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Precise Leveling, Space Geodesy and Geodynamics

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IV. Vertical Crustal Movements in the Vicinity of the 1931 Valentine, Texas Earthquake (Bull. Seismol. Soc. Am., in press, 1981).

V. Detection of Vertical Fault Displacements by Precise Leveling in Western Kentucky (Bull. Seismol. Soc. Am., submitted, 1981).

Introduction and Background

Monitoring and understanding crustal movements has been identified as a major goal of NASA's Office of Space and Terrestrial Applications, with implications for earthquake prediction, seismic risk evaluation, engineering application, resource exploration, and crustal dynamics in general. In developing this new space based capability it is essential to all aspects of the program (independent of which system is being considered) to incorporate and build upon currently available information regarding crustal dynamics. In the earliest stages of development, the successful deployment of any crustal movement monitoring network utilizing space techniques will depend upon obtaining accurate information on such questions as: where should land based stations be deployed to most effectively monitor crustal movements in both stable areas for calibration purposes and active areas where movements are most likely to be detected? What precision is necessary to detect crustal movements in different areas? How often should movements be repeated (this is particularly important in light of the possible complex temporal behavior of crustal movements)? In addition, when attempting to measure continent wide deformation and plate tectonic movements, regional and local movements (whether due to tectonic or non-tectonic causes) will be a source of noise which could greatly complicate interpretation of the observations. Tha ability to identify and remove such effects could prove to be the fundamental limit on the applicability of space techniques to studies of plate tectonic movements and continent wide deformation.

For the past 2 years a group at Cornell University under NASA sponsorship has been using U.S. releveling data to investigate these questions. This report briefly summarizes the most recent results of this effort.

Recent Results (for details of particular studies see appropriate appendix)

I. Neotectonic Deformation, Near Surface Movements and Systematic Errors in U.S. Releveling Measurements: Implications for Earthquake Prediction (Appendix II, III)

Analyses of U.S. releveling mearsurements indicate that derivative crustal movement estimates may reflect tectonic deformation, near-surface movements, and/or systematic errors. Discriminating the contributions of these factors is especially crucial for unambiguous geodetic detection of possible precursory seismic deformations. While reliable leveling measurements of co-seismic and post-seismic movements are well documented for some of the larger ($M > 6$) dip-slip earthquakes, leveling evidence for pre-seismic motion is generally sparse and often ambiguous. Subtle earthquake-related motions may be masked by both aseismic movements and systematic errors. For example, deep magma injection and surficial groundwater withdrawal are two mechanisms which have been documented to cause surface movements which, under some circumstances, could be misidentified as seismic-related. Of more concern, perhaps, are systematic measurement errors. Topography dependent errors are an exceptionally troublesome type, perhaps affecting as much as 20% of U.S. leveling. However, other varieties of systematic error also contribute to the uncertainty. Discrepancies between leveling and tide gauge data and within nets of leveling alone suggest large, long baseline accumulations of error. In many cases, aseismic and erroneous contributions cannot be unequivocally determined ex post facto. However, a comprehensive examination of the NGS crustal movement data base, representing a large sampling of the entire U.S. Level Net, provides perspective and criteria needed to begin to recognize movement directly related to earthquake activity.

Perhaps the most extensive set of relevant measurements exists in

southern California, where much attention has recently been focused. Reevaluation of some of these leveling observations indicates that while some appear to reflect tectonic deformation, others are suspect because of indications of systematic errors and/or near-surface, non-tectonic movements. Specifically, possible preseismic movements reported for the 1971 San Fernando earthquake in the vicinity of the earthquake fault as well as approximately 30 km northwest of the epicenter may be due to systematic errors. Movements near the San Gabriel fault, initially ascribed to the Palmdale Bulge and more recently to preseismic effects of the 1971 San Fernando earthquake apparently reflect near-surface sediment compaction due to water table fluctuations. Similarly, there is strong evidence of contamination by rod calibration errors in the releveling observations used to define the southwestern portion of the "Palmdale Bulge" (Llano to Azusa, California). The reality of the "Palmdale Bulge" itself must be questioned in view of this reevaluation. In contrast, possible tilting southwest of Palmdale between 1961 and 1964 is not easily related to systematic errors or near-surface movements and thus may represent tectonic deformation. Whether this tilt anomaly was due to preseismic effects of the San Fernando earthquake or a mechanically separate tectonic event is presently unknown.

II. Vertical Crustal Movements in the Vicinity of the 1931 Valentine, Texas, Earthquake (Appendix IV)

Deformation possibly related to an earthquake in Texas has been revealed from analysis of precise leveling data collected by the National Geodetic Survey. Relative elevation changes deduced from first-order leveling surveys conducted between 1910 and 1957 show that at least two areas in western Texas near the epicenter of the 1931 Valentine earthquake (M = 6.4) subsided

relative to their surroundings. Apparent subsidence east of Van Horn, Texas is attributed primarily to the combined effects of groundwater withdrawal and topography-related survey errors. In contrast, relative subsidence near Valentine, Texas, of 11.2 ± 1.0 cm extending over a distance of about 20 km does not appear to be due to either near-surface effects or leveling errors and thus may represent cosesimic deformation of the Valentine earthquake. The predominance of subsidence over uplift indicated by the leveling data suggests a normal fault mechanism for the earthquake. This is consistent with the Cenozoic structure of the region which is characterized by Basin and Range morphology. If the observed subsidence was caused by the 1931 earthquake, the leveling data suggest that the epicenter for this event lies considerably closer to Valentine than originally reported. This new location is supported by a recent relocation for the Valentine earthquake, and by local intensity reports.

III. Detection of Vertical Fault Displacements by Precise Leveling in Western Kentucky (Appendix V)

Relative vertical displacements of bench marks in extreme western Kentucky have been determined by comparison of successive leveling surveys in 1947 and 1968. The resulting pattern of apparent surface deformation shows a steep offset which can be closely modeled by a normal fault buried in an elastic half-space. The offset is located near the northern boundary of the Mississippi Embayment and the New Madrid seismic zone, an area where faults have previously been inferred on the basis of both geological and geophysical evidence. If the apparent movement is due to slip along a fault, several lines of evidence (regional structure, earthquake data, and lineations) suggest that the postulated fault trends north-northeast.

Thirteen earthquakes were recorded in this area between the times of leveling; focal mechanisms exist for three of these. The nearest of these three focal mechanisms to the leveling offset implies normal faulting. The magnitude of the earthquake, however, appears to be too small to account for the amount of slip required by the fault model. Thus the apparent deformation may have accmulated with several undetected small earthquakes, or gradually as aseismic creep.

IV. An Empirical Basis for Defining Releveling Anomalies: Implications for Crustal Deformation in Southern California

Standard error estimates for precise leveling are based on loop misclosures and on the repeatability (usually two determinations) of height differences between consecutive benchmarks. Standard error estimates as such contain little information on systematic errors. Primarily as a result of crustal movement studies (e.g. Savage and Church, 1974; Brown, 1978; Citron and Brown, 1979; Chi et al., 1980; Jackson et al., 1980; Strange, 1981; Reilinger and Brown, 1981) it is now known that systematic leveling errors can surpass standard error estimates over distances of only a few kilometers. While systematic errors can be recognized in some instances (e.g. Brown, 1978; Chi et al., 1980; Ni et al., 1981; Reilinger and Brown, 1981), in many cases erroneous measurements cannot be unequivocally identified (e.g. Citron and Brown, 1979; Mark et al., 1981).

Because of existing uncertainties as to the cause or magnitude of systematic leveling errors we have begun developing an empirical basis for identifying anomalous releveling signals. This project utilizes the standardized U.S. releveling data base (all movement profiles plotted on common scales) developed at Cornell (Fig. 1). Each profile of apparent elevation change (Fig. 2a shows an example from our data base of a standardized ele-

vation change profile) is approximated by a series of linear tilt segments which are determined by the particular character of the profile (e.g., Fig. 2b). To minimize the influence of unstable benchmarks, each segment must be defined by at least 5 benchmarks and must extend at least 10 km along the route. The relative movement between the end points of each segment is then catalogued along with the information listed in Table 1 (e.g. location, segment length, topography, benchmark spacing, orientation, time interval, etc.). This geodynamics data base will allow quantitative determination of the dependence of apparent movements on any combination of the parameters listed in Table 1. Perhaps most importantly, it will be possible to determine what constitutes an anomalous signal both in terms of the magnitude and spatial extent of observed movements.

There are a variety of problems directly relevant to earthquake prediction which can be addressed with this new data base. For example, the dependence of tilt on topography can be evaulated as a function of tectonic setting (i.e. are apparent movements which correlate with terrain confined to tectonically active areas such as Southern California or do such correlations exist in presumably stable regions?). Addressing this question should help determine to what extent correlations with terrain, which clearly occur in Southern California, result from real movements or systematic errors. Similarly we will investigate the differences and similarities (tilt magnitude and wavelength) between apparent movements in contrasting tectonic settings. Such an investigation may reveal characteristic "signatures" which are unique to seismically active regions and which may prove useful for seismic hazard evaulation.

Although this project is still in the development stage, we have conucted a preliminary study incorporating all U.S. profiles of apparent elevation change in areas of significant relief (slope of $\geq$ 5 m/km). Each profile was approximated by a series of linear tilt segments. Histograms showing the number of segments in each tilt range (.2μ rad. range) were con-

structed for leveling within the Southern California uplift and for leveling throughout the rest of the U.S. (Fig. 3a, b). The following preliminary observations can be made from these histograms: 1) in general, tilts within the region of the Southern California uplift do not differ substantially in magnitude from tilts observed in other areas of significant relief, 2) tilts of less than about 2μ rad. are difficult to distinguish (at least on the basis of tilt magnitude) from "background noise"; 3) many of the larger tilt anomalies (>2μ rad.) have been convincingly associated with tectonic deformation (e.g., Alaska postseismic deformation, 1952, Kern County earthquake deformation, Rio Grande rift, Hebgen Lake, etc.; i.e., this approach to characterizing the data base has effectively separated tectonic deformation from the "Background noise"); 4) at least some of the Southern California releveling measurements fall well above the background level and thus may represent tectonic deformation. In regard to this last point, it is interesting to note that a number of the releveling measurements defining preseismic effects of the 1971 San Fernando earthquake (S.F.P.S. on Fig. 3a) are characterized by rather anomalous tilts. While some of these movements are due to groundwater variations (Reilinger, 1980) others may be due to precursory phenomena. Because of their obvious importance to earthquake prediction in Southern California, further investigation of these particular observations is receiving our highest priority.

V. New Releveling Measurements in the Eastern U.S.

The National Geodetic Survey (NGS) is currently involved in a massive releveling program to update the North American vertical datum (Whalen, 1980). These new measurements are being incorporated in our crustal movement data base as they become available. Figure 4 shows those leveling routes along the east coast of the U.S. for which new crustal movement information has

been obtained. These recent observations are being used to further investigate possible deformation suggested by our earlier studies, including; uplift of the Appalachian Mountains (Citron and Brown, 1979); eastward tilting of the Atlantic Coastal Plain (Brown and Oliver, 1976); and apparent movements associated with a number of structural features along the east coast (Brown, 1978).

Perhaps the most interesting result obtained thus far concerns possible contemporary deformation in the vicinity of Cape Hatteras, North Carolina. Brown (1978) produced and interpreted profiles of apparent elevation change from leveling and tide gauge measurements along the east coast from Calais, Maine to Key West, Florida. This study revealed large discrepancies between the two techniques which are now believed to be due to systematic errors in leveling (Reilinger and Brown, 1981). On the other hand, individual movement features along this coastal line correlate with geologic structures, and may reflect tectonic deformation (Brown, 1978). Steep tilting is indicated near a fault bounding the Connecticut graben, subsidence is noted at the Chesapeake embayment. Sharp subsidence near Cape Canaveral and relative uplift near Cape Fear are in regions where geomorphic and seismic evidence suggest recent movements. The steepest regional tilting indicated by the coastal profile occurs near Cape Hatteras, N.C. Figure 5 shows apparent elevation changes along the Cape Hatteras segment of the coastal route incorporating the 1979 releveling meaasurements. As indicated by Figure 5, the new survey shows continued tilting in the same sense and at approximately the same rate as the earlier measurements. The progressive deformation indicated by multiple surveys, the large magnitude of the observed tilting and the lack of significant terrain along the route (ruling out elevation dependent errors) is strong evidence that these apparent movements represent crustal deformation and not measurement errors. The long wavelength of the

observed tilting argues against near surface effects and for a deep seated tectonic mechanism. Our analysis of these new observations, which is currently in progress, involves testing the leveling measurements for internal consistency and incorporating other geophysical and geological information on neotectonic activity (e.g. tide gauge measurements, deformation of coastal terraces, recent seismic activity, etc.). Should the measurements stand up to this analysis, they will constitute the most persuasive evidence for neotectonic deformation in the east yet uncovered by leveling. As the east coast of the U.S. has experienced major historic earthquakes (e.g. Charleston, 1886), the possibility that anomalous deformation in the Cape Hatteras area reflects a potential seismic hazard will be carefully examined. This example illustrates the importance of the new leveling measurements being collected by the NGS for evaluating the current state of tectonic activity in the U.S. Incorporating these measurements into the existing data base will represent a significant part of our future effort.

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TABLE I.

Information Tabulated for U.S. Releveling Data Base

profile designation -	all profiles have been assigned a unique identifier which allows immediate retrieval of profile and location on U.S. index map.
profile segment # -	all profiles are divided into a number of sections each of which has been approximated by a straight line segment.
relative movement -	relative movement (mm) between the ends of segment.
length -	distance along leveling route (km) for segment.
topography -	height difference (m) between ends of segment.
topography max. -	maximum topographic height difference (m).
bench mark spacing -	average bench mark spacing for segment (km).
orientation -	approximate orientation (10 degree intervals) for segment.
location -	location code for beginning and ending points of profile segment (using U.S. Physiographic Provinces)
base year -	time (year) when base survey was conducted (base survey is generally the last survey along route)
releveling year -	time of releveling (note: if a route has been releved more than once, each releveling defines a different segment).
Southern California code -	this code identifies profile segments located within the area of the Southern California uplift.

