. The only events for which sufficient data are available are from Australian earthquakes (Vogfjord and Langston, 1987; McCue and others, 1987; Denham, 1988). These also have somewhat greater than average displacements, relative to fault length, but not by as much as is apparent for the Meers fault. The values plotted for these events (fig. 28) are taken from Denham (1988) and represent vertical throw, rather than net displacement, which would be somewhat greater. These are all dip-slip events, however, so any differences should be fairly small. Aspects about a couple of these events (1968 Meckering and 1986 Marryat Creek) are worth noting. Both had arcuate surface traces, possibly indicating localized faulting in response to high levels of horizontal compressive stress, rather than rupture along a throughgoing fault system. The Meckering earthquake is quite unusual in that it is well constrained at a very shallow depth (about 3 km) (Vogfjord and Langston, 1987). It is likely that this presents a better representation of source parameters, but this may be misleading, since it is being compared to surface manifestations of deeper events.

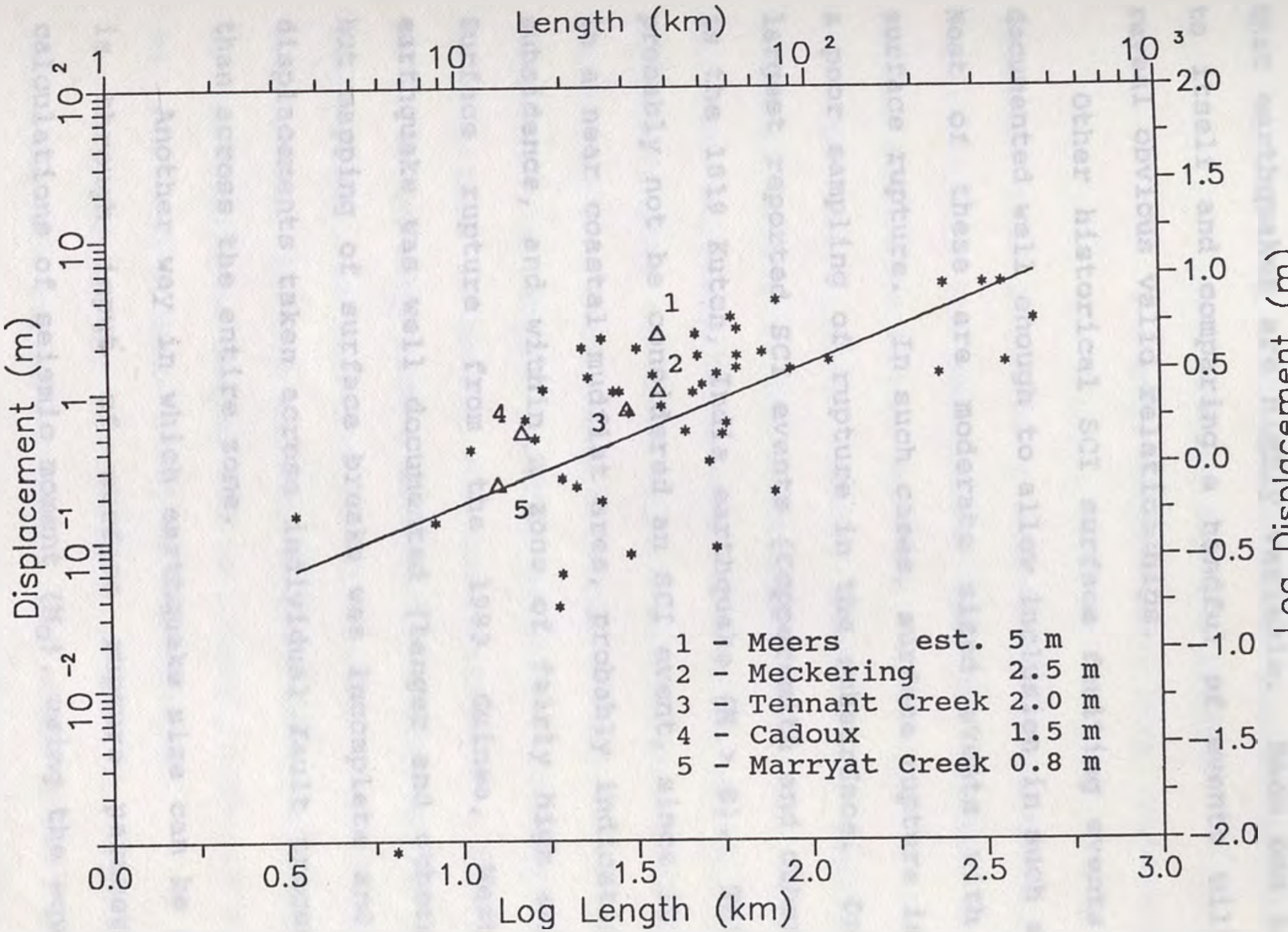


Figure 28: Comparison of stable continental interior (SCI) earthquakes (triangles) and surface rupture length versus maximum surface displacement for historical earthquakes (after Bonilla and others, 1984). Numbers used for Australian events (Denham, 1988) are vertical throw, rather than net displacement, which would be slightly greater.

This is not to say that these events are any less typical of SCI earthquakes than the recent Meers fault events, but that earthquakes are highly variable. Each one is unique to itself and comparing a handful of events will rarely reveal obvious valid relationships.

Other historical SCI surface faulting events are not documented well enough to allow inclusion in such analyses. Most of these are moderate sized events with limited surface rupture. In such cases, surface rupture is usually a poor sampling of rupture in the subsurface. One of the largest reported SCI events (Coppersmith and others, 1987) is the 1819 Kutch, India earthquake ($M > 8$). This should probably not be considered an SCI event, since it occurred in a near coastal mudflat area, probably indicating active subsidence, and within a zone of fairly high seismicity. Surface rupture from the 1983 Guinea, West Africa earthquake was well documented (Langer and others, 1987), but mapping of surface breaks was incomplete and measured displacements taken across individual fault traces, rather than across the entire zone.

Another way in which earthquake size can be estimated is through input of surface rupture parameters into calculations of seismic moment (M_o), using the equation:

$$M_o = \mu A\bar{u} \tag{3}$$

where; μ = shear modulus of faulted materials

A = area of rupture surface

$\bar{u}$ = avg. displacement over rupture surface

Although this requires a number of assumptions, possible variations in the assumptions used are limited enough that useful estimates can be derived.

The width of a rupture surface is limited by the vertical extent of the seismogenic zone, the base of which can most reliably be determined from the depth distribution of aftershocks of large magnitude events. Earthquakes are generally confined above a fairly abrupt cut-off, taken to represent a rapid brittle-ductile transition (Sibson, 1984). This is fairly well constrained at a depth of about 10-15 km in many tectonically active regions, in particular along the San Andreas fault zone. It is less well constrained in SCI regions due to less seismic activity, but it should be deeper and is likely in the range of 20-25 km (Chen and Molnar, 1983) or even deeper.

The length of a rupture area can be approximated by using surface rupture length. For small events that barely reach the surface, this can drastically underestimate rupture length in the subsurface, but as length increases, the surface length should percentage-wise approach the subsurface length. It is unlikely that a particular limit exists at which surface length becomes an accurate approximation of subsurface length, but this may occur with a length of about two times the down-dip width of the rupture surface (Bonilla and others, 1984).

Utilizing the assumptions of a commonly used shear modulus value of 3×10^{11} dyne/cm^2 (Bonilla and others, 1984), surface rupture parameters reasonably approximating the entire rupture surface, a simple length times width rupture model, rupture of the entire width of a 20 km thick seismogenic zone, and an average displacement of 5 m, a calculated seismic moment of about 1×10^{27} dyne-cm can be derived. This would be a very large event, corresponding to a moment magnitude (M_W) of about 7 1/2.

The vast majority of historical earthquakes have occurred at or near tectonic plate margins and relations like those discussed above are dominated by them. Examinations of the few historical "intraplate" events have suggested that these events may have differences from their "interplate" counterparts. Intraplate events may rupture with higher stress drops, and thus have greater displacements and larger magnitudes for a given rupture length.

Nuttli (1983a, 1983b) examined reported values of m_b (body wave magnitude taken at 1 second periods), M_S (surface wave magnitude), and M_O (seismic moment) for mid-plate and plate-margin earthquakes. From the derived empirical relations he concluded that, for mid-plate earthquakes, seismic moment varies as the fourth power of the corner period (fig. 29), implying increasing stress drop with increasing moment. He thus proposed that, with higher stress drop at larger magnitudes, large rupture lengths are not required for mid-plate events.