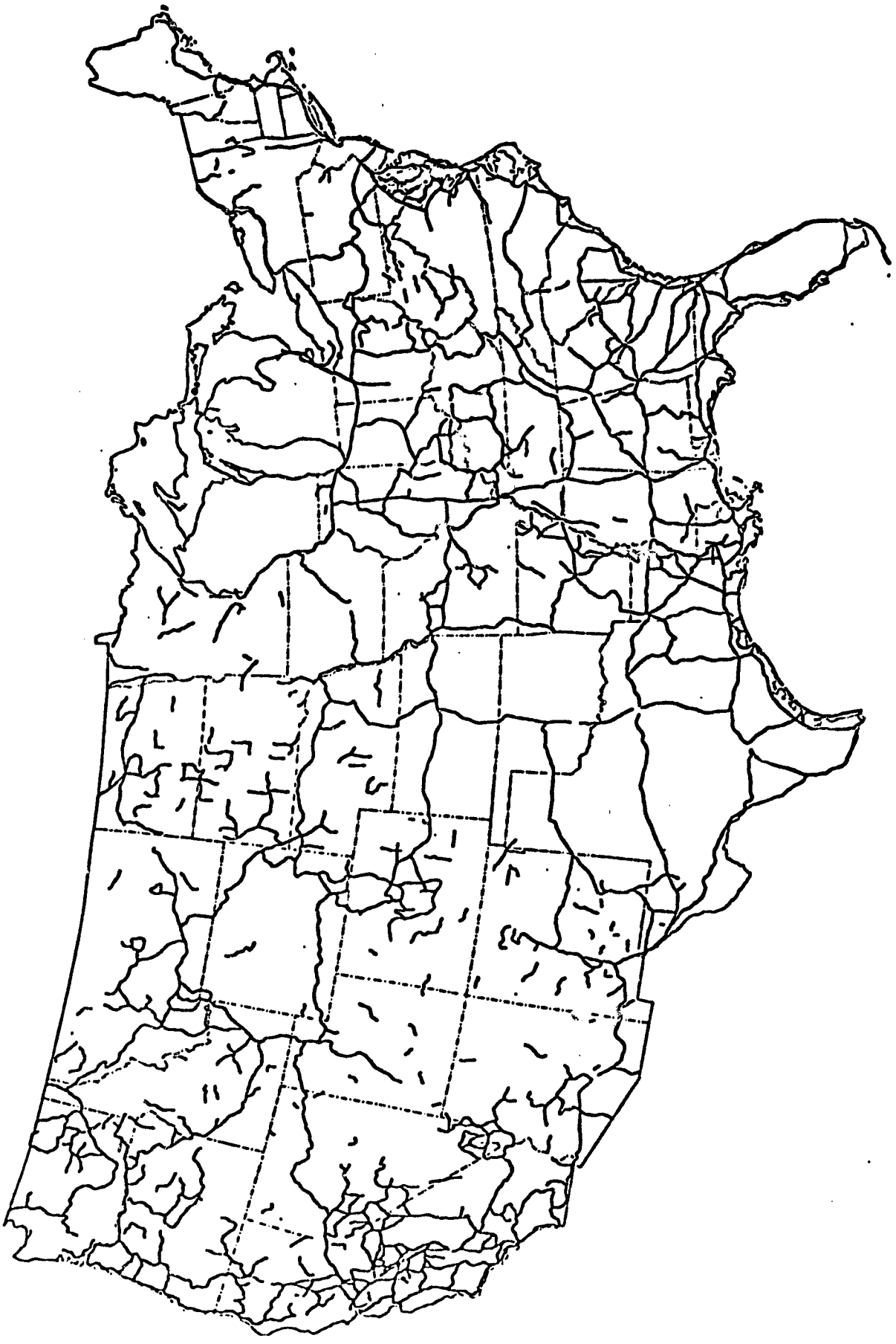

Figure 1: Leveling routes for which crustal movement information is available.

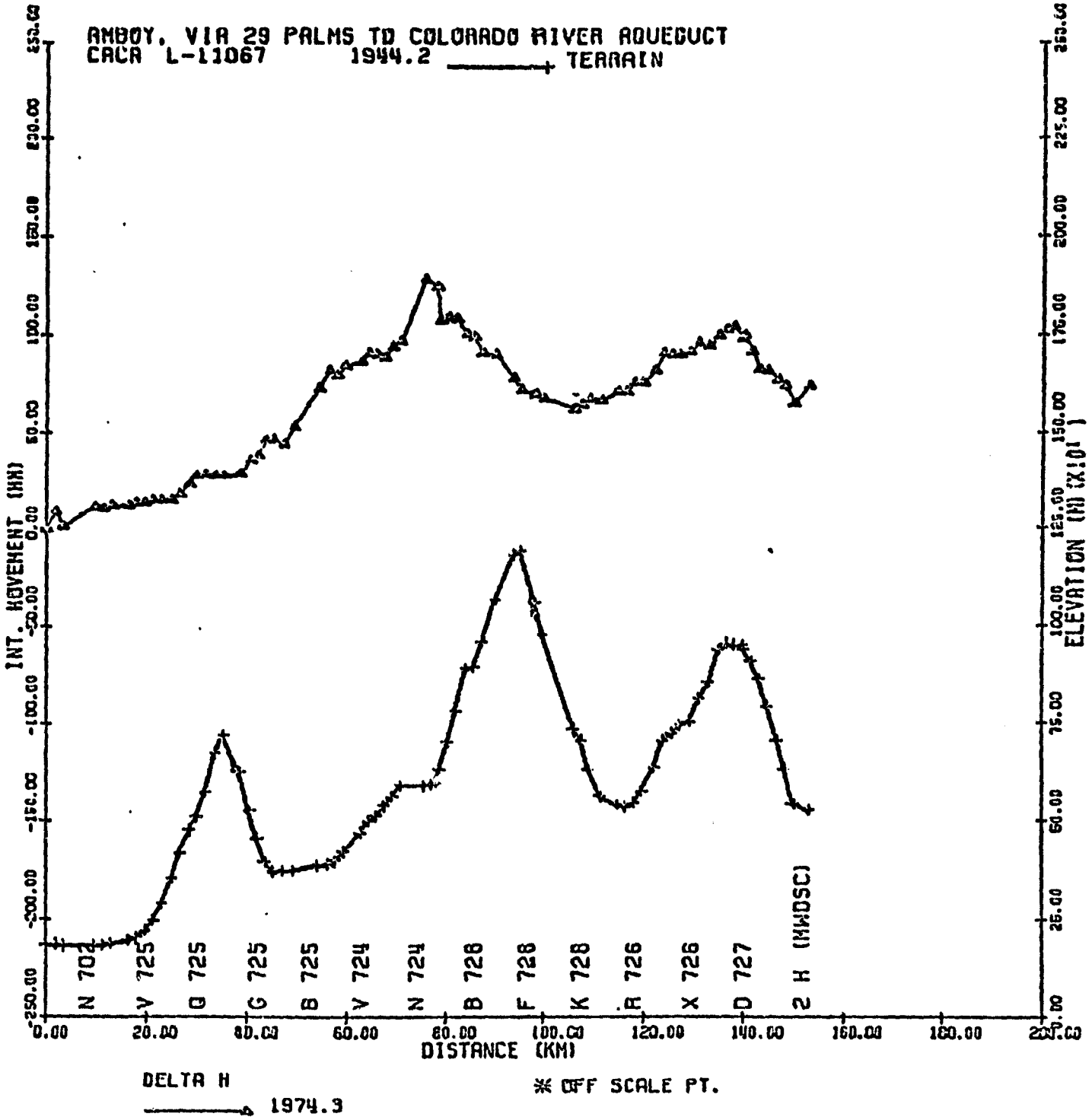

Figure 2a: Example from standardized crustal movement data base developed at Cornell.

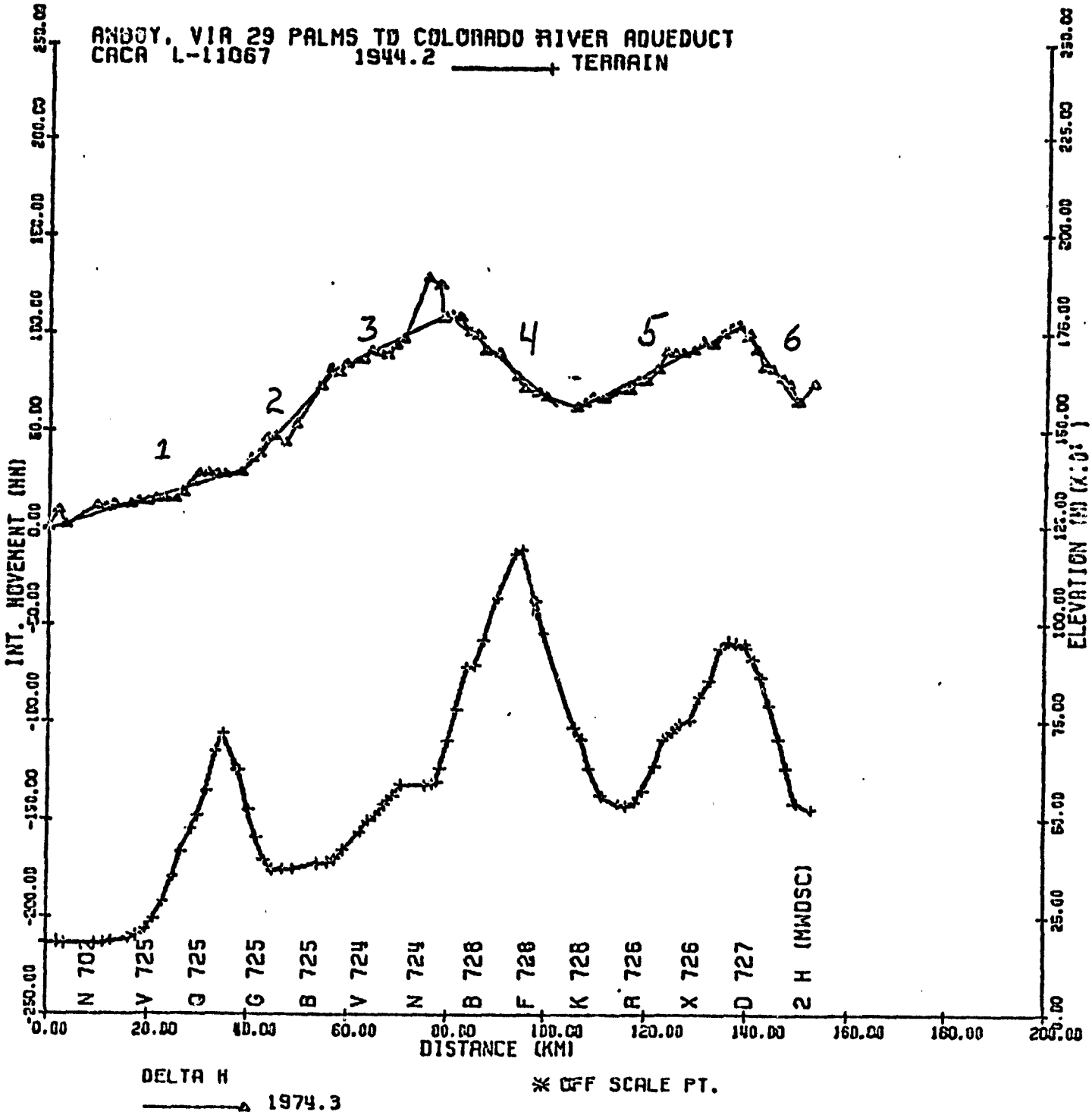

Figure 2b: Example of tilt segments used to approximate elevation change profiles.

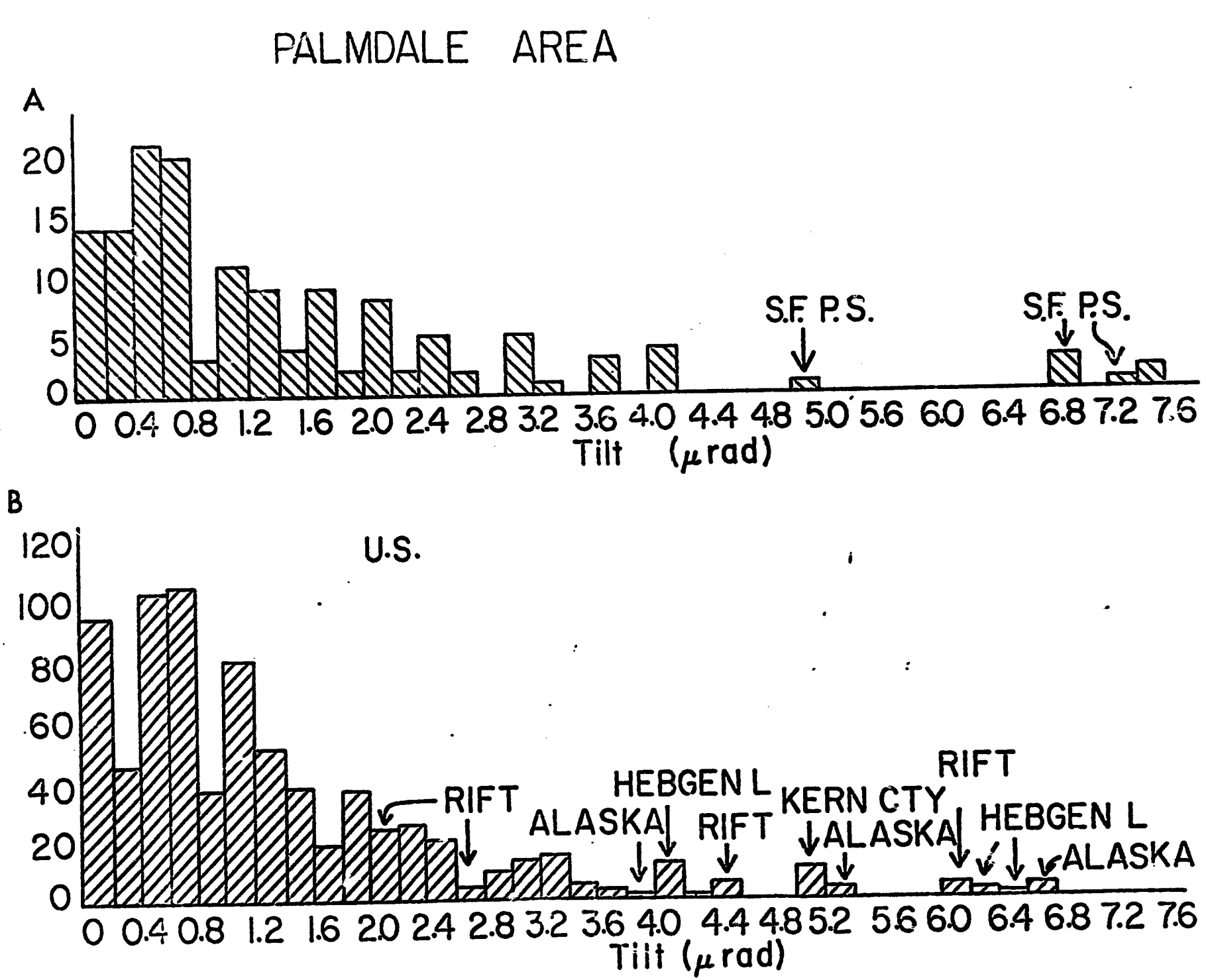

Figure 3: Histograms showing number of tilt segments in each tilt range(0.2 μrad) for: a) elevation change profiles within the region of the Palmdale Bulge; b) in other areas of significant relief.