Scholz and others (1986) compared thirty large earthquakes for their relations of length and seismic moment (fig. 30), and distinguished these events on the basis of fault type and tectonic setting (see Table 2). They interpreted that a simple scaling relationship is maintained for both interplate and intraplate settings, in which:

M_o is proportional to L^2 or $u = aL$; where;

M_o = seismic moment

L = length

u = slip

a = constant

Values for the constant a were found to be about 1×10^{-5} for the interplate events and about 6×10^{-5} for the intraplate ones, suggesting that, on the average, displacements are six times greater during intraplate earthquakes. Using this relation, an estimated seismic

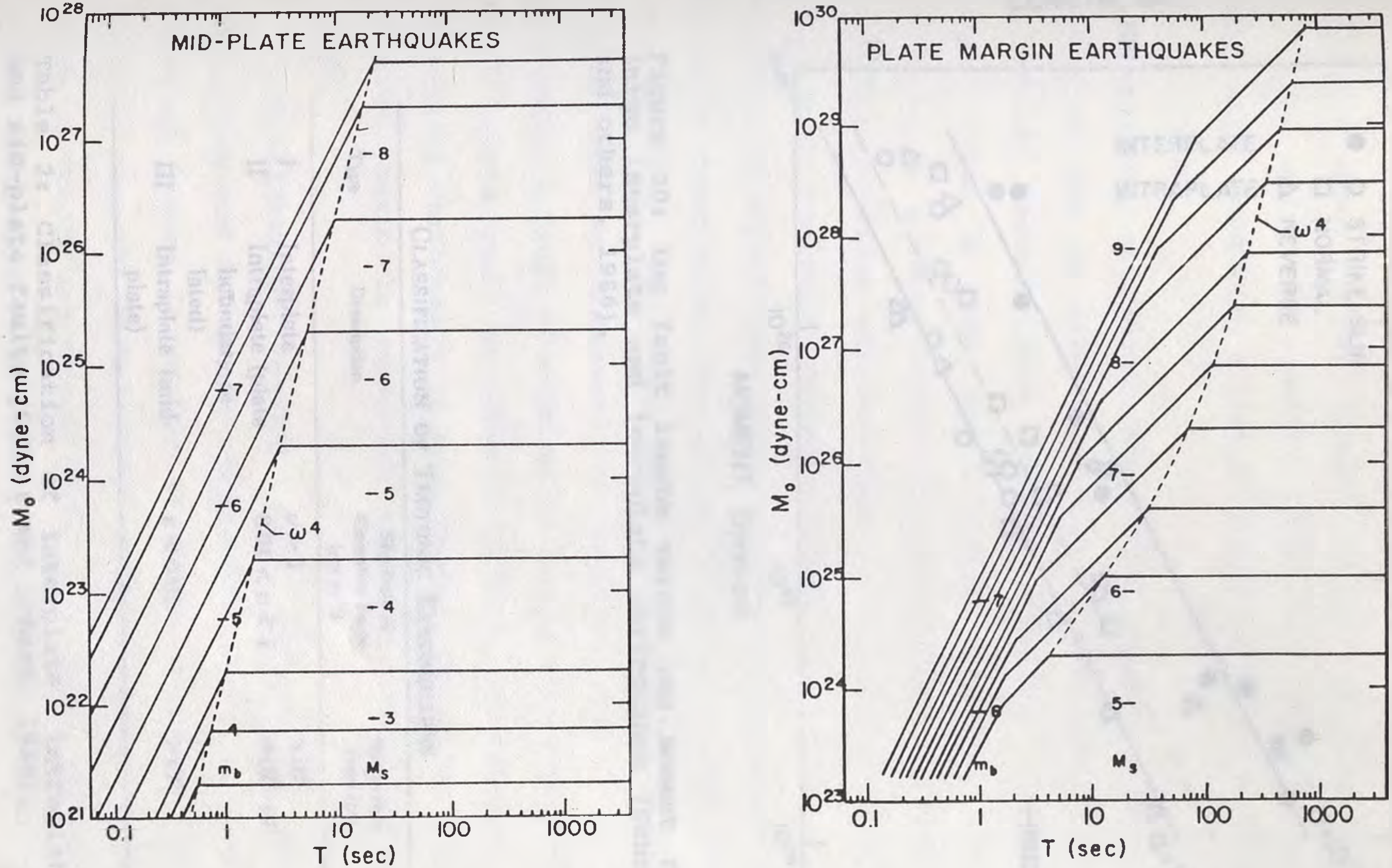


Figure 29: Comparison of spectral scaling relations for mid-plate and plate-margin earthquakes (Nuttli, 1983).

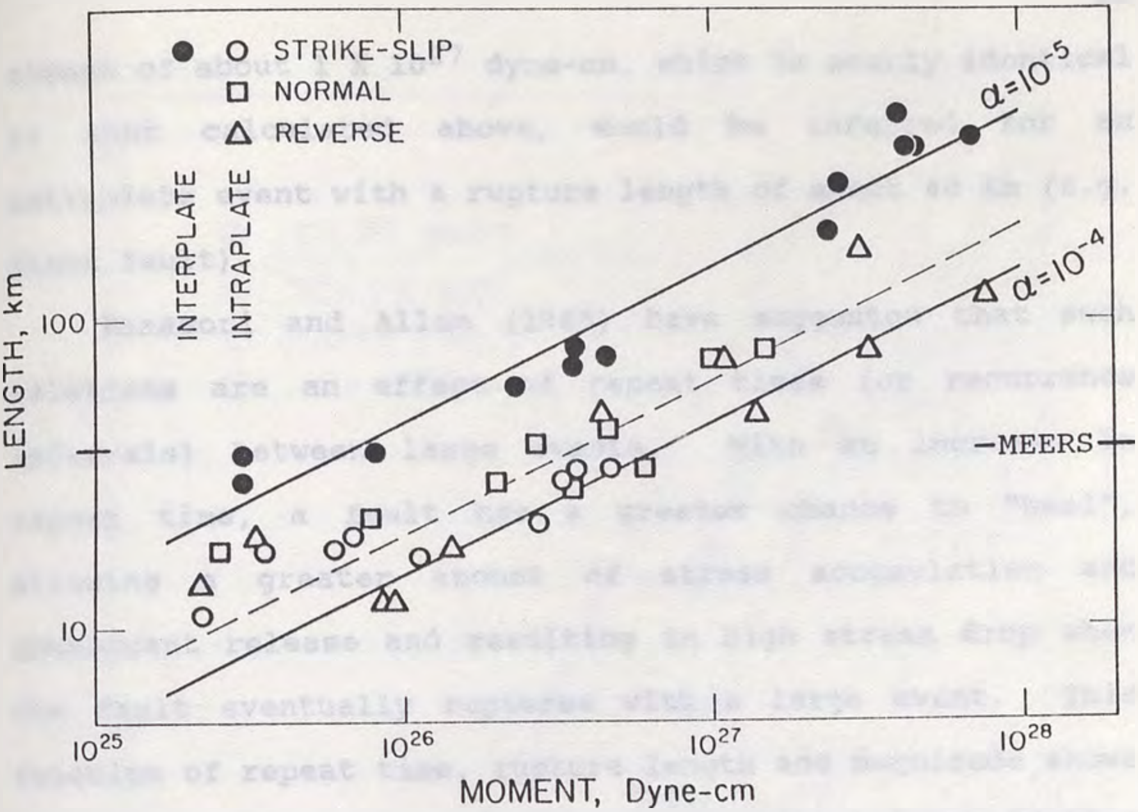


Figure 30: Log fault length versus log moment for large interplate and intraplate earthquakes (Scholz and others, 1986).

CLASSIFICATION OF TECTONIC EARTHQUAKES

Type	Description	Slip Rate of Causative Fault (cm yr^{-1})	Recurrence Time (yr)
I	Interplate	$\upsilon > 1$	$\approx 10^2$
II	Intraplate (plate boundary-related)	$0.01 < \upsilon < 1$	$\approx 10^3$–10^4
III	Intraplate (mid-plate)	$\upsilon < 0.01$	$>10^4$

Table 2: Classification of interplate, intraplate, and mid-plate faults (Scholz and others, 1986).

moment of about 1 X 10^{27} dyne-cm, which is nearly identical to that calculated above, would be inferred for an intraplate event with a rupture length of about 40 km (e.g. Meers fault).

Kanamori and Allen (1985) have suggested that such relations are an effect of repeat times (or recurrence intervals) between large events. With an increase in repeat time, a fault has a greater chance to "heal", allowing a greater amount of stress accumulation and subsequent release and resulting in high stress drop when the fault eventually ruptures with a large event. This relation of repeat time, rupture length and magnitude shows definite correlation, even though there is a large scatter of data (fig. 31). The Meers fault appears to have a repeat time of thousands of years or more. By this treatment, a fault of about 40 km length with a repeat time >2,000 years would be expected to have the potential for generating earthquakes of approximately magnitude 7 1/2 or greater, again in agreement with the calculation made above.