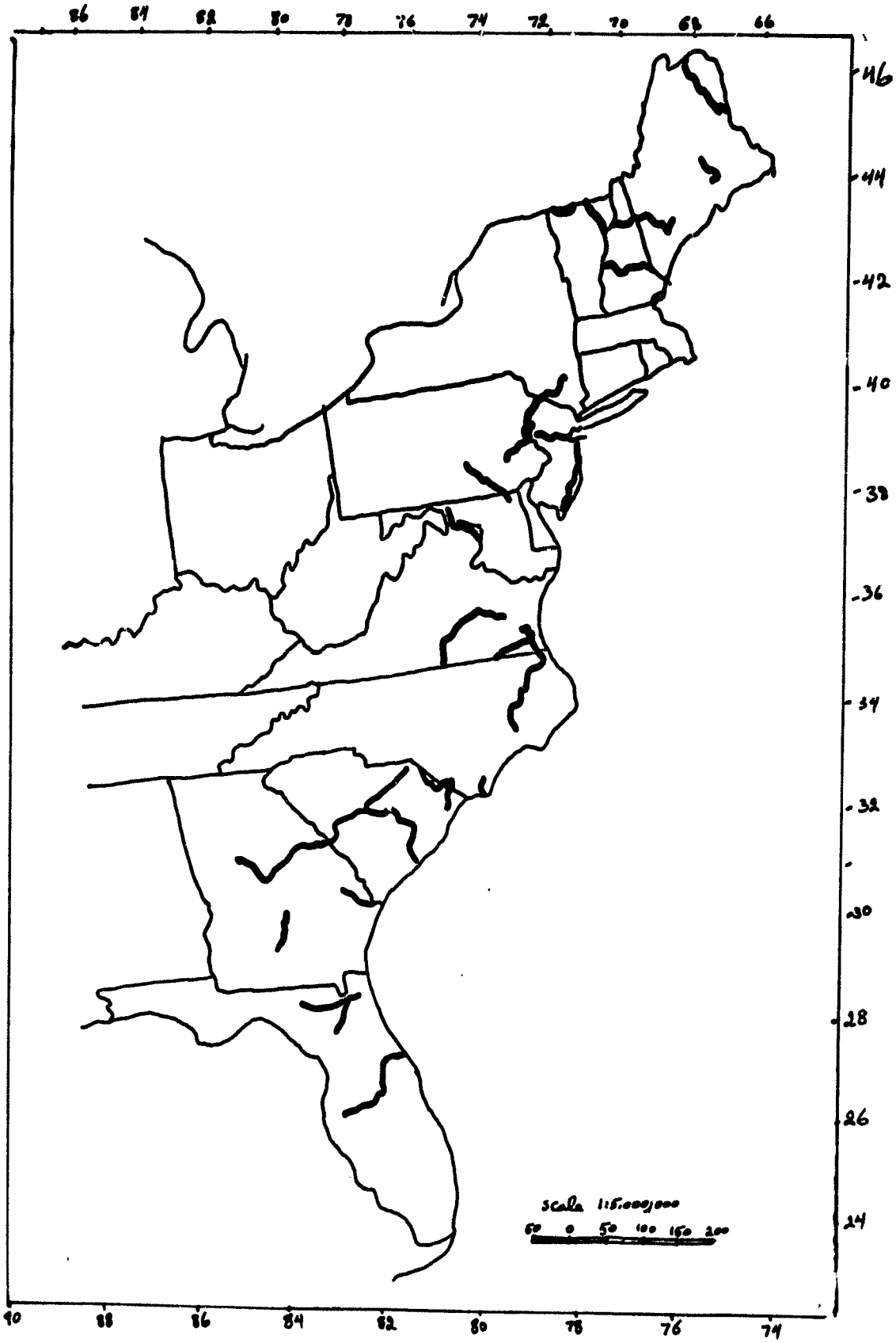

Figure 4: Location of recently resurveyed leveling routes in the eastern U.S.

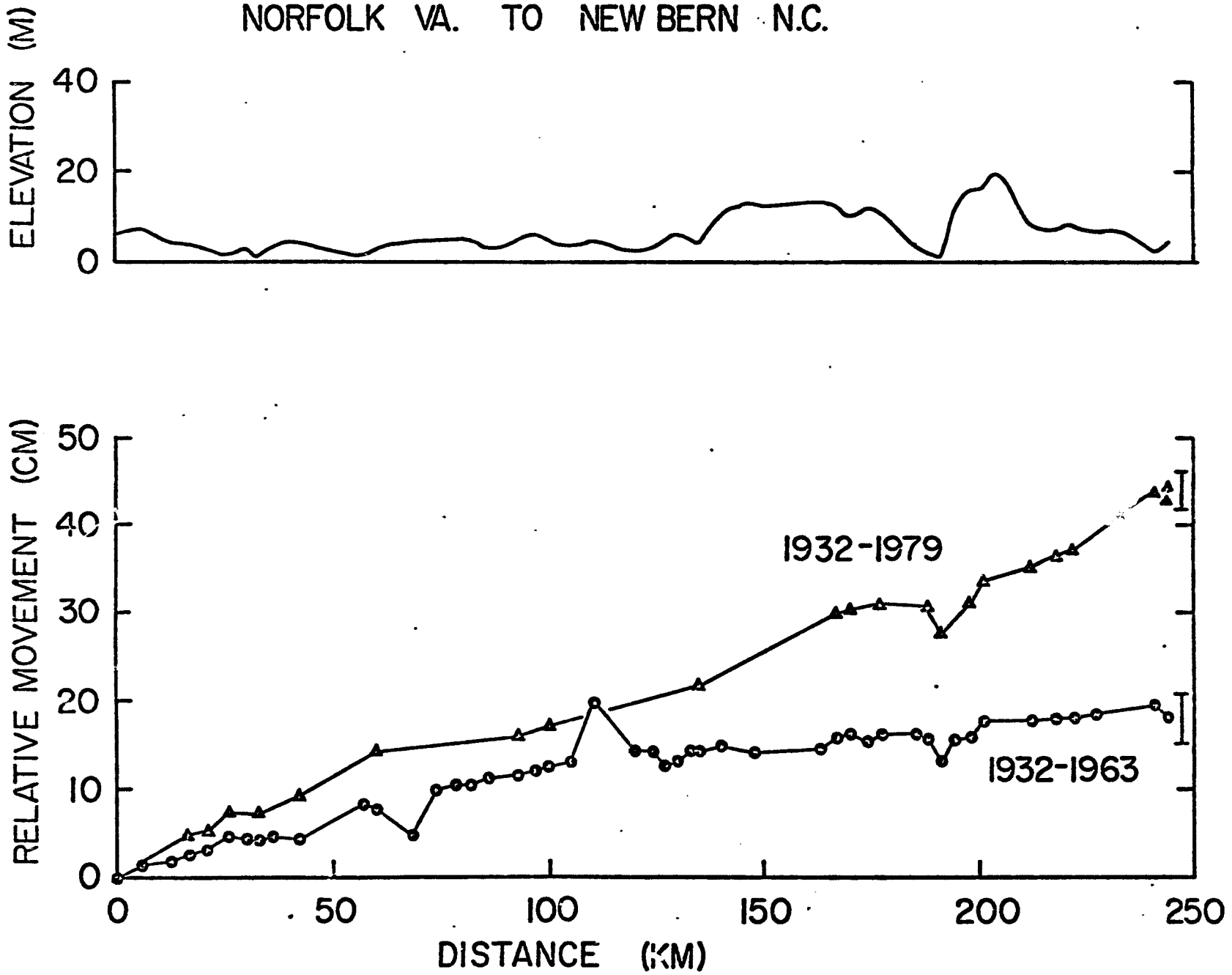

Figure 5: Profiles of elevation change for leveling route crossing Cape Hatteras, North Carolina (see Figure 4 for location).

APPENDIX I

CRUSTAL MOVEMENT STUDIES AT CORNELL*
(numbers refer to index map)

January 1981

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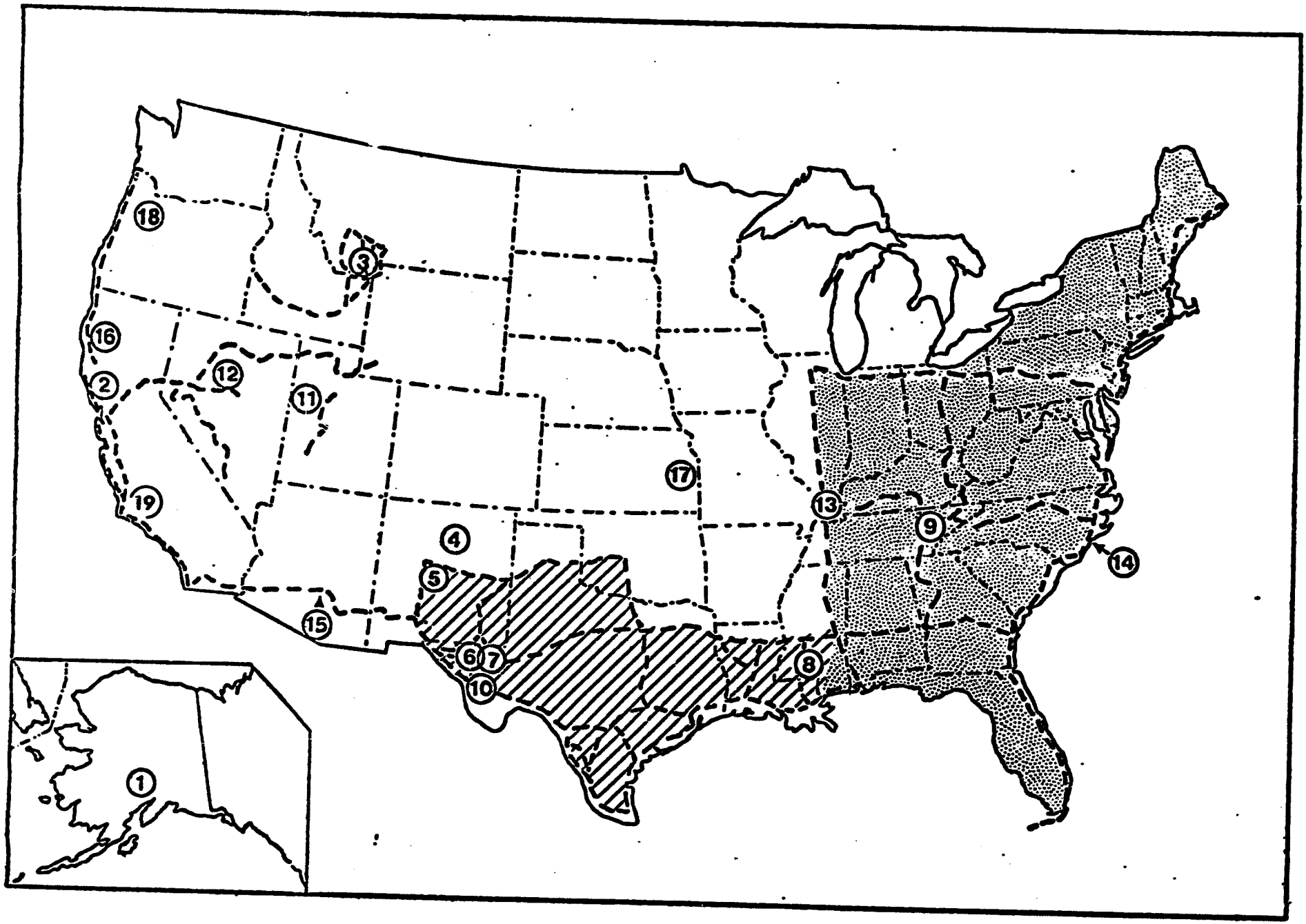
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APPENDIX II

NEOTECTONIC DEFORMATION, NEAR-SURFACE MOVEMENTS AND SYSTEMATIC ERRORS IN U.S. RELEVELING MEASUREMENTS: IMPLICATIONS FOR EARTHQUAKE PREDICTION

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Abstract. Analyses of U.S. releveling measurements indicate that derivative crustal movement estimates may reflect tectonic deformation, near-surface movements, and/or systematic errors. Discriminating the contributions of these factors is especially crucial for unambiguous geodetic detection of possible precursory seismic deformations. While reliable leveling measurements of co-seismic and post-seismic movements are well documented for some of the larger (M > 6) dip-slip earthquakes, leveling evidence for pre-seismic motion is generally sparse and often ambiguous. Subtle earthquake-related motions may be masked by both aseismic movements and systematic errors. Deep magma injection and surficial groundwater withdrawal are two mechanisms which are shown to cause surface movements which, under some circumstances, could be misidentified as seismic-related. Of more concern, perhaps, are systematic measurement errors. Topography dependent errors are an exceptionally troublesome type, perhaps affecting as much as 20% of U.S. leveling. However, other varieties of systematic error also contribute to the uncertainty. Discrepancies between leveling and tide gauge data and within nets of leveling alone suggest large, long baseline accumulations of error. In many cases, aseismic and erroneous contributions can not be unequivocally determined ex post facto. However, a comprehensive examination of the NGS crustal movement data base, representing a large sampling of the entire U.S. Level Net, provides perspective and criteria needed to begin to recognize movement directly related to earthquake activity.