Relationships showing such differences between intraplate and interplate earthquakes have been developed based on events in regions of higher levels of tectonic activity than the south-central United States. As shown in Table 2, Scholz and others (1986) treat these "intraplate" events as an intermediate case, where faults are subject to

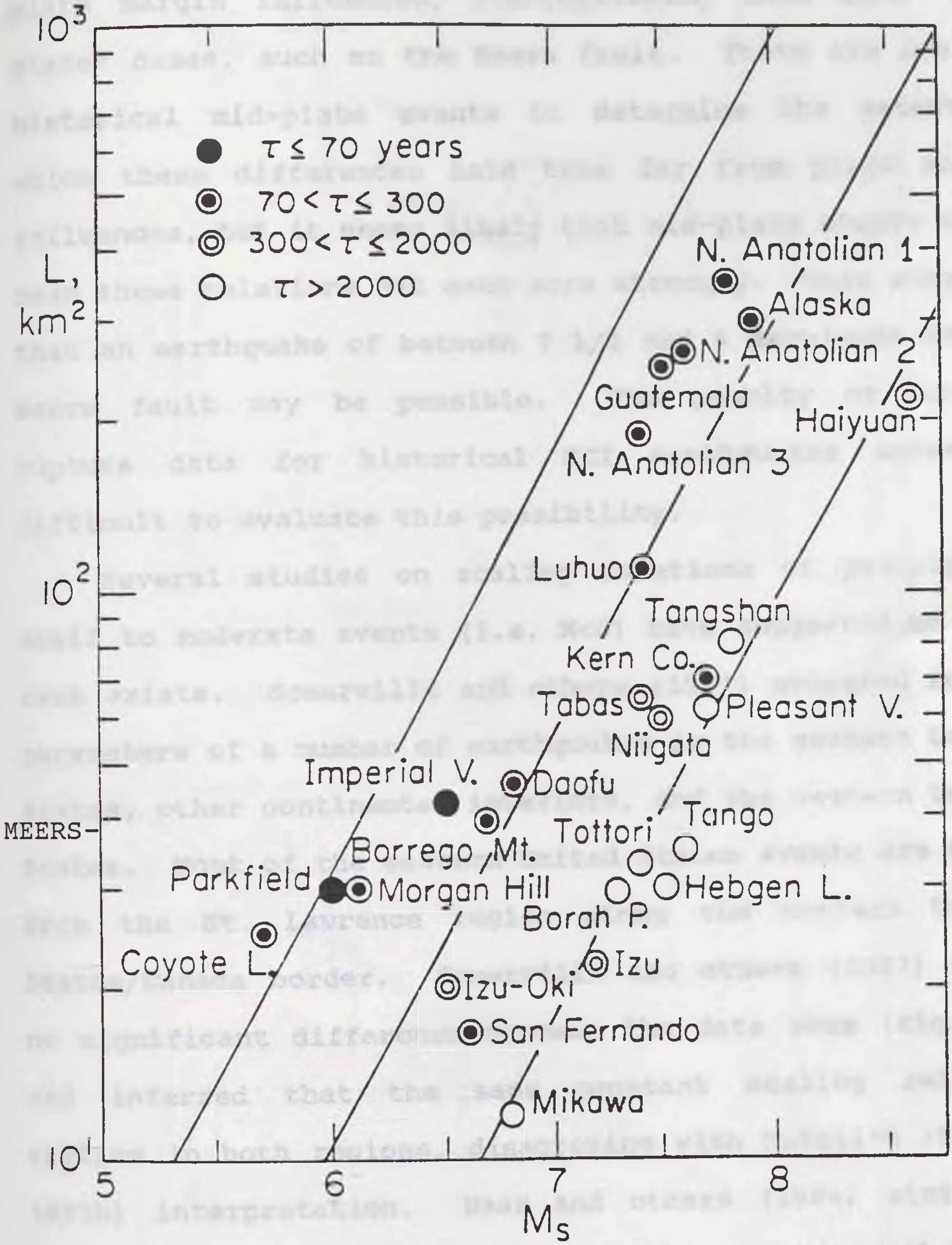


Figure 31: Relation between surface-wave magnitude (M_S), fault length (L), and repeat time for large historical earthquakes (Kanamori and Allen, 1986).

plate margin influences, distinguishing them from "mid-plate" cases, such as the Meers fault. There are too few historical mid-plate events to determine the extent to which these differences hold true far from plate margin influences, but it seems likely that mid-plate events would bear these relations out even more strongly. This suggests that an earthquake of between 7 1/2 and 8 magnitude on the Meers fault may be possible. The paucity of surface rupture data for historical SCI earthquakes makes it difficult to evaluate this possibility.

Several studies on scaling relations of principally small to moderate events (i.e. M<6) have suggested no such case exists. Somerville and others (1987) compared source parameters of a number of earthquakes in the eastern United States, other continental interiors, and the western United States. Most of the eastern United States events are taken from the St. Lawrence region along the eastern United States/Canada border. Somerville and others (1987) found no significant difference between the data sets (fig. 32) and inferred that the same constant scaling relation applies in both regions, disagreeing with Nuttli's (1983a; 1983b) interpretation. Haar and others (1984) similarly found no obvious differences between a series of well-recorded events of the 1983 Enola, Arkansas earthquake swarm and aftershocks of earthquakes in the Mammoth and Oroville, California areas.

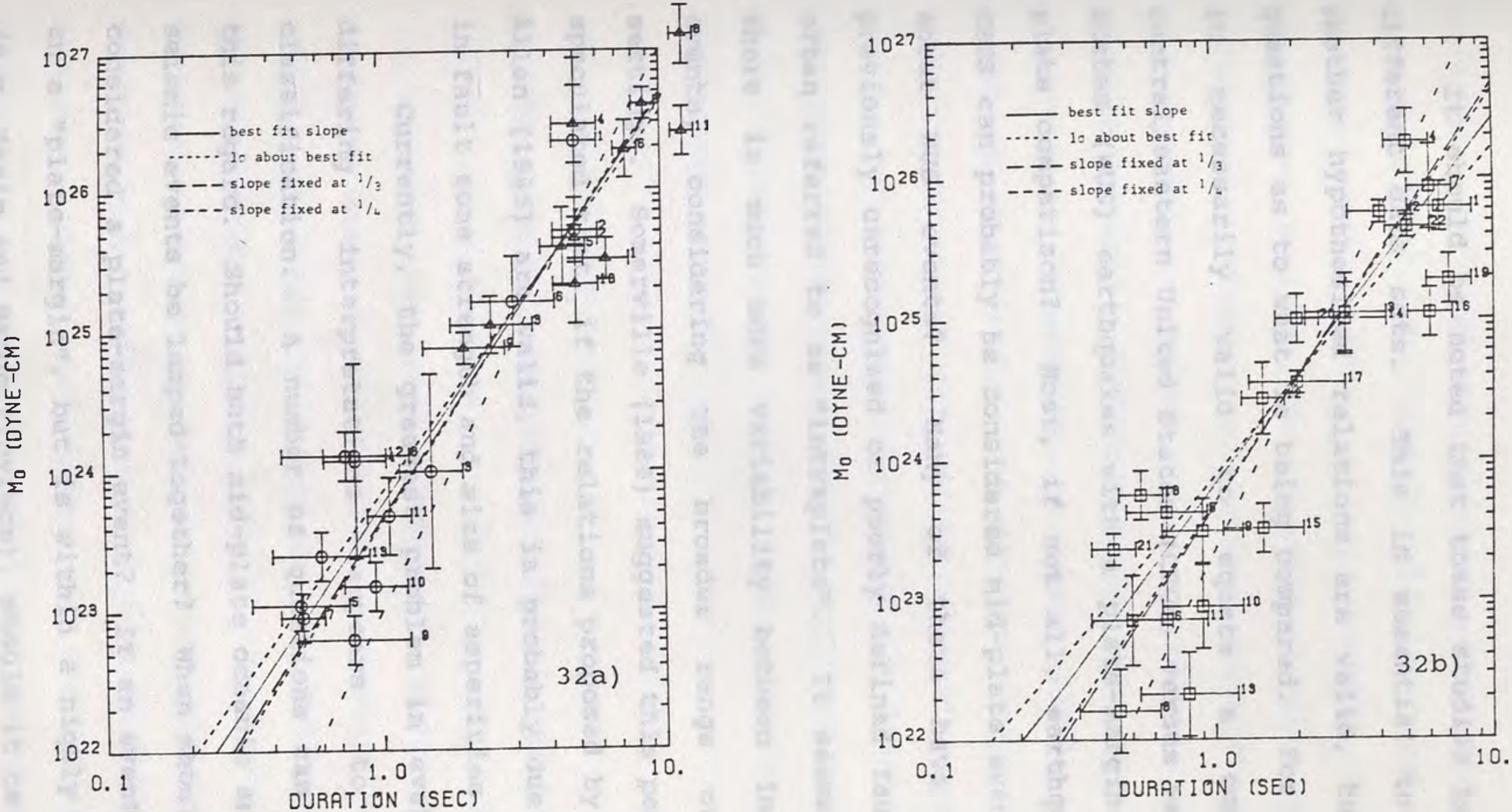


Figure 32: Source scaling relations for 32a) eastern North America and other continental interiors and 32b) western North America (Somerville and others, 1987).

It should be noted that these studies have all used different data sets. This is essential to demonstrate whether hypothesized relations are valid, but it raises questions as to what is being compared. For instance, is it necessarily valid to equate a comparison of central/eastern United States (CEUS) versus western United States (WUS) earthquakes with a plate-margin verses mid-plate comparison? Most, if not all, earthquakes in the CEUS can probably be considered mid-plate events, but what about WUS events? Many of these have occurred on previously unrecognized or poorly defined faults that are often referred to as "intraplate". It seems likely that there is much more variability between individual WUS events, considering the broader range of structural settings. Somerville (1986) suggested this possibility and speculated that, if the relations proposed by Kanamori and Allen (1985) are valid, this is probably due to increases in fault zone strength and size of asperities over time.