Perhaps the most extensive set of measurements relevant to earthquake activity exists in southern California, where much attention has recently been focused. Reevaluation of some of these leveling observations indicates that while some appear to reflect tectonic deformation,

others are suspect because of indications of systematic errors and near-surface, non-tectonic movements. Specifically, possible preseismic movements reported for the 1971 San Fernando earthquake in the vicinity of the earthquake fault as well as approximately 30 km northwest of the epicenter may be due to systematic errors. Movements near the San Gabriel fault, initially ascribed to the Palmdale Bulge and more recently to preseismic effects of the 1971 San Fernando earthquake apparently reflect near-surface sediment compaction due to water table fluctuations. Similarly, there is strong evidence of contamination by rod calibration errors in some of the releveling observations used to define the southern portion of the Palmdale Bulge (Llano to Azusa, California). The reality of the Palmdale Bulge itself must be questioned in view of this reevaluation. In contrast, possible tilting southwest of Palmdale between 1961 and 1964 is not easily related to systematic errors or near-surface movements and thus may represent tectonic deformation. Whether this tilt anomaly was due to preseismic effects of the San Fernando earthquake or a mechanically separate tectonic event is presently unknown.

Introduction

Vertical movements of the earth's crust are commonly expected to accompany the various phases of strain buildup and release associated with major earthquakes. Observations of vertical co-seismic and post-seismic movements using precise leveling are well-documented. However, reports of preseismic movement in the U.S. are rare and, as will be argued, questionable. Recognition of true pre-seismic motion is complicated by systematic leveling errors, near-surface non-tectonic processes (e.g., fluid withdrawal), the general lack of sufficiently redundant and extensive surveys, and the fact that significant changes in elevation have been identified which are unrelated to earthquakes. Such "noise" could easily hide a pre-seismic signal. Considerable uncertainty exists as to the extent and magnitude of these obscuring "movements". Direct determination of their effects is often extremely difficult. However, some perspective on these problems can be obtained from empirical analyses of existing National Geodetic Survey (NGS) releveling (Figure 1). Such an analysis forms

the basis of this report.

This paper reviews some factors that must be considered when attempting to extract tectonic information, especially those relevant to earthquake prediction, from historic releveling observations. Evidence on the extent and nature of systematic errors, non-tectonic movements, and tectonic deformation (both earthquake related deformation and tectonic movements unassociated with earthquakes) from U.S. releveling measurements is presented. Specific criteria to help recognize suspect movements are developed and illustrated by application to a reevaluation of certain southern California leveling results of particular interest in earthquake prediction.

Systematic Errors in Leveling

At the root of much of the current debate regarding leveling-derived estimates of crustal motion is the prevailing uncertainty as to the role of systematic measurement errors. In particular, systematic errors which accumulate with relief have become a central issue in crustal movement research . While errors of this type have been known to geodesists for some time (e.g., Bomford, 1971), their influence has been considered too small to be of concern in most geodetic applications. However, new field experiments carried out by the NGS (Whalen, 1980) and empirical analyses (Brown et al., 1980) confirm earlier suspicions (e.g., Savage and Church, 1974; Brown and Oliver, 1976; Citron and Brown, 1979; Jackson et al., 1980; Chi et al., 1980) that topography-induced systematic errors are larger and more common than heretofore established and consequently that such errors can be and probably have been misinterpreted as tectonic motions of the crust.

Topography-correlated errors can arise from improperly calibrated leveling rods and from unequal atmospheric refraction of the foresight and backsight readings. The effects of these two sources of error should differ in a number of respects and thus in principle may be distinguished. For example, ficticious movements resulting from rod calibration errors should correlate rather closely with detailed topography, and change magnitude only where rod pairs are changed. In contrast, atmospheric refraction will be independent of the rods used in the survey. Furthermore, refraction errors

can be expected to accumulate in a more complex manner because refraction depends on a variety of parameters which may vary significantly during the course of a given survey (e.g., near surface temperature gradients, individual sight lengths, wind, etc.). Because of procedural changes (a tendency towards shorter sight lengths in newer surveys), atmospheric refraction should more often than not result in ficticious movements which show a positive correlation with topography (i.e., high areas will appear to be rising) while errors due to rod miscalibration should have no preference for positive or negative correlations. In practice, these distinctions are not always easy to draw. However, a preliminary survey of NGS releveling estimates of elevation change which correlate with topography indicates that about 75% display positive correlations. While rod calibration errors appear to effect some leveling observations (e.g. Jackson et al., 1980; this paper) this preliminary result suggests that refraction may be the more pervasive source of elevation correlated error.

The expected magnitude for refraction error is a point of considerable uncertainty. According to one approximation (i.e., Kukkamaki, 1938), this error is proportional to the height difference between benchmarks, the temperature difference between 0.5 m and 2.5 m above the ground, and the square of the sight length used in the observation. Figure 2 shows representative values for refraction error using constants given by Holdahl (1980). For reasonable temperature differences (1-2 degrees C: Whalen, 1980) and sight lengths (25-75 m), errors as large as 30-40 mm or more can easily accumulate over height differences of 100 m (300-400 ppm). Since refraction error will usually have the same sign, its effect should tend to cancel when differencing surveys to compute movement. This rationale has often been used as an argument for ignoring the effect. However, if surveys are conducted using different sight lengths (Strange, 1981) and/or under different micrometeorological conditions, the refraction effect will result in what appear to be movements that roughly correlate with relief.

Examples of apparent movement correlating with elevation are numerous (Brown et al., 1980). Approximately 20% of U.S. releveling observations show visual correlations between apparent movement and topography. The magnitude of the effect often reaches 30-40 mm per 100 m change in elevation. Although, correlation with topography

alone is insufficient to warrant rejection of a tectonic interpretation (e.g., see Reilinger et al., 1977), it is clearly grounds for suspicion particularly when correlations persist at short wavelengths (e.g., Figures 18 and 19) or when multiple surveys give inconsistent results. For example, Figure 3a shows apparent vertical movements and terrain along the route from Colorado Springs to Leadville, Colorado based on surveys conducted in 1925, 1953, and 1954. The reversal of the 1953-1925 apparent tilt during the period 1954-1953 strongly suggests systematic error, and the correlation with terrain (Figure 3b) points toward an elevation correlated error. Since elevation-dependent errors may contaminate a significant portion of the NGS releveling data base, the possibility of such errors must always be considered prior to invoking tectonic explanations for apparent movements which correlate with relief.

Topography dependent error is not the only type of systematic error affecting U.S. leveling measurements. Table 1 lists a number of areas where comparison of repeated leveling measurements show large systematic discrepancies which may be due to errors in the observations. These examples occur in areas of generally subdued relief, thus ruling out elevation-correlated errors. If leveling errors are solely responsible for these discrepancies, whatever their cause, they range in magnitude from .3 mm/km to 1 mm/km and remain systematic (i.e., accumulate monotonically) for distances ranging from about 100 km to over 600 km.

Figure 4 shows three different estimates of elevation change (assuming constant rates of movement over the time period between levelings) along the east coast of the U.S. from Maine to Florida: 1) from unadjusted leveling measurements; 2) from tide gauge records (squares); and 3) from the same leveling observations adjusted with standard least squares procedures for consistency with other repeated leveling lines which form circuits extending inland from the coast (tide gauge data were not used in the adjustment, Jurkowski et al., 1979). The leveling measurements span an approximately 30 yr. time interval. The fact that the relative movement between Maine and Florida is substantially reduced through the adjustment and the fact that the adjusted leveling profile is more consistent with tide gauge data (although serious discrepancies still remain) indicate that the regional north-south tilt results from

systematic errors in the leveling observations. The error remains more or less systematic over distances of 1000's of kilometers, and on some sections (e.g., 1800-2600 km) reaches .5 mm/km.

Comparison of leveling and tide gauge estimates of crustal movement along the west coast of the U.S. between Astoria, Oregon, and Crescent City, California show similar discrepancies (Brown et al., 1980). Unlike the east coast profile, apparent crustal movements along the west coast were derived from only two surveys and were thus not subject to possible temporal bias due to stringing together segments covering different time intervals. For the west coast profile, the north-south error reaches .3 mm/km and remains systematic over a distance of 380 km.

The cause of the apparent errors in these coastal surveys is presently unknown. The predominantly north-south orientation of the coastal profiles may suggest unequal lighting or other factors which are believed to preferentially accumulate on north-south lines (Bomford, 1971). However, the substantial reduction of the apparent tilt indicated by the east coast profile when adjusted with inland data suggests that the error may be related to the proximity of the leveling route to the coast (i.e., the error did not effect, or had less of an effect, on profiles further inland).

Suspect movements are not restricted to coastal and north-south profiles. For example, consider the large apparent tilt across the U.S. midcontinent identified by Brown and Oliver (1976) from releveling between Davis Junction, Illinois and Willard, Ohio (Figure 5). This tilt is perhaps the largest apparent movement defined by leveling in the eastern U.S. The tilt anomaly shows no clear relationship to geologic structure and is inconsistent with movements inferred from comparisons of water level gauges in the great lakes (Brown and Oliver, 1976). Figure 5 shows the results of a loop closure analysis (see Chi et al., 1980 for discussion of method) for circuits including the Davis Junction to Willard route. The fact that misclosures are considerably larger when the circuits are closed with the 1967-1969 surveys between Davis Junction and Willard than with the 1930-1947 surveys, even though the remainder of the loop was surveyed more closely in time to the 1967-1969 interval, suggests that the large apparent tilt of the interior plains is due to systematic error and not real ground motion. Such an error would have to reach 1 mm/km and remain systematic for

distances of well over 500 km.

In both the coastal and interior examples cited above, the discrepancies are characterized by consistent accumulation over large distances. The effects of these long baseline discrepancies are often reduced below the level of concern for most geodetic applications by network adjustments. However, when examined in the context of evaluating crustal movements such discrepancies become quite significant. Although the net apparent movement over a profile can be large, the tilt rates resulting from these errors are rather low, especially when compared with those exhibited by unequivocal examples of real movement. Tilt may therefore be the more diagnostic parameter in evaluating reliability of crustal movement estimates.

Near-Surface Movements

In addition to systematic errors, releveling measurements are influenced by near-surface movements which can mask, or be mistaken for deep-seated tectonic motion. Near-surface effects include: a) Benchmark instability (frost heave, soil moisture and temperature changes, human disturbance), b) Surface failure (land slides, mine and cavern collapse), c) Loading (reservoir impoundment, building settlement), and d) Fluid withdrawal (water, oil, gas). Benchmark instability and surface failure are often easily identified or are of such local extent that they are not a serious problem for regional tectonic studies. Such effects can, however, complicate local investigations - for example, of movements near earthquake faults. Near-surface soil or sediment compaction due to earthquake ground-shaking may be responsible for the predominance of subsidence over uplift near many earthquake faults (Savage and Hastie, 1966). Subsidence due to surface loading and fluid withdrawal is, in general, easily related to human activity. In fact, leveling has proven quite effective at monitoring such movements, with important engineering applications (e.g., Poland and Davis, 1969). However, near-surface movements, and in particular movements due to variations of water levels in aquifers, appear to be more widespread than previously reported. In addition, such effects can be subtle and subsequently misidentified as tectonic deformation.

Figure 6 shows NGS releveling profiles in the

U.S. which indicate subsidence relative to surrounding areas and which overlie aquifer systems which have experienced variations in water levels due either to pumping or natural causes. These movements may therefore represent sediment compaction associated with these water level variations, and are consequently suspect as indicators of tectonic motion. It is interesting to note from Figure 6 that these apparently near-surface movements are quite common in southern California, affecting much of the area peripheral to the Palmdale Bulge. Arguments will be presented later suggesting that in at least one case such groundwater subsidence in southern California has been misidentified as tectonic uplift of the adjacent areas.

A particular example, not previously reported, which illustrates criteria which can be used to recognize near-surface sediment compaction is the relative subsidence of the Los Angeles basin. Figure 7 gives a map of the L.A. Basin showing contours of basement depth and the location of a leveling route traversing the basin. Also shown are elevation changes along the leveling route and the history of water level decline measured in an observation well near the center of the basin. Subsidence near the center of the basin reaches 15 cm relative to the periphery and extends over a distance of 40 km. The observed subsidence correlates spatially with aquifer geometry and temporally with the history of water level decline. In addition, the magnitude of the effect (i.e., the ratio of subsidence to water level decline) is comparable to observations in other areas (Poland and Davis, 1969). Had the relationship between aquifer geometry and subsidence not been noticed, it is possible that these measurements could have been misinterpreted as tectonic motion.

Vertical Movements and Earthquakes

In spite of the substantial difficulties associated with releveling estimates of crustal movement, some of which have been described in previous sections, the capability of the leveling technique for monitoring tectonic earth movements is well established. In a number of cases, relatively subtle earth movements (i.e., tilts few x 10^{-6} rad and tilt rates few x 10^{-8} rad/yr) have been identified. In this section we briefly review releveling evidence for earthquake related

deformation in the U.S. and use specific examples to illustrate some of the criteria employed to identify real tectonic movements.

The best examples of tectonic deformations measured by leveling in the U.S. are those in the vicinity of major earthquakes. Figure 1 and Table 2 review those U.S. earthquakes for which vertical movements have been reported. All of these earthquakes are associated with faults that have a significant component of dip-slip movement (with the possible exception of the 1940 Imperial Valley earthquake). Up to the present, there is no clear evidence from U.S. releveling measurements for permanent vertical deformation associated with purely strike-slip faulting although present observations are not sufficient to rule such out.

The most obvious vertical movements are those accompanying the earthquake (coseismic). Coseismic deformation has been well-documented for several of the larger ($M > 6$) normal and thrust earthquakes which have occurred in areas of preexisting geodetic control (Table 2). Observed movements range in magnitude from a few cm to a few m depending on the size of the earthquake and the proximity of the leveling measurements to the epicentral area. In general, coseismic movements are well explained by elastic dislocation theory (Savage and Hastie, 1966) although complications can arise from such factors as near-surface soil or sediment compaction due to ground shaking.

Post-seismic vertical movements have also been observed by releveling for some of the larger dip-slip earthquakes (see Table 2). These movements are usually smaller than associated coseismic movements; however like coseismic movements they can often be identified by their close spatial and temporal association with earthquakes, and in some cases surface faulting. Where sufficient observations exist, post-seismic deformation rates appear to decrease exponentially from the time of the earthquake. For example, movements near Anchorage following the 1964 Alaska earthquake are shown in Figure 8 (Brown et al., 1977). The Alaska earthquake, one of the largest events ever recorded, occurred where the oceanic Pacific plate is being thrust under the continental North American plate at a rate of over 5 cm/yr (Plafker, 1972). Savage and Hastie (1966) and Hastie and Savage (1970), using a dislocation model of thrust faulting, showed that the coseismic displacements were consistent with low-angle thrusting. Post-seismic movements

near Anchorage (Figure 8) amounted to as much as 0.55 m of land uplift at an exponentially decreasing rate during the decade following the earthquake. Additional evidence for deformation following the Alaska earthquake was reported by Prescott and Lisowski (1977) from analysis of detailed leveling arrays on Middleton Island in the Gulf of Alaska. Tilts associated with the Alaska post-seismic movements were on the order of 10^{-5} to 10^{-6} rad. There is still considerable debate as to the mechanism responsible for post-seismic movements, but at least some of the observations appear consistent with after-slip on the fault, or an extension of the fault that ruptured during the earthquake, although other explanations have been proposed (e.g., Nur and Mavko, 1974; Wahr and Wyss, 1980).

While co-seismic and post-seismic movements are well established for at least some earthquakes, clear evidence for preseismic deformation from U.S. releveling measurements is quite rare. This may be due to a lack of appropriate measurements as opposed to the absence of such movements since it is unusual to have multiple levelings of sufficient proximity prior to an earthquake. Precursory vertical movements have been suggested from leveling measurements for only three U.S. earthquakes: the 1959 magnitude 7.1 Hebgen Lake, Montana, the 1971 magnitude 6.4 San Fernando, California and the 1973 magnitude 6.0 Point Mugu, California earthquakes (see Table 2). The evidence for preseismic movement near Point Mugu is marginal, both because the proposed movements are barely significant relative to random error estimates and because the area was subject to surficial subsidence due to groundwater withdrawal during the period of interest (Castle et al., 1977). The 1971 San Fernando earthquake is exceptional in that significant releveling was available for the epicentral area prior to the earthquake. These observations were analyzed after the earthquake and were interpreted to indicate precursory movements (Castle et al., 1974). However, reevaluation of the relevant leveling observations, described in a later section of this paper, cast some doubts on the reliability of these measurements and hence on their tectonic significance. Reilinger et al. (1977) found evidence for possible precursory uplift throughout a broad region surrounding the area of major co-seismic movement of the 1959 Hebgen Lake earthquake which apparently accumulated at a rate of 3-5 mm/yr (Figure 9). The zone of uplift is

defined by five independent elevation change profiles derived from 12 independent surveys. Although three of the five movement profiles show positive correlation with topography (i.e. high areas going up), one shows a negative correlation and one shows no correlation; yet all indicate a consistent sense of movement. This consistency argues strongly against elevation correlated errors as the cause of the observed uplift. The doming stands out distinctly in relation to movements in surrounding areas, and shows a close spatial correlation with the zone of major coseismic deformation and aftershock activity for the 1959 earthquake (Figure 9). In addition, the geodetically measured deformation is consistent in sign with Cenozoic deformation deduced from geologic structure (Reilinger et al., 1977). Tilts associated with this uplift range from 3 - 7 x 10^{-6} rad with associated tilt rates between 1 - 3 x 10^{-7} rad/yr. Although Reilinger et al. (1977) suggest that doming began prior to the earthquake, because of the limited number of pre-earthquake leveling measurements, it is impossible to prove that the activity was precursory (i.e., doming may have accompanied, and/or immediately followed the earthquake). In sum, therefore, leveling evidence for vertical movements preceding any U.S. earthquake is weak in both quantity and quality.

Other Tectonic Deformation

Recognizing real tectonic deformation from releveling, although necessary, is not sufficient grounds to infer that they are directly relevant to the earthquake prediction problem. Earthquake related movements must be separated from movements due to other deep seated processes, such as inelastic deformation (fault creep, folding; e.g. Thatcher, 1981), isostatic adjustments and magmatic activity. These mechanisms are believed, on the basis of observational evidence, to result in contemporary vertical movements which are sufficiently rapid to be detected by releveling measurements.

Movements due to subsurface magmatic activity, for example, are not restricted to volcanically active regions (e.g., Hawaii, Iceland, Japan), having been reported in Yellowstone National Park (Reilinger et al., 1977; Pelton and Smith, 1979), and the Rio Grande rift (Reilinger et al., 1980) as well. Crustal uplift in the Central Rio

Grande rift illustrates tectonic deformation which appears to be unrelated to major earthquake activity. The existence of an active magma body beneath the central Rio Grande rift was inferred primarily on the basis of geophysical, and some geological information (Sanford et al., 1977). The magma body is believed to consist of a thin sill at a depth of about 20 km (Figure 10). Elevation change profiles along the routes shown in Figure 10 are given in Figure 11. All three profiles indicate uplift of the area overlying the magma body. The observed uplift is believed to be due to tectonic deformation and not measurement errors or near-surface movements because: 1) uplift is defined by three independent elevation change profiles; 2) while the two east-west profiles show a rough negative correlation with topography near the area of uplift, the north-south profile shows no correlation, thus ruling out elevation-dependent errors as the primary cause of the observed movements; 3) the Belen to Amarillo profile demonstrates that the uplift of the rift is anomalous relative to points to the east; 4) geomorphic evidence for post-Pliocene deformation (Bachman and Mehnart, 1978) is consistent in sign with the geodetic observations; 5) anomalous uplift occurs directly above the magma body; and 6) modeling studies indicate that uplift could result from activity within the magma body. If uplift is accumulating more or less continuously as suggested (Reilinger et al., 1980), it is characterized by an average rate of 4 mm/yr, with corresponding tilt rates of 5 to 10 $\times 10^{-8}$ rad/yr. In spite of the rather compelling independent evidence for an active magma body beneath the area of uplift, it has been suggested that these movements may be due to an impending earthquake (Koseluk and Bischke, 1981).

Reliability Criteria

The selected cases described above demonstrate both the utility and limitations of geodetic leveling to detect tilts of a few x 10^{-6} rad and tilt rates of a few x 10^{-8} rad/yr. Thus while non-tectonic influences (e.g. systematic error) can obscure real earth movement, the technique has clearly proven effective at monitoring relatively subtle tectonic deformation. It is essential, however, that individual releveling observations be examined in detail for possible contamination by systematic errors and near surface movements prior to invoking tectonic explanations. Particularly effective quantitative techniques include comparison of forward and reverse levelings (e.g., Savage and Church, 1974) and loop closure analysis (e.g., Chi et al., 1980). In addition, the following qualitative criteria, some of which were illustrated by the previous examples, have proven useful for evaluating the reliability of particular data sets (Brown et al., 1980): 1) magnitude of apparent movements relative to possible errors (since many errors remain poorly understood this is equivalent to determining whether the movements in question stand out in relation to "background noise"); 2) consistent temporal behavior when multiple levelings are available (e.g., Alaska); 3) relations with independent geophysical or geologic estimates of recent movement (e.g., tide guage, lake levels, tilt meters, horizontal movements, geomorphic evidence, etc.); 4) consistent movements when multiple leveling lines cross a given feature (e.g., Hebgen Lake, Rio Grande rift); 5) correlation with geologic structure and tectonic activity (e.g., Hebgen Lake); 6) lack of correlation with topography ruling out possible elevation correlated errors; 7) lack of relationship to possible near-surface processes (e.g., fluid withdrawal, reservoir impoundment, etc.); 8) lack of relationship between apparent movements and procedural changes (changes in sight lengths, rod or instrument changes); and 9) consistency of inferred mechanism with tectonic setting (e.g., Alaska).

Although no single criteria is sufficient to warrant acceptance or rejection of a set of observations, when multiple criteria give consistent results they provide strong evidence for the reliability of the measurements in question.

A Case Study: Southern California Releveling Measurements

Much attention has recently been focused on leveling in southern California, where there is both considerable concern about future earthquakes and an abundance of leveling observations. Using the above reliability criteria, developed through analysis of the much broader data base of U.S. releveling, we have reevaluated some of the observations used to deduce preseismic movements for the 1971 San Fernando earthquake as well as the Palmdale Bulge. Our reevaluation, representing a different perspective, suggests that many of the southern California measurements are significantly affected by both topography-dependent errors and near-surface movements. On the other hand, at least some of the observations may still reflect tectonic deformation. Thus, the configuration of the Palmdale Bulge will, at the very least, require revision in light of improved understanding of those factors which can influence releveling measurements.

In our analysis of southern California releveling observations, data have been displayed in terms of relative movements for sequential time intervals along the pertinent segments of the leveling routes. This contrasts with previous attempts to tie the observations to a tide gauge in order to relate movements to sea level. Analyzing relative movements minimizes the effects of systematic errors, which can accumulate to rather substantial amounts over the 100-200 km distance to the tide gauge, and as will be demonstrated, greatly simplifies interpretation of the observations.

Figure 12 shows those leveling routes in Southern California for which crustal movement information has been investigated for this study. This information was used to define both the configuration of the Palmdale Bulge and preseismic movements for the 1971 San Fernando earthquake. Possible preseismic movements were reported by Castle et al. (1974) in the vicinity of the earthquake fault (segment I) and 30 km northwest of the epicenter (segment II). Strange (1981) subsequently reported evidence of preseismic movements just north and east of Saugus (along segment II and III). These observations were previously used to define the

Palmdale Bulge (Castle et al., 1976; Castle, 1973). Apparent movements along segment III and along the Llano to Azusa route (Figure 12) were also instrumental in determining the configuration of the Palmdale Bulge (Castle et al., 1976; Castle, 1978). Each of these features will be discussed separately below.

The sequence of movements along segment I crossing near the area of surface faulting are shown in Figure 13. Coseismic movement consisting of relative subsidence south of the San Fernando Fault and relative uplift north of the fault are clearly indicated for the 1969-1971 interval. These movements are roughly consistent with elastic rebound accompanying thrust faulting (Savage et al., 1975). Possible preseismic tilting up to the north is indicated by the profiles for the time intervals 1955-1961, 1961-1964, and 1964-1965. Apparently no tilt accumulated along this section between 1965-1969. Figure 14a shows relative movements between points near the ends of this profile segment plotted as a function of time. The temporal consistency of these movements is, in itself, normally evidence that the measurements reflect real movements (criteria 2). However, there are two reasons to suspect systematic error, and in particular refraction errors, rather than true ground motion.

Examination of the profiles in Figure 13 indicates that the observed tilting correlates with topography (criteria 6). This correlation, although suggestive, is not sufficient to confirm systematic error because real movements can in some cases correlate with relief (e.g., Reilinger et al., 1977). However, the sequence of apparent tilts between 1955 and 1969 show a systematic relationship to the sight lengths used for different surveys (Figure 14b); a relationship that is consistent with that expected from refraction errors (Holdahl, 1980) (criteria 8). The $3^{o}C$ temperature difference that results in a good fit to the observations, although somewhat higher than daily average temperature differences observed in other areas ($1-2^{o}C$, Whalen, 1980), may not be unreasonable for the spring and summer months in southern California (Holdahl, 1980 reports that temperature differences "may frequently attain values up to $4^{o}C$ shortly after noon during the summer"). Although Strange (1981) contends that significant tilting persists along this segment of the profile in spite of refraction corrections, the effectiveness of corrections made without on site temperature

measurements is not well established. In view of the possibility of refraction errors, the tectonic significance of the sequence of apparent tilts shown in Figure 13 remains ambiguous.

Figure 15 shows profiles of relative elevation change for sequential time intervals crossing the two areas of reported preseismic deformation northwest of the epicenter (Segment II; Figure 12). During the 1953-1964 interval the main deformation consisted of subsidence in the vicinity of Saugus relative to points farther north (ruled area of plot). This movement was originally attributed to the Palmdale Bulge (Castle et al., 1976; Castle, 1978) and more recently to preseismic effects of the San Fernando earthquake (Strange, 1981). However, analysis of releveling measurements throughout the Saugus Basin indicates that relative subsidence shows a close correlation with the geometry of the Saugus aquifer and the history of water level decline (Reilinger, 1980). The spatial correlation between the zone of subsidence and the Saugus aquifer is shown in Figure 16. In addition to the spatial and temporal correlation, the degree of subsidence of individual benchmarks within the aquifer is roughly proportional to the product of aquifer thickness and water level decline (Figure 17), the expected relationship for near surface compaction (e.g. Terzaghi and Peck, 1967). These observations strongly suggest that subsidence above the Saugus aquifer results from near-surface sediment compaction due to fluctuations of the water level and not from tectonic deformation (criteria 7). This result is particularly important to the current controversy surrounding the Palmdale Bulge (e.g. Stein, 1981) since, unlike many of the measurements defining the Bulge, those in the Saugus area do not correlate with relief.

The other large possible movements shown in Figure 15 occurred between 1965 and 1968 and between 1968 and 1969. These observations were the primary evidence used to infer pre-earthquake slip at depth on the earthquake fault (Thatcher, 1976). The 1965-1968 movements consisted of uplift of the north section relative to the south by about 6 cm. The 1968-1969 movements were, in essence, a reversal of the 1965-1968 movements (criteria 2). The important point is that both sets of apparent movements were dependent upon the 1968 survey. This is illustrated by the bottom-most plot in Figure 15, which shows the general absence of movement for the 1965-1969

interval. Therefore, either we were fortunate enough to catch preseismic deformation at a time of significant deflection (1968), and again when the movements had exactly reversed themselves (1969), or the 1968 survey was in error. While oscillatory movements may have occurred, the possibility of errors in the 1968 survey is at least as likely, particularly in light of the now suspect results south of the epicenter, and similarly suspect trends identified in leveling observations in other parts of the country.

The possibility that refraction errors contaminate leveling measurements south of the epicenter naturally raises the question as to whether this same effect is responsible for the apparent error in the 1968 survey northwest of the earthquake. The steep tilts indicated by the 1965-1968 and the 1968-1969 movement profiles occur where there is a corresponding steep slope in topography (25 to 40 km). However, the topographic slope is so steep (>.04 rad) that only short sight-lengths could be used, making it unlikely that atmospheric refraction coupled to sight-lengths was a significant effect. Unusual near-surface temperature differences at the time of the survey, or other elevation-correlated errors, such as miscalibrated leveling rods (Jackson et al., 1980) may have affected these observations (see also Stein, 1981).