Currently, the greatest problem in evaluating these differing interpretations appears to hinge on classification. A number of questions can be raised in this regard. Should both mid-plate oceanic and continental seismic events be lumped together? When should an event be considered a plate-margin event? If an event does not lie on a "plate-margin", but is within a highly active region (e.g. Basin and Range Province), should it be classified as intraplate, or would subclasses of active tectonic domains be appropriate? Do certain mid-plate areas (e.g. St. Lawrence seaway) more closely resemble the WUS than surrounding regions, due to higher rates of activity? Systematic examination of various classification schemes will hopefully resolve these questions. A promising approach toward classification for worldwide events was undertaken by Coppersmith and others (1987). They grouped events by associations with differing types and ages of geologic structures and age of affected crust. However, with so few large magnitude events, it is not possible to make such distinctions for surface faulting events and still retain large enough data sets to make valid statistical arguments.

Impact of a Large Central United States Earthquake

A large magnitude earthquake (M>7) in a heavily populated area of the central or eastern United States (CEUS) would likely be one of the most catastrophic natural disasters in our nation's history. Not since the 1906 San Francisco earthquake has an earthquake of this size occurred in a heavily populated area in the contiguous 48 states. Since that time, our population has grown considerably and we have settled many new areas, increasing the potential for damage in the event of a large earthquake. In many parts of the western United States (WUS), the recognition of a significant seismic potential has led to numerous measures to reduce losses due to earthquakes. Much more could yet be done in the WUS, but such measures lag far behind in the CEUS.

Historical seismic events in the CEUS, notably the 1811-1812 New Madrid, Missouri earthquakes, have affected much larger areas than those in the WUS. The New Madrid series of three great earthquakes were probably felt over all of the United States and southern Canada east of the Rocky Mountains (Nuttli, 1982). The crust in this region is more structurally homogeneous than that in the WUS, resulting in lower attenuation of seismic energy.

We currently cannot hope to accurately predict when or where large earthquakes will occur in the near future, but we can try to recognize potential seismic sources and estimate the sizes and probabilities of events that could be expected. In the absence of a long historical record, geologic indicators can provide the only available evidence of late Quaternary seismic activity. An increased emphasis on such studies could lead to a much better understanding of the overall seismic potential in this region. Even in the WUS, intensive paleoseismic study is relatively new. Most data on ages, extents, and recurrence intervals of prehistorical surface ruptures has been collected in recent years. Collection of such data is a formidable, but not overwhelming, task.

Large magnitude earthquakes in the CEUS are infrequent events. On the one hand, this is fortunate, since most people stand a low probability of being directly affected by such an event and the expected lifespan of most man-made structures is much less than the repeat times of potentially damaging earthquakes. On the other hand, this causes a lower awareness of seismic hazards and hampers efforts to implement earthquake hazard reduction programs. This can make the occurrence of such events all the more damaging.

If a great enough probability for large magnitude earthquakes exists, it is important to assess what impact, if any, such events might have on existing critical structures, and what design requirements should be imposed on new structures. The vast majority of damage and loss of life during earthquakes results from improperly designed structures. If the potential for large earthquakes can be recognized and accurately characterized, steps can be taken to mitigate the impact of such an event. It is impractical to retrofit every existing structure, but a large magnitude earthquake would cause much more damage if no improvements were made on critical structures.

CONCLUSIONS

The Meers fault is unusual in that it is one of only a few known active faults in stable continental interior (SCI) regions. Understanding of this fault could greatly aid in our comprehension of tectonic processes and seismic hazards in the central and eastern United States (CEUS) and other SCI regions.

This fault exhibits clear evidence of at least one large magnitude earthquake in the late Holocene. This is not remarkably surprising, since large magnitude earthquakes do occasionally occur in the CEUS and this fault lies within a major structural zone (i.e. Wichita frontal fault system). However, the prominent expression of this fault scarp is very surprising, since no other such feature as prominent as this is known in a cratonic region. The fact that this fault scarp escaped recognition for so long implies that there is much we need to learn about the late Quaternary tectonic activity and frequency of large magnitude earthquakes in cratonic regions.

The recent displacements have occurred along a reactivated Paleozoic fault. This fault is subvertical, trends N60°W, and is part of the complex Wichita frontal fault system, which bounds the north side of the Wichita uplift. Various lines of evidence along the Amarillo-Wichita-Arbuckle uplift indicate that this zone may have the potential to cause large magnitude earthquakes along much of its length.

The zone of surface ruptures along the Meers fault extends onto Fort Sill Military Reservation and has a minimum length of 37.0 km. Late Quaternary displacement apparently has a left-lateral component dominating an up-to-the-north component by about 3:1 to 5:1. Vertical displacement reaches a maximum of about 5 m and lateral displacement reaches at least 12 and probably 20 m. This apparently occurred during at least two events, with the most recent one accounting for roughly 1/2 the vertical displacement and an undetermined amount of lateral displacement. The most recent event is very young. It is estimated to have occurred between about 1,300 and 600 years before present (Luza and others, 1987). Previous events may be pre-Holocene, possibly indicating a recurrence interval of greater than 10,000 years.

Many key questions remain about the late Quaternary activity of this fault, including specific information on the number, timing, displacements, and sizes of late Quaternary earthquakes. Nonetheless, it seems reasonably clear that the most recent large magnitude earthquake on this fault caused surface displacements of several meters, forming a significant part of the present fault scarp. It is less clear whether this event is representative of seismic events on this or other faults in SCI regions, or whether such an event is simply indicative of the high degree of earthquake variability.

The displacement observed at the surface requires large magnitude events. Comparison of surface rupture length to historical events causing surface rupture would indicate the maximum earthquake magnitude that could be expected would be greater than 6 3/4. Earthquakes in the CEUS and other SCI regions may typically rupture with shorter lengths and larger displacements than similar sized earthquakes in plate-margin settings. If this is true, the underlying cause is believed to be the various processes that act to strengthen fault zones during periods between large magnitude events. A fault zone with a high rate of activity (e.g. San Andreas fault zone) does not have a long enough interseismic interval to allow prolonged growth of asperities and strengthening of the fault zone. With displacements of several meters and a length of only about 40 km, the most recent large magnitude event on the Meers fault seems to support this interpretation.

The possibility that fault zones with low rates of activity have the potential for larger magnitude events than previously believed presents a situation that contradicts conventional methods of assessing seismic hazards. In the past, the larger, more active fault zones were considered to have a higher seismic potential than less active zones. While this holds true for the probability of earthquake occurrence, it may not be valid to assume large magnitude events are confined to highly active zones.

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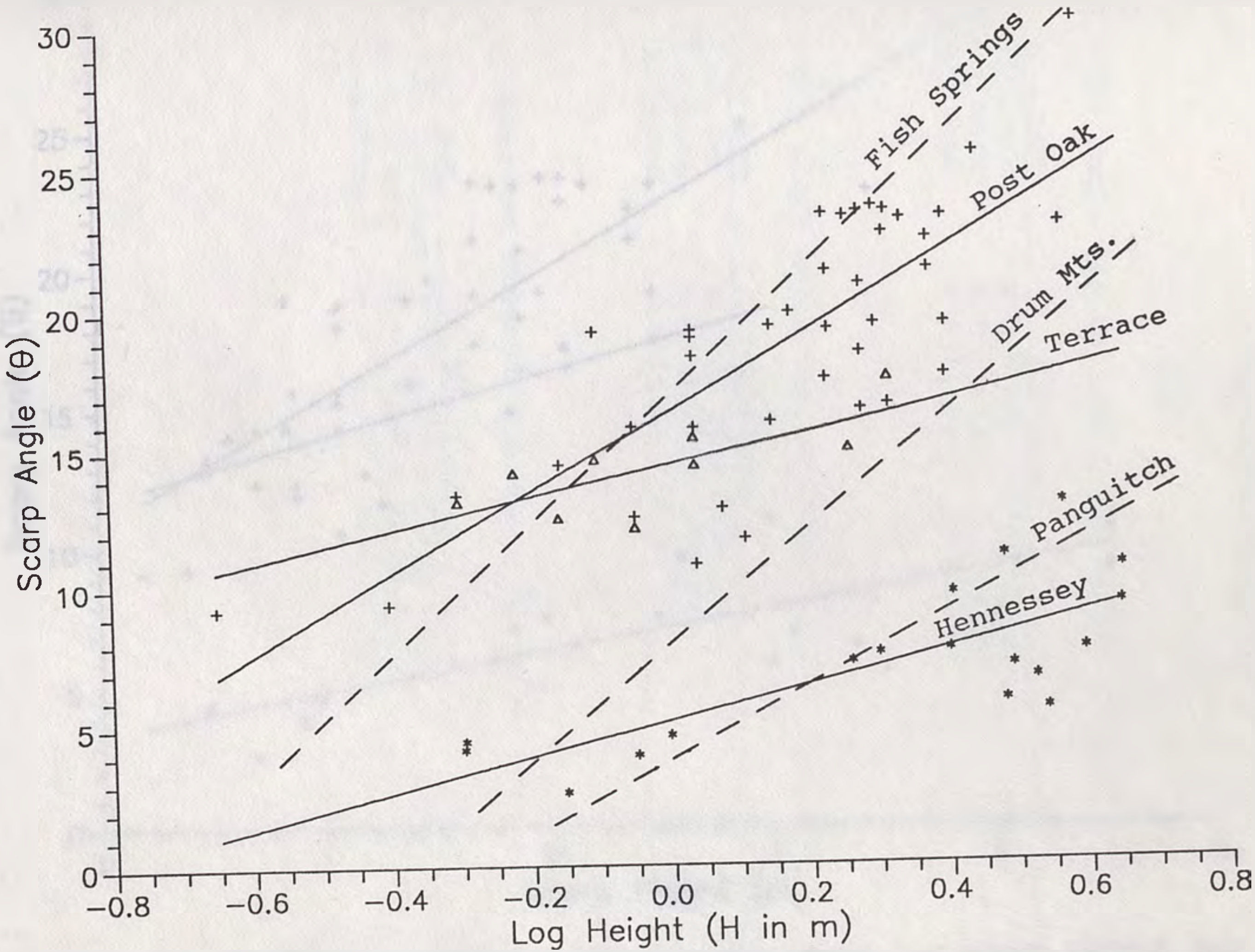

Figure A1: Relation of maximum scarp slope angle and log of scarp height for Meers fault scarp and Bucknam and Anderson (1979) data sets (see Table A1). Crosses = Post Oak; triangles = stream terraces; stars = Hennessey.