Apparent movements along the route from Llano to Azusa, California were used to define the configuration of the Palmdale Bulge (Fig. 12). Elevation change profiles along this route derived from surveys in 1934, 1962, and 1971 are shown in Figure 18. The 1962-1971 profile indicates about 10 cm uplift at Llano relative to Azusa. However, the following observations strongly suggest that this apparent movement is due to systematic error, and in particular rod calibration error in the 1962 survey (note that these are different data than those reported by Jackson et al., 1980): 1) elevation changes along the segment of the 1962-1971 profile where most of the relative movement occurs (between A and B, Fig. 18) show a detailed correlation with topography along the leveling route. The relationship between apparent movement and terrain, which has a correlation coefficient of .98 is illustrated in Figure 19. Such a correlation with detailed topographic features is suggestive of systematic errors (criteria 6); 2) point A in Figure 18 where the correlation between movement and terrain ends rather abruptly is precisely the point where the leveling

instrument and rods were changed in the 1962 survey (criteria 8); 3) the elevation change profile derived from the 1934 and 1962 surveys along this route indicates movements in the opposite sense and of the same magnitude as the 1962-1971 profile. This is clearly illustrated by the 1934-1971 profile (Fig. 18) which shows no significant movement. This reversal in the sense of apparent movement casts serious doubts on the 1962 survey (criteria 2). These three points strongly suggest that apparent movements between Llano and Azusa, previously used to define the Palmdale Bulge, actually result from rod calibration errors in the 1962 survey.

Movements along leveling route III in Figure 12 were also important for determining the geometry of the Palmdale Bulge (Castle et al., 1976; Castle, 1978). In contrast to the previous examples, these apparent movements do not seem to be due to either systematic errors or near-surface effects and thus may represent tectonic deformation.

Figure 20 shows the sequence of relative movements along the survey route south of Palmdale. The major tilt event occurred between 1961 and 1964 and amounted to more than 10 cm of relative movement over a distance of 20 km. This corresponds to a tilt of 5×10^{-6} rad. The general absence of movements for the 1955-1961 interval, the 1964-1965 interval and the 1965-1971 interval attests to the reliability of all of these surveys (i.e., comparison of the 1955 or 1961 surveys with any of the later surveys will give roughly the same result). This implies that the 1961 to 1964 tilt event is in fact defined by five independent surveys. This is illustrated by the bottom most profile in Fig. 20 which shows significant tilting for the period 1961-1965 which, although reduced in amplitude, is quite similar to that shown by the 1961-1964 profile (criteria 2). In addition, this tilt anomaly does not show a strong correlation with topography (i.e., the direction of tilting does not reverse where the topographic slope reverses) (criteria 6). Furthermore, the sequence of relative movements show no relationship either to changes in leveling rods or to changes in sight lengths (criteria 8). This evidence suggests that the apparent tilting south of Palmdale reflects real crustal movements. Whether the tilt anomaly was a precursor to the San Fernando earthquake or represented a mechanically separate event is presently unknown.

Discussion and Conclusions

In spite of laudable progress in developing sophisticated new geodetic methods (e.g., VLBI, Laser Ranging, GPS) releveling measurements continue to be the most accurate (over appropriate distances) and widespread source of information on contemporary vertical movements of the continental crust. As such they constitute an important input to the earthquake prediction problem. Previous investigations, a few of which have been described here, clearly demonstrate the potential of the technique for monitoring subtle earth movements. However, it is equally clear that releveling estimates of crustal movement are influenced by near-surface movements and as yet poorly understood systematic errors which can obscure or be mistaken for tectonic deformation. Thus, uncritical interpretation of releveling observations can lead to erroneous tectonic conclusions, which in the case of earthquake prediction could entail serious social ramifications. The checking techniques (e.g., circuit closure analysis) and reliability criteria illustrated in this study, represent an attempt to quantify specific procedures for evaluating the tectonic significance of particular leveling data sets. Although not foolproof, these procedures have proven effective in a number of cases at discriminating tectonic movements from suspect effects. However, even when spurious effects can be eliminated, relating observed deformation to preseismic mechanisms may be quite difficult because of our limited understanding of precursory phenomena and our general inability to distinguish them from vertical movements due to other causes (e.g., magmatic activity, isostatic movements, aseismic orogenic deformation, etc.). Furthermore, the sparse distribution of leveling surveys in both space and time, even in areas like southern California, makes it highly unlikely that precursory movements for all but the largest earthquakes will ever be detected. In order for leveling to become more than an accidental contributor to earthquake prediction, a systematic leveling program designed for geodynamic rather than geodetic objectives is needed to develop the observational background required to recognize possible preseismic movement.

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FIGURE CAPTIONS

Figure 1. Leveling routes for which crustal movement information has been obtained in the U.S. Locations of U.S. earthquakes for which releveling evidence of crustal movement has been reported are also shown. Numbers refer to Table 2.

Figure 2. Magnitude of refraction error (R_E) normalized by height difference (ΔH) versus sight length (L) for various temperature differences. Based on relationship and theoretically determined constant given by Holdahl (1980).

Figure 3. a) Profiles of apparent elevation change and topography from Colorado Springs to Leadville, Colorado. Reversal of apparent tilt and correlation with topography strongly suggest elevation correlated error.
b) Apparent elevation change versus elevation difference for 1954-1953 profile in Figure 3a. Correlation coefficient (r) and regression slope (magnitude of error) are also shown.

Figure 4. Elevation change profiles along east coast of U.S. (map at right). Unadjusted profile based on observed elevations from leveling assuming constant velocity movement (modified from Brown, 1978). Adjusted profile is same data adjusted by other levelings lying inland from coast (tide gauge data not used in adjustment). Squares show similar profile derived from tide gauges.

Figure 6. Location of elevation change profiles which indicate subsidence possibly due to groundwater effects. Shaded areas represent previously published cases of near surface subsidence.

Figure 7. Map of Los Angeles basin. Contours indicate depth to basement (M), heavy dashed lines are faults, stippled areas are bedrock outcrops, heavy solid line (San Pedro to Los Angeles) is leveling route - crustal movements for period 1955-1964, indicating sediment compaction, shown below map. Asterisk shows location of observation well for which water level history is shown at

right.

Figure 9. Contours of elevation change (1 mm/yr and 5 mm/yr contours shown) for doming of Hebgen Lake region. Movements may reflect preseismic movements for 1959 Hebgen Lake earthquake (Reilinger et al., 1977).

Figure 10. Locations of leveling routes and benchmarks (dots) in Socorro-Albuquerque, New Mexico area. Outline of mid-crustal magma body is also shown (from Rinehart et al., 1979).

Figure 12. Leveling routes in Southern California for which crustal movements have been investigated in this study. 1971 San Fernando earthquake epicenter (*) and surface fault are shown along with contours of Palmdale Bulge as reported by Castle (1978).

Figure 13. Relative movements for sequential time intervals and topography for route I in Figure 12.

Figure 15. Relative movement for sequential time intervals and topography for route II in Figure 12.

Figure 16. (Top) Leveling routes traversing the Saugus aquifer.
(Bottom) Relative movements along routes shown at top. Note correlation between subsidences and aquifer location. Section labelled S.C.R. occurs within the alluvial aquifer of the Santa Clara river and may be subject to subsidence due to compaction of this aquifer (Mark et al., 1981). Section labelled REF may reflect refraction error (Figure 14b and text).

Figure 17. Plot of subsidence versus change in potentiometric surface (i.e., water level: ΔP) times aquifer thickness (T) for benchmarks in and immediately adjacent to aquifer (from Reilinger, 1980). Different symbols refer to different releveled segments: circles-Saugus to North (1953-1964); squares-Saugus to South (1955-1964); triangles-Saugus to East (1955-1961). The three circled points lie in the southeastern part of the aquifer and may reflect either different sediment characteristics or some effect other than

sediment compaction in this area.

Figure 18. Profiles of apparent elevation change from Llano to Azusa, California used to define the Palmdale Bulge (see Figure 12 for location).

Figure 19. Correlation between apparent movement and terrain for segment (A to B) of 1962-1971 profile shown in Figure 18. Correlation coefficient (.98) and slope of regression line are also given. Correlation and reversal of apparent movement strongly suggest elevation-dependent systematic error.

Figure 20. Relative movement for sequential time intervals and topography for route III in Figure 12.

Figure 5. Leveling loops used to investigate apparent tilting between Davis Junction, Illinois and Willard, Ohio. Elevation change profile and topography along Davis Junction to Willard route shown at right. Misclosures for each loop, given in brackets, suggest systematic error in 1967-1969 leveling from Davis Jct. to Willard, Ohio.

Figure 8. Elevation changes and topography between Anchorage and Whittier following 1964 Alaska earthquake (modified from Brown et al., 1977). Profiles are tied to sea level at Anchorage. Elevation change versus time for benchmark near center of uplift is shown at right. Note exponentially decreasing uplift.

Figure 11. Profiles of elevation change and topography used to infer uplift above Socorro magma body. Dates of leveling are indicated at the top of each plot.

Figure 14. a) Relative movement between benchmarks near ends of profile shown in Figure 13 plotted versus time.
b) Relative movement [$\Delta(\Delta H)$] indicated by profile in Figure 13 normalized by height difference between these points (ΔH) for different time intervals plotted versus difference in square of sight length for corresponding leveling surveys. The straight lines represent the expected relationship from refraction error for a range of temperature differences (from relationship given by Holdahl, 1980).

Table 1. Magnitude of systematic discrepancies between repeated levelings (apparently due to leveling errors) in areas of subdued relief. Discrepancies are given in terms of apparent tilt (mm/km) and apparent tilt rate ($10^{-8}yr^{-1}$). Distance over which discrepancies accumulate monotonically and average relief along route are shown.

AREA	MAGNITUDE (MM/KM)	MAGNITUDE (10^{-8} YR^{-1})	DISTANCE (KM)	AVERAGE RELIEF (M)
East Coast U.S.	0.5	1.6	650	<5
West Coast U.S.	0.3	3.3	380	<10
Midcontinent U.S.	1.0	2.6	140	<100

Quoted Limit for Systematic Error in a Single Survey <0.2 mm/km (Bomford, 1971)

Table 2. Earthquakes for Which Vertical Movements Derived from Leveling have been Reported. Maximum Observed Deformation and Typical Dimension of Effected Area are also Given. Numbers refer to Figure 1.

	Earthquake	Movement (cm)	Dimensions (km)	Comments	Reference
1	1931 Valentine, Texas Mw6.1 coseismic	10	30	Observed movements could include possible preseismic and/or postseismic deformation	Ni et al., 1980
2	1933 Long Beach, California; Mw6.3 coseismic	20	15	Complicated by effects of fluid withdrawal	Parkin, 1958
3	1940 Imperial Valley, California; Mw 7.1 coseismic	20	25	Complicated by possible near-surface compaction	Miller et al., 1970
4	1952 Kern County, California; Mw7.7 coseismic	60	30		Lofgren, 1966
5	1954 Western Nevada series of earthquakes; Mw6.6 to Mw7.1				
	coseismic	100	35		Whitten, 1957
	postseismic	7	8		Savage and Church, 1974
6	1959 Hebgen Lake, Montana; Mw7.1			Movements interpreted as preseismic could include coseismic and postseismic effects	
	preseismic	20	100		Reilinger et al., 1977
	coseismic	700	35		Myers and Hamilton, 1964

Table 2 (cont'd.)

	Earthquake	Movement (cm)	Dimensions (km)	Comments	Reference
7	1964 Alaska; Mw8.4				
	coseismic	600	400		Plafker, 1969
	postseismic	55	100		Brown et al., 1977
8	1971 San Fernando, California; Mw6.4				Prescott and Liskowski, 1977
	preseismic	20	30	Preseismic movements questionable - see text	
	coseismic	230	30		Castle et al., 1974
	postseismic	6	10		Burford et al., 1971
					Savage and Church, 1975
9	1973 Point Mugu, California; Mw6.0				
	preseismic	4	50	Complicated by subsidence due to groundwater effects	
	coseismic	3	50		Castle et al., 1977
10	1975 Oroville, California; Mw5.7				
	coseismic	18	20		
	postseismic	3	20		Savage et al., 1977
11	1975 Yellowstone, Wyoming; Mw6.0				
	coseismic	6	8	Complicated by possible postseismic movements of 1959 Hebgen Lake earthquake	Pitt et al., 1979

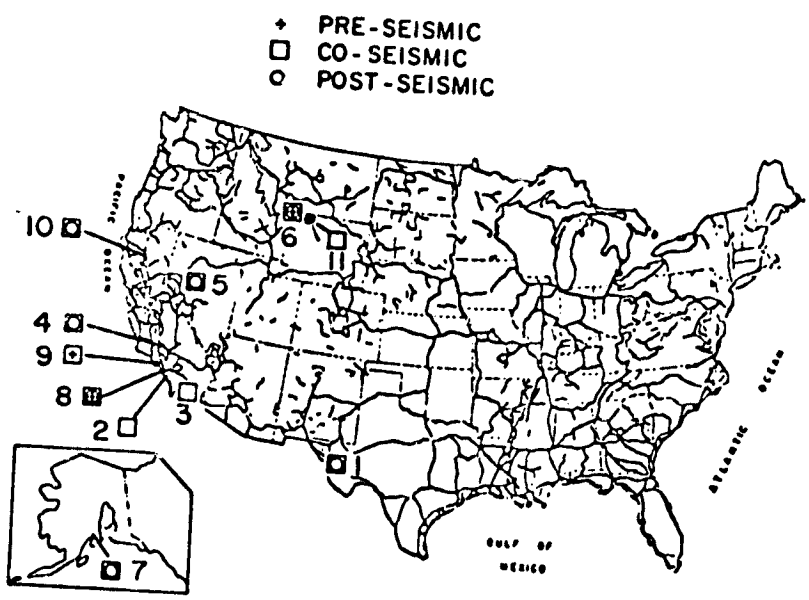


Fig 1

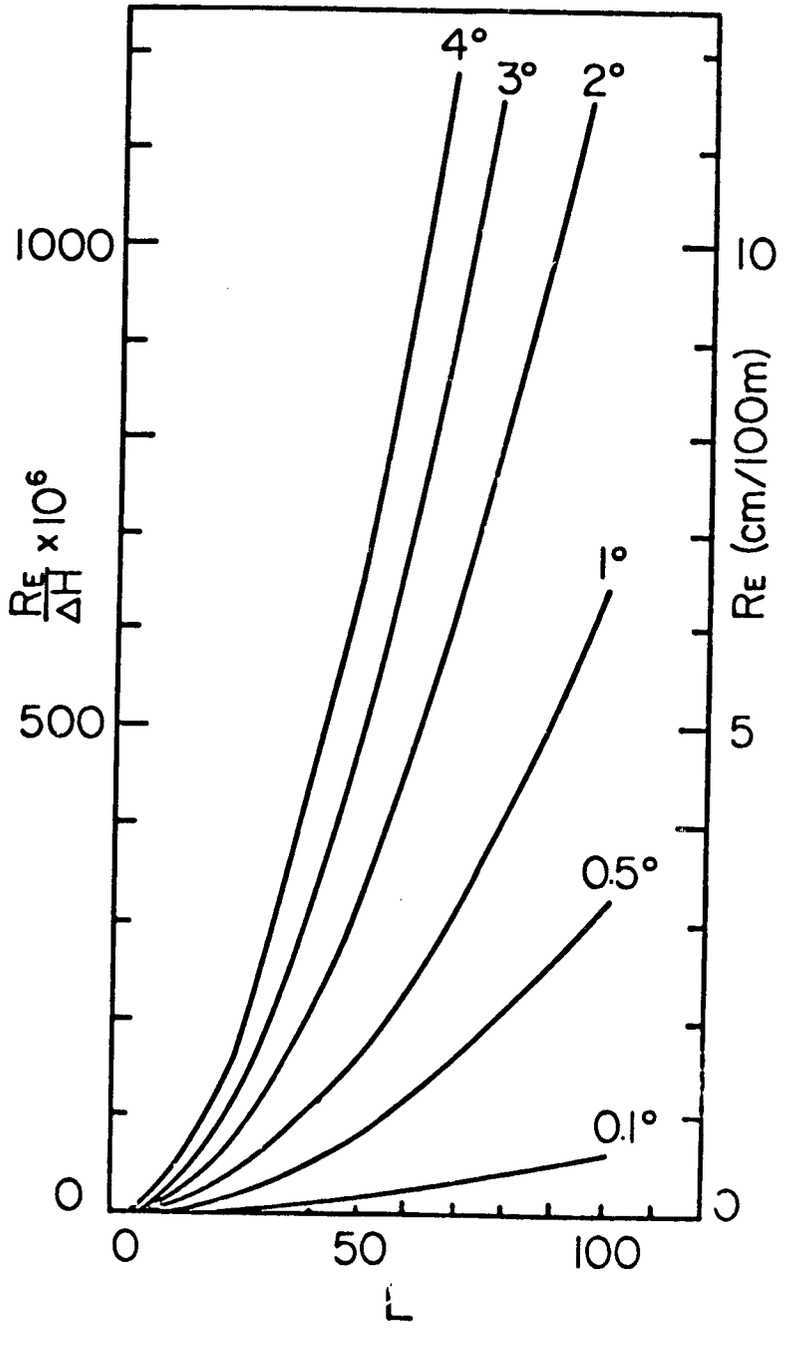

4°
3°
2°
1°
0.5°
0.1°
1000
500
0
10
5
0
0
50
100
L
$\frac{R_E}{\Delta H} \times 10^6$
R_E (cm/100m)

Fig 2

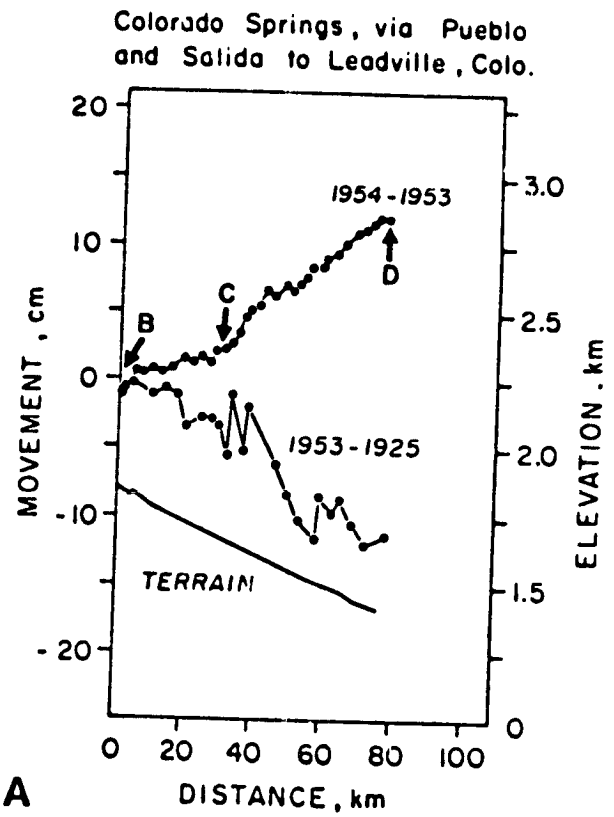
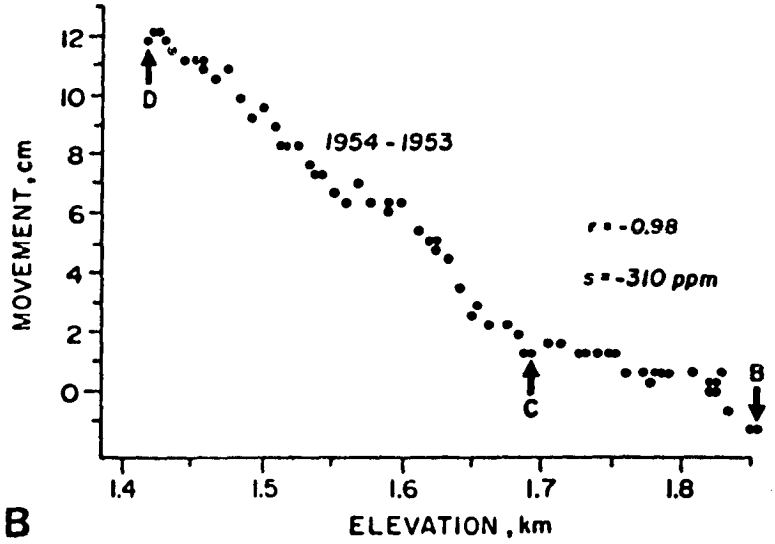

Colorado Springs, via Pueblo and Salida to Leadville, Colo.
20
10
0
-10
-20
MOVEMENT, cm
1954-1953
1953-1925
TERRAIN
B
C
D
3.0
2.5
2.0
1.5
0
ELEVATION, km
0
20
40
60
80
100
DISTANCE, km
A

Fig 3A


Colorado Springs, via Pueblo and Salida to Leadville, Colorado
MOVEMENT, cm
12
10
8
6
4
2
0
D
C
B
1954 - 1953
r = -0.98
s = -310 ppm
1.4
1.5
1.6
1.7
1.8
ELEVATION, km

B

3B

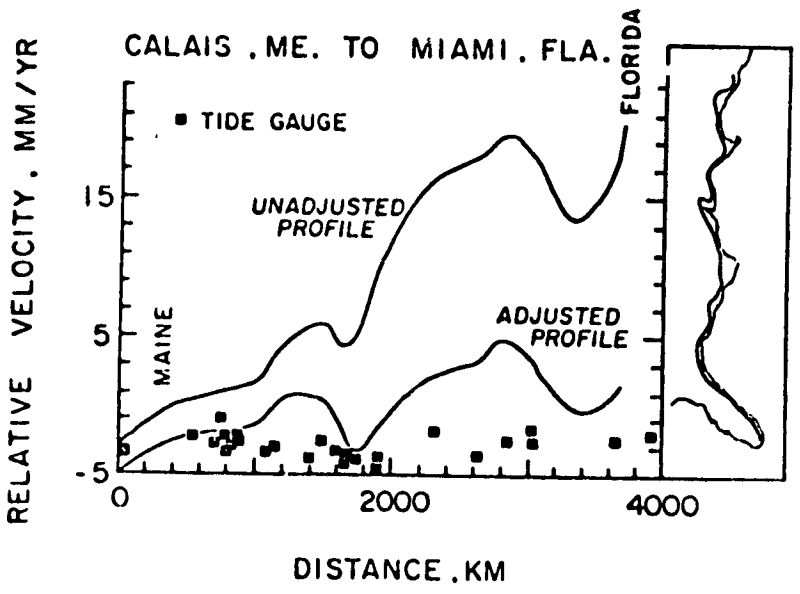
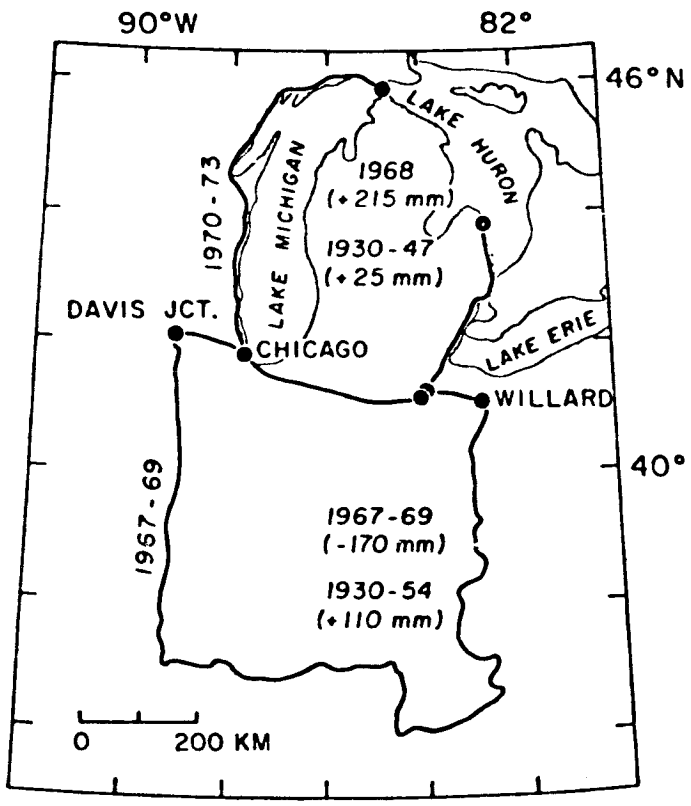

CALAIS . ME. TO MIAMI . FLA.
TIDE GAUGE
UNADJUSTED PROFILE
ADJUSTED PROFILE
MAINE
FLORIDA
RELATIVE VELOCITY . MM/YR
15
5
-5
0
2000
4000
DISTANCE . KM

Fig 4


90°W
82°
46°N
40°
LAKE HURON
LAKE MICHIGAN
LAKE ERIE
1968
(+215 mm)
1930-47
(+25 mm)
1970-73
DAVIS JCT.
CHICAGO
WILLARD
1957-69
1967-69
(-170 mm)
1930-54
(+110 mm)
0
200 KM

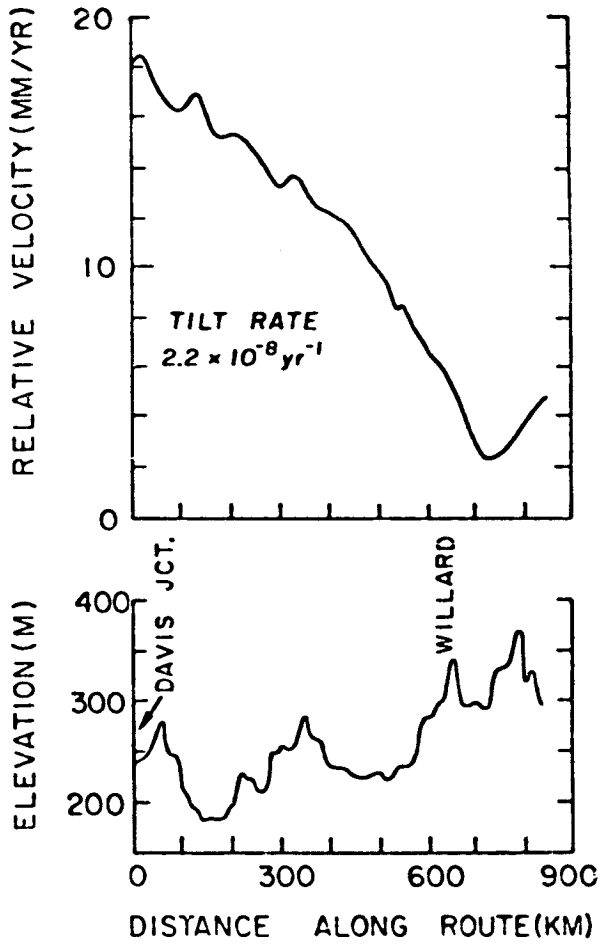

RELATIVE VELOCITY (MM/YR)
20
10
0
TILT RATE
2.2 × 10⁻⁸ yr⁻¹
ELEVATION (M)
400
300
200
DAVIS JCT.
WILLARD
0
300
600
900
DISTANCE ALONG ROUTE (KM)

Fig 5

Fig 6

ORIGINAL PAGE IS
OF POOR QUALITY

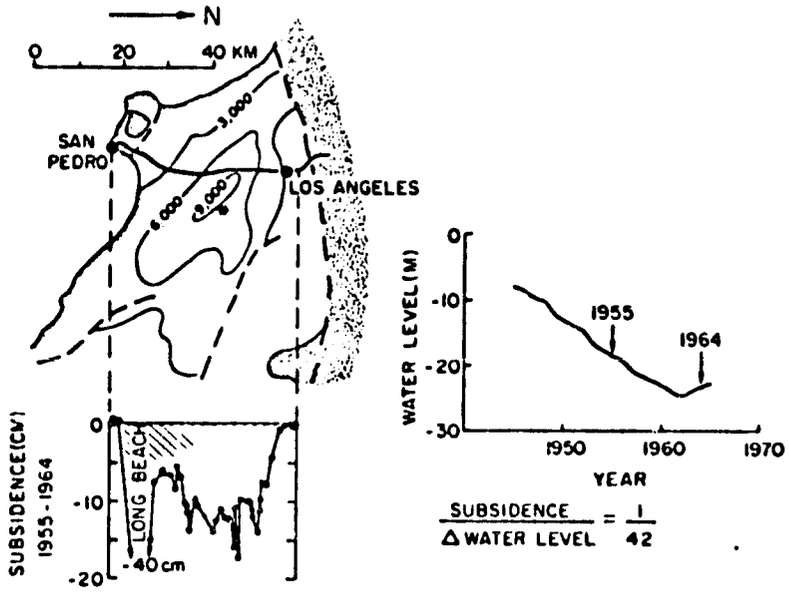

N
0 20 40 KM
SAN PEDRO
LOS ANGELES
3,000
6,000
9,000
SUBSIDENCE(CM)
1955-1964
0
-10
-20
LONG BEACH
-40 cm
WATER LEVEL(M)
0
-10
-20
-30
1955
1964
1950
1960
1970
YEAR
SUBSIDENCE / Δ WATER LEVEL = 1/42