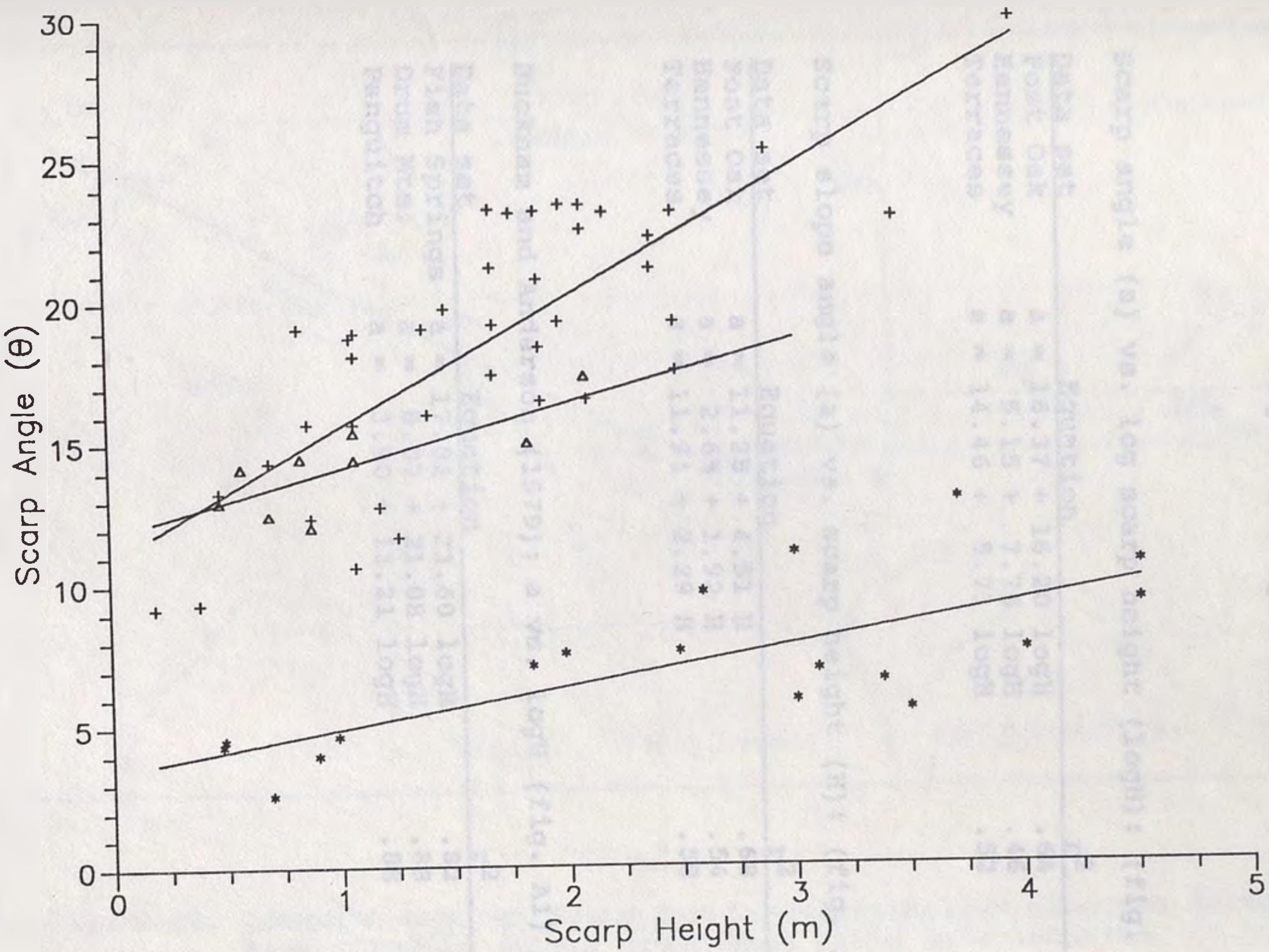

Figure A2: Relation of maximum scarp slope angle and scarp height for Meers fault scarp data (see Table A1).

TABLE A1: Slope Angle vs. Scarp Height
Regression Equations

Scarp angle (a) vs. log scarp height (logH): (fig. A1)

Data set	Equation	r^2	no.
Post Oak	a = 16.37 + 16.20 logH	.64	42
Hennessey	a = 5.15 + 7.76 logH	.46	19
Terraces	a = 14.46 + 5.71 logH	.52	9

Scarp slope angle (a) vs. scarp height (H): (fig. A2)

Data set	Equation	r^2	no.
Post Oak	a = 11.28 + 4.51 H	.68	42
Hennessey	a = 2.65 + 1.92 H	.54	19
Terraces	a = 11.91 + 2.29 H	.58	9

Bucknam and Anderson (1979): a vs. logH (fig. A1)

Data set	Equation	r^2	no.
Fish Springs	a = 17.01 + 23.60 logH	.82	38
Drum Mts.	a = 8.07 + 21.08 logH	.88	49
Panguitch	a = 3.90 + 13.21 logH	.88	11

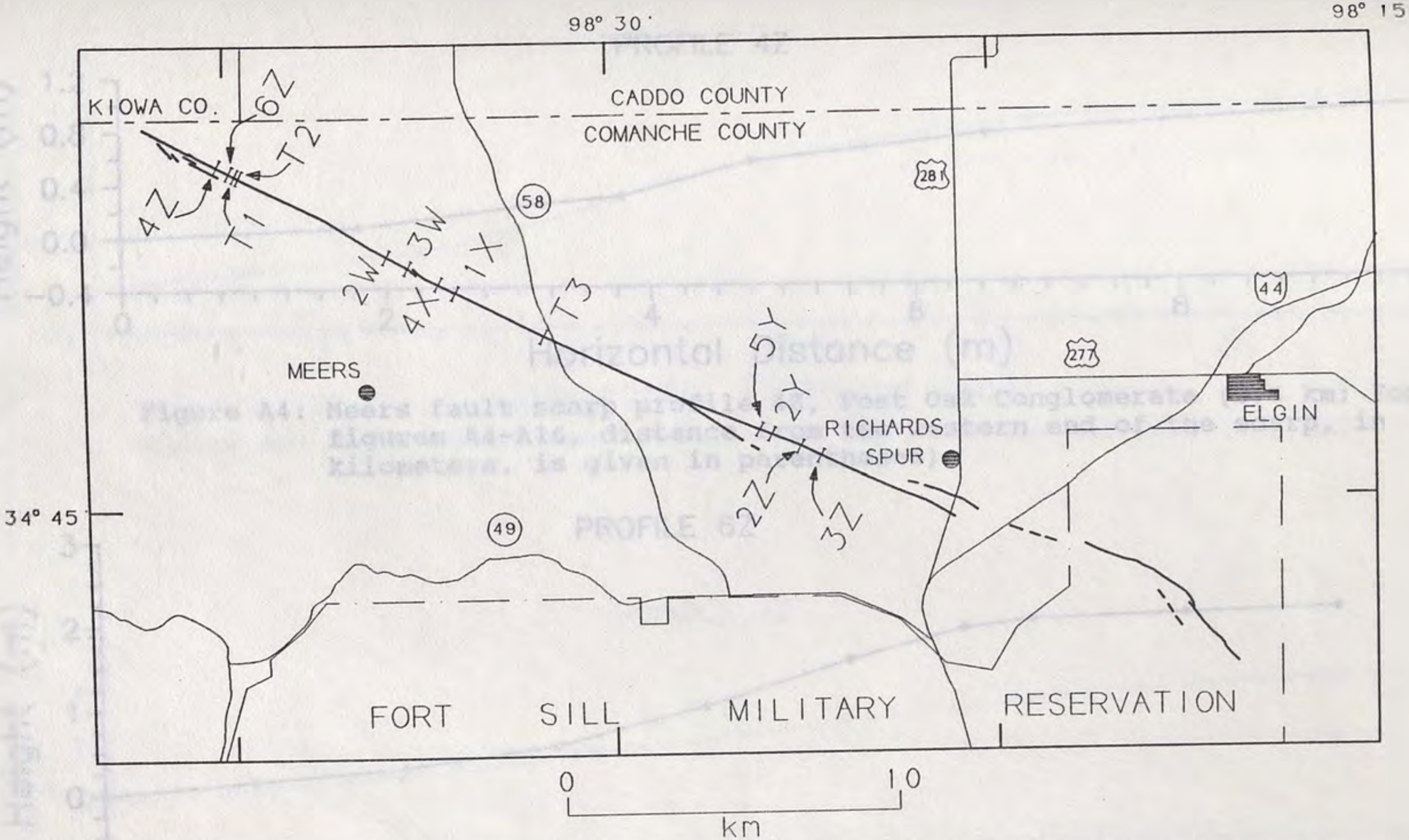

Figure A3: Location map for Meers fault scarp profiles (figures A4 to A16). Profile labels based upon field note location numbers.

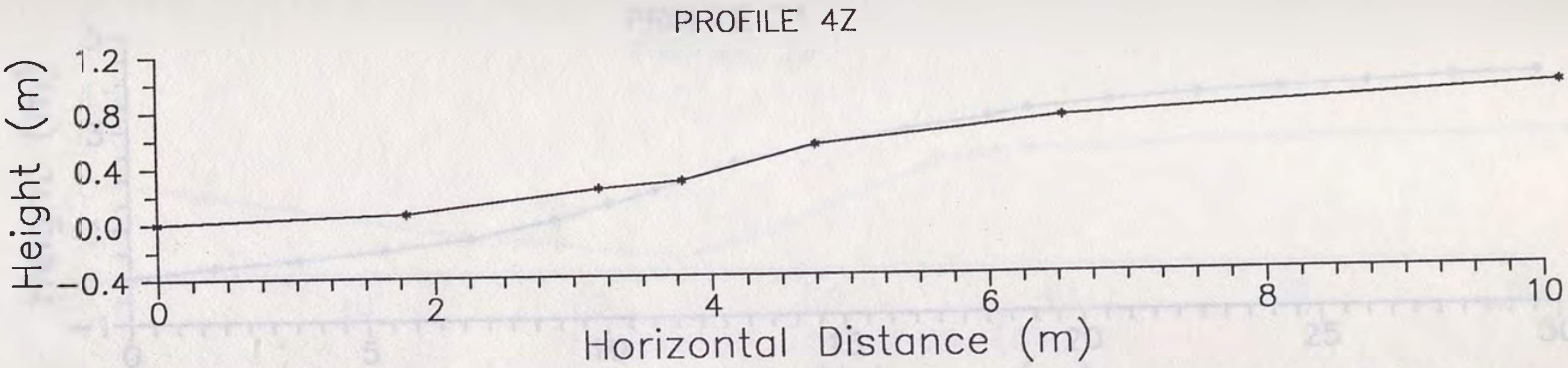

Figure A4: Meers fault scarp profile 4Z, Post Oak Conglomerate (2.6 km; For figures A4-A16, distance from the western end of the scarp, in kilometers, is given in parentheses).

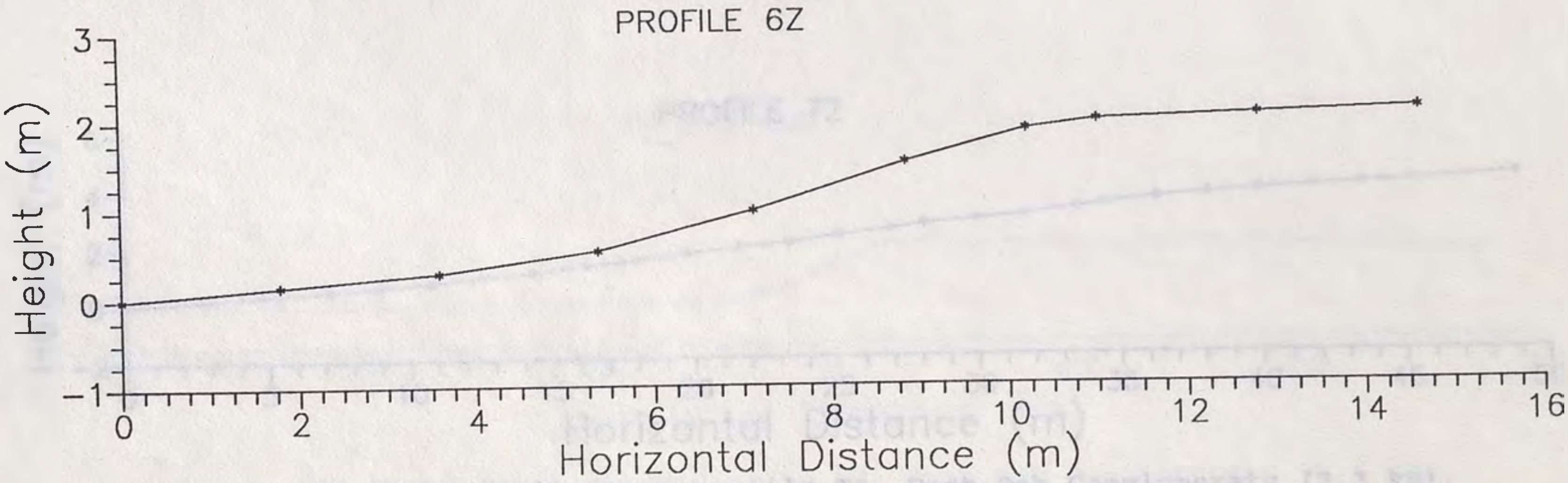

Figure A5: Meers fault scarp profile 6Z, Post Oak Conglomerate (3.0 km).

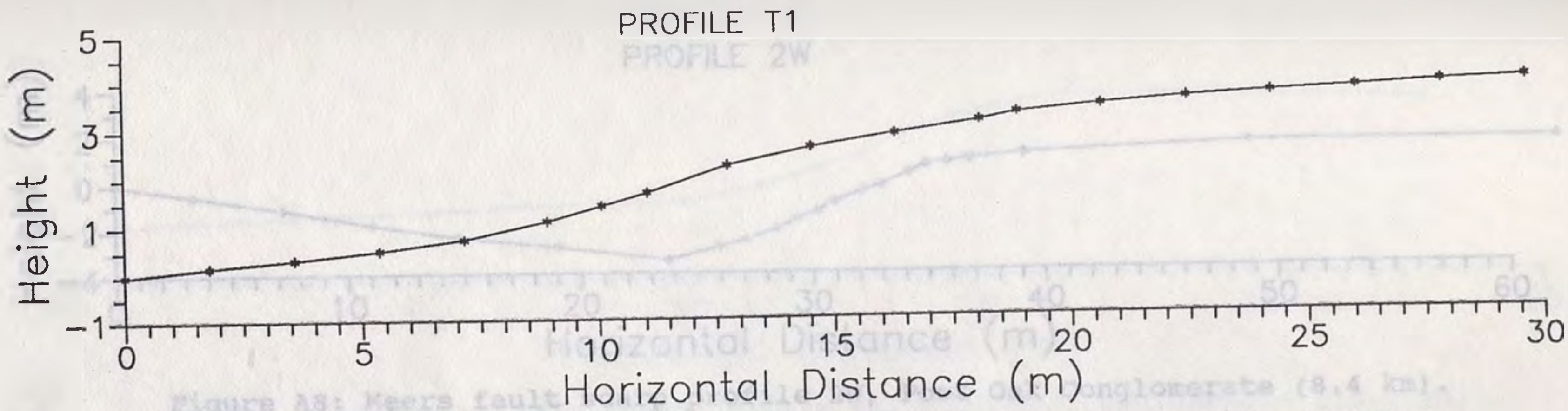

Figure A6: Meers fault scarp profile T1, Post Oak Conglomerate (3.2 km).

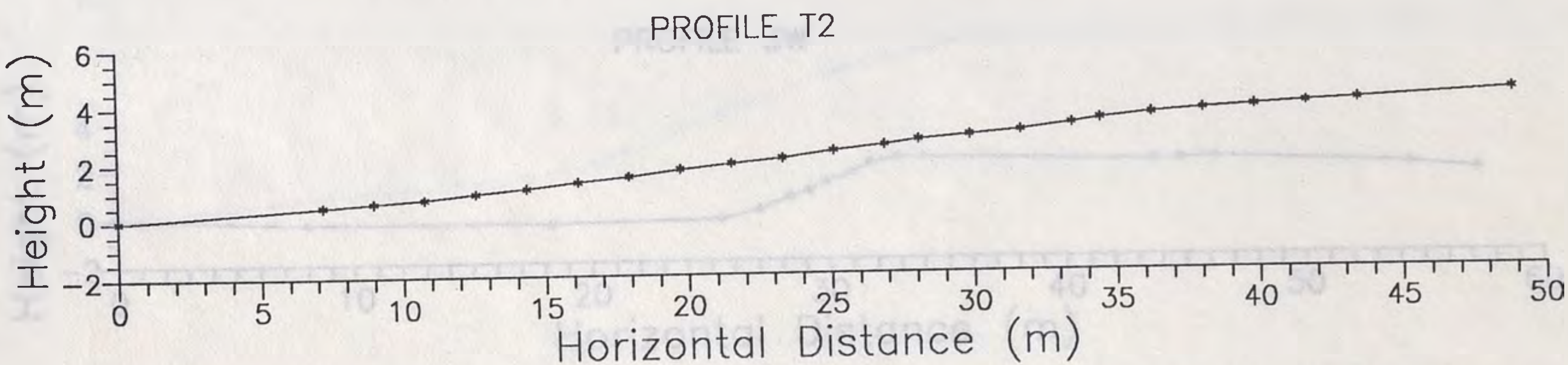

Figure A7: Meers fault scarp profile T2, Post Oak Conglomerate (3.3 km).

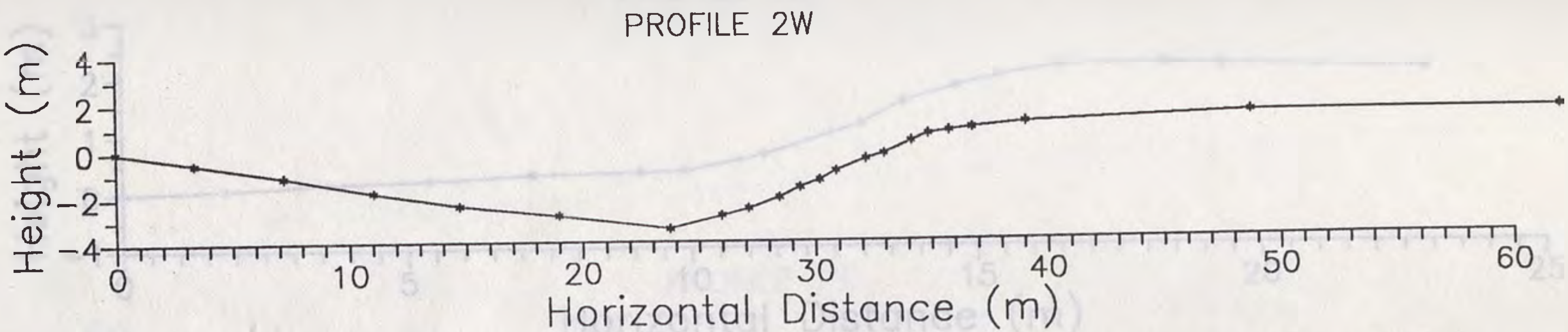

Figure A8: Meers fault scarp profile 2W, Post Oak Conglomerate (8.4 km).

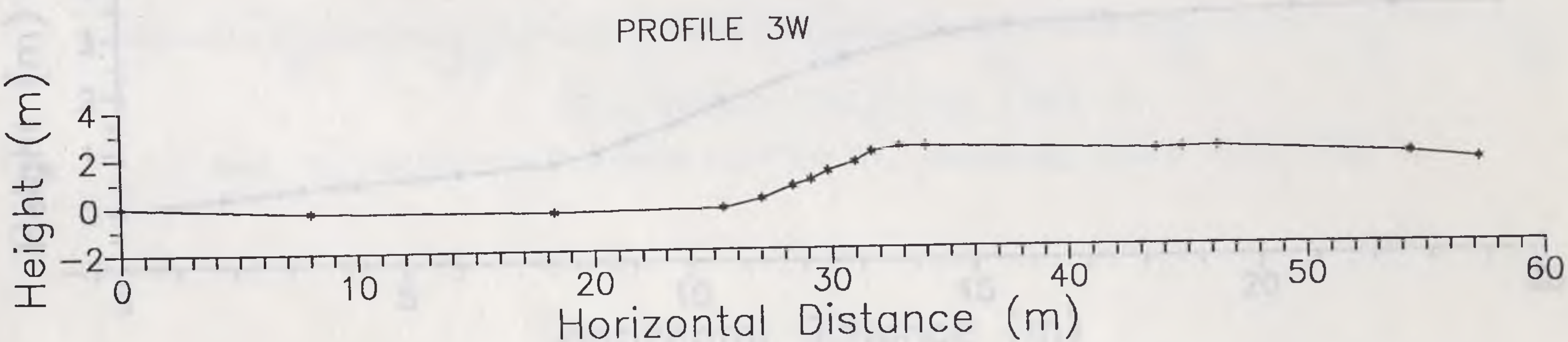

Figure A9: Meers fault scarp profile 3W, Post Oak Conglomerate (9.1 km).

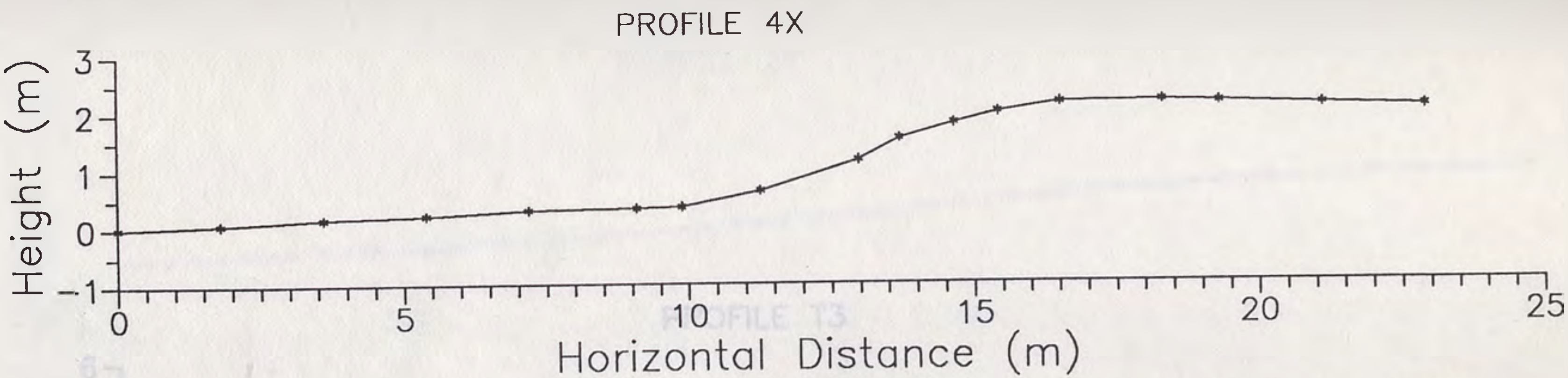

Figure A10: Meers fault scarp profile 4X, Post Oak Conglomerate (10.1 km).

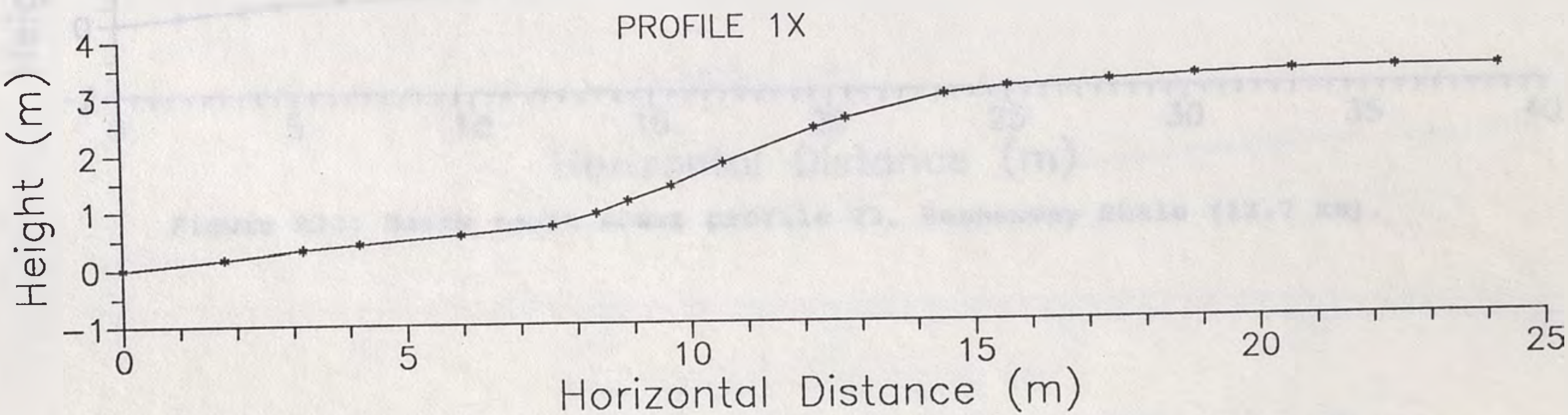

Figure A11: Meers fault scarp profile 1X, Post Oak Conglomerate (10.6 km).

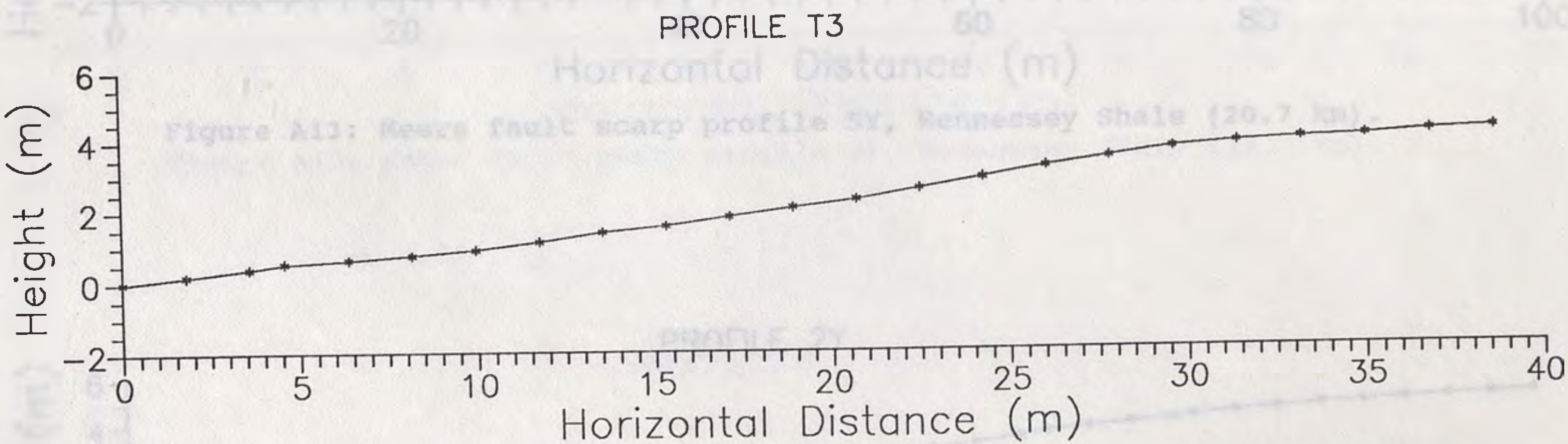

Figure A12: Meers fault scarp profile T3, Hennessey Shale (13.7 km).

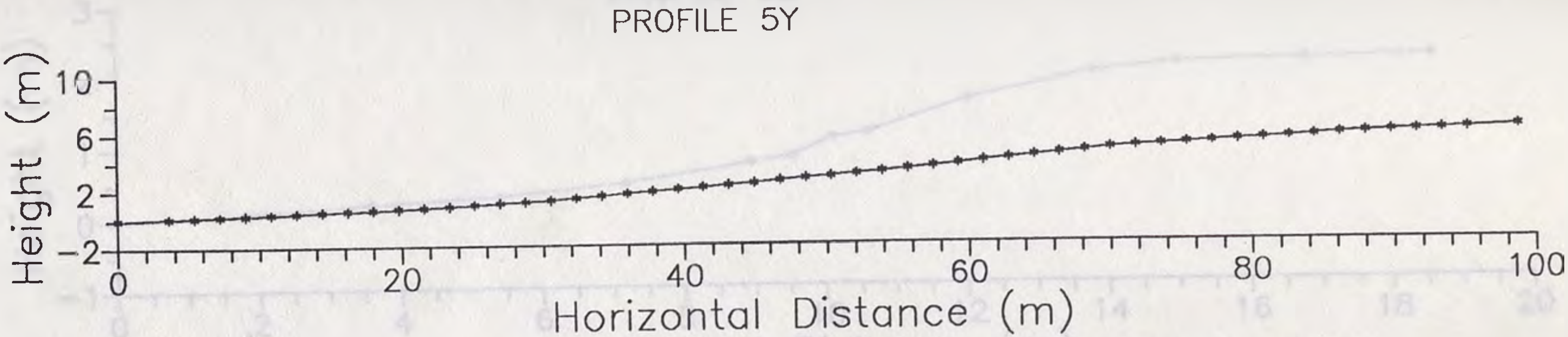

Figure A13: Meers fault scarp profile 5Y, Hennessey Shale (20.7 km).

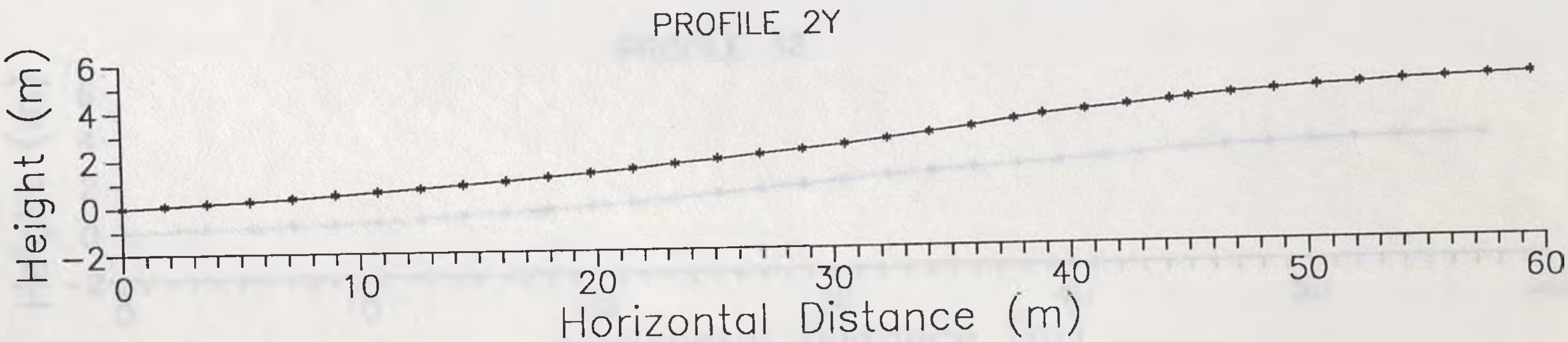

Figure A14: Meers fault scarp profile 2Y, Hennessey Shale (21.1 km).

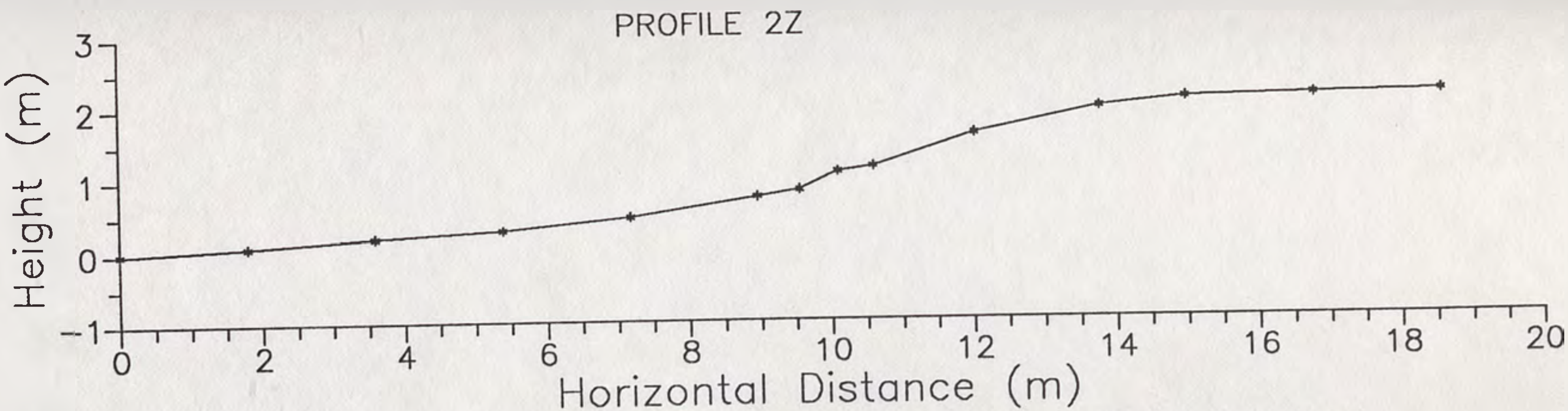

Figure A15: Meers fault scarp profile 2Z, Hennessey Shale (22.0 km).

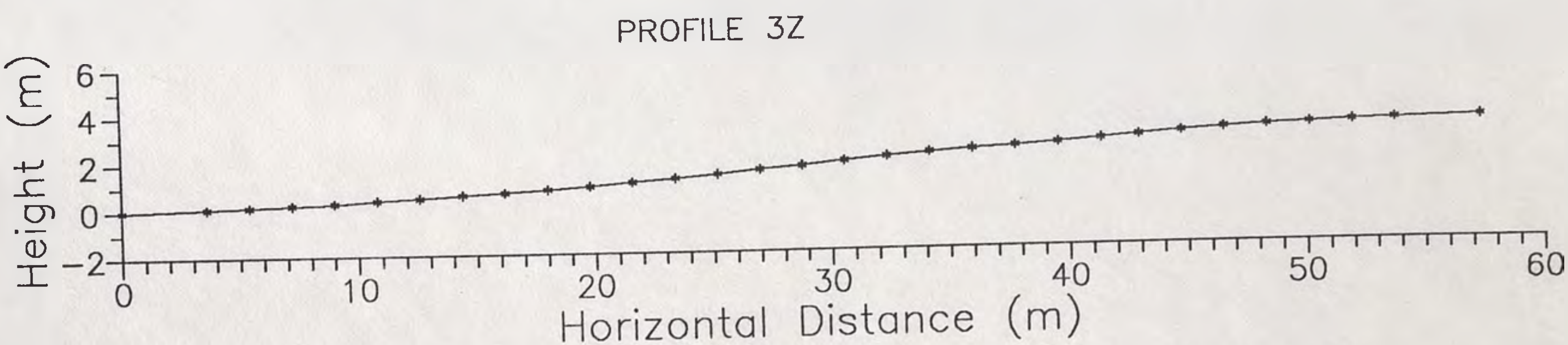

Figure A16: Meers fault scarp profile 3Z, Hennessey Shale (22.8 km).