Fig 7

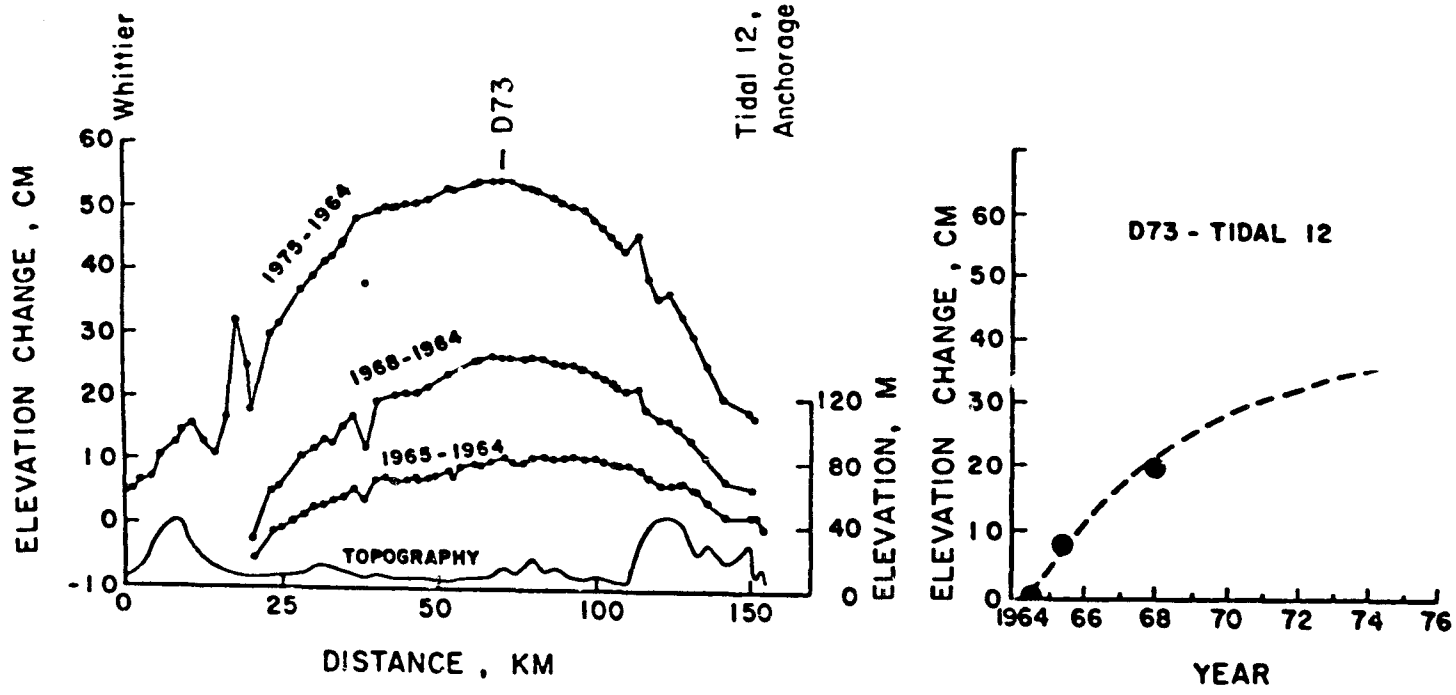

Whittier
D73
Tidal 12, Anchorage
ELEVATION CHANGE , CM
60
50
40
30
20
10
0
-10
1973-1964
1968-1964
1965-1964
TOPOGRAPHY
0
25
50
100
150
DISTANCE , KM
120
80
40
0
ELEVATION, M
D73 - TIDAL 12
ELEVATION CHANGE , CM
60
50
40
30
20
10
0
1964
66
68
70
72
74
76
YEAR

Fig 8

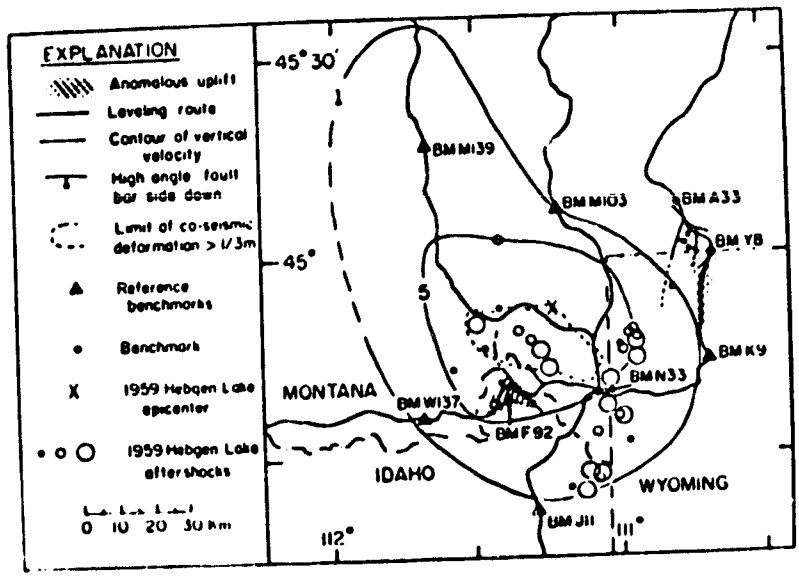

EXPLANATION
Anomalous uplift
Leveling route
Contour of vertical velocity
High angle fault bar side down
Limit of co-seismic deformation > 1/3m
Reference benchmarks
Benchmark
1959 Hebgen Lake epicenter
1959 Hebgen Lake aftershocks
0 10 20 30 km
45°30'
45°
112°
111°
BM MI39
BM MI03
BM A33
BM Y8
BM K9
BM N33
BM W137
BM F92
BM J11
MONTANA
IDAHO
WYOMING
5

Fig 9

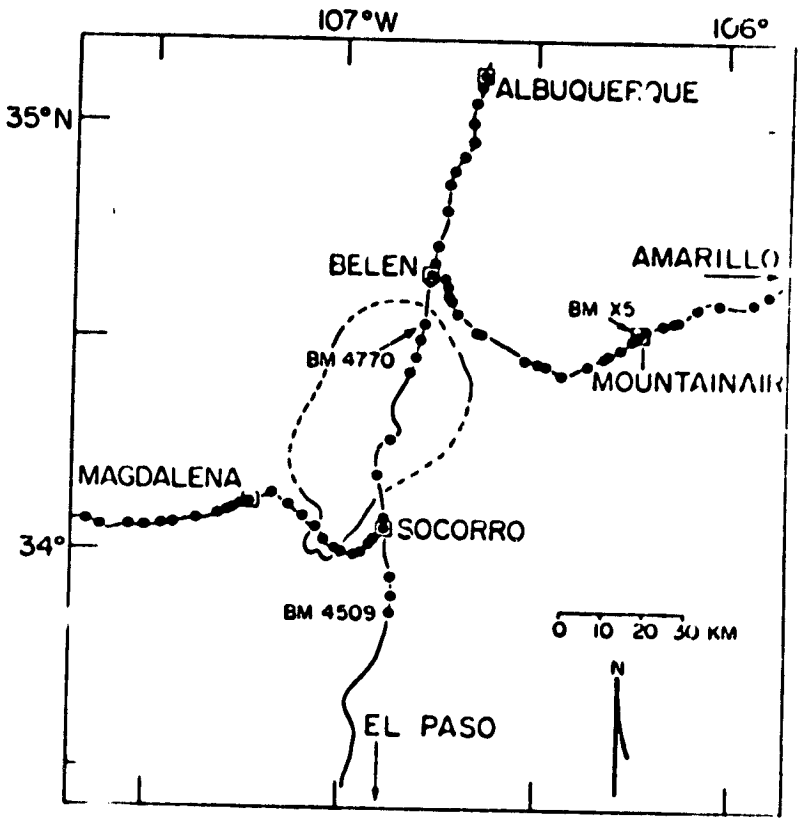

107°W
106°
35°N
34°
ALBUQUERQUE
BELEN
AMARILLO
BM X5
BM 4770
MOUNTAINAIR
MAGDALENA
SOCORRO
BM 4509
EL PASO
0 10 20 30 KM
N

Fig 10

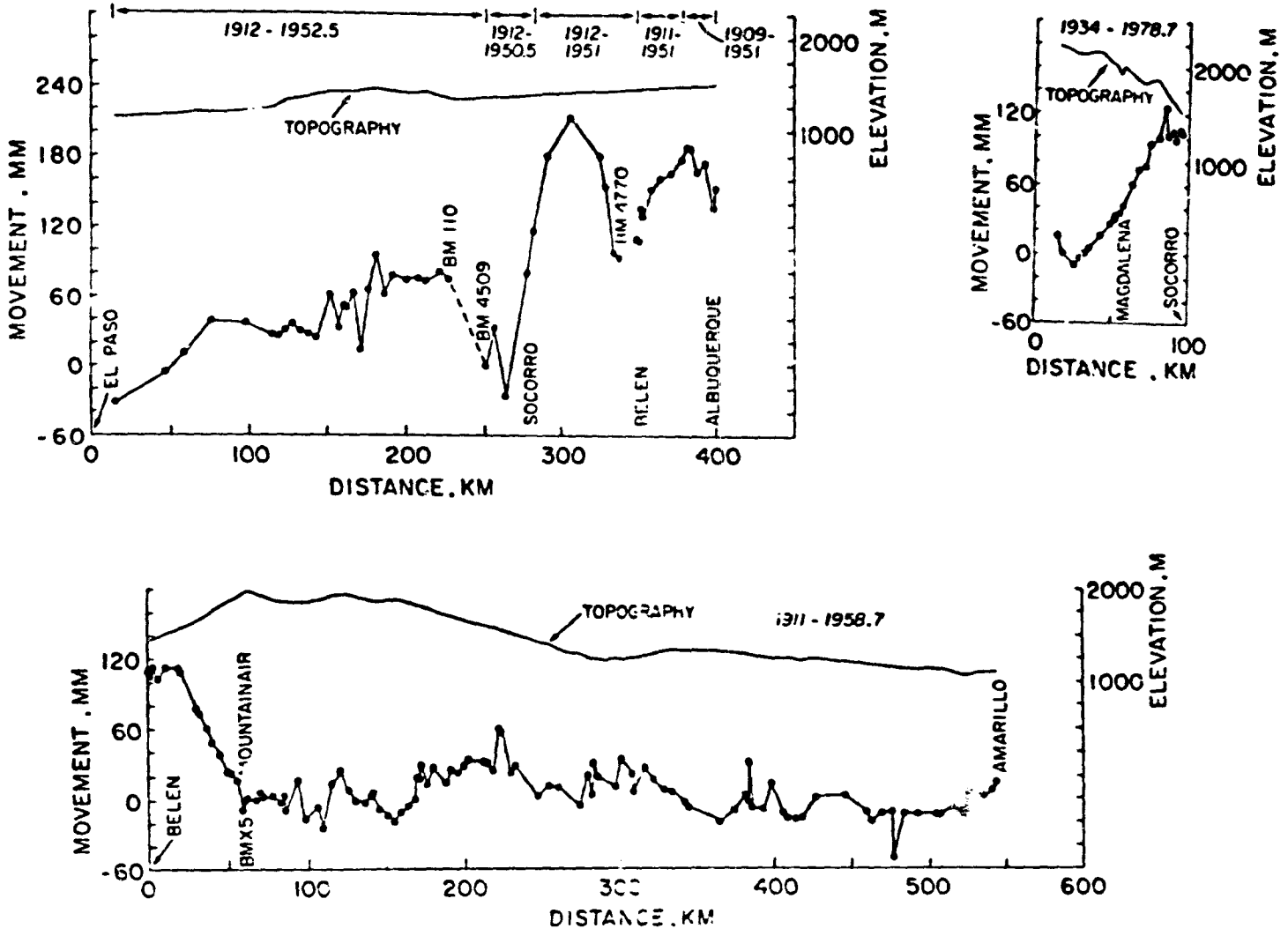

1912 - 1952.5
1912-1950.5
1912-1951
1911-1951
1909-1951
TOPOGRAPHY
MOVEMENT . MM
ELEVATION . M
DISTANCE . KM
EL PASO
BM 110
BM 4509
SOCORRO
BELEN
ALBUQUERQUE
1934 - 1978.7
MAGDALENA
SOCORRO
1911 - 1958.7
BELEN
AMARILLO

Fig 11

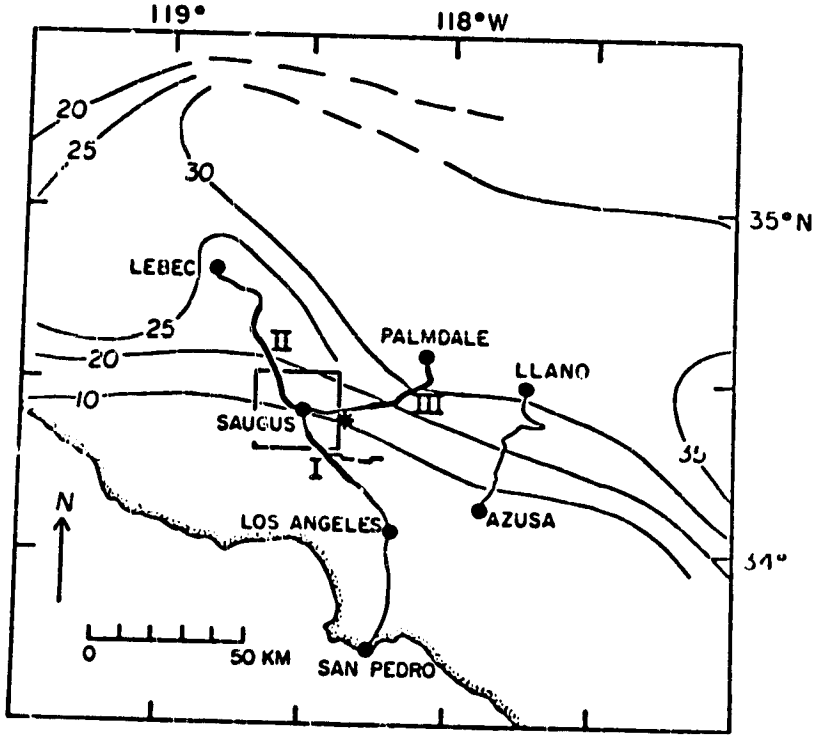

119°
118°W
35°N
34°
20
25
30
LEBEC
25
20
10
II
PALMDALE
LLANO
III
SAUGUS
I
N
LOS ANGELES
AZUSA
0
50 KM
SAN PEDRO

Fig 12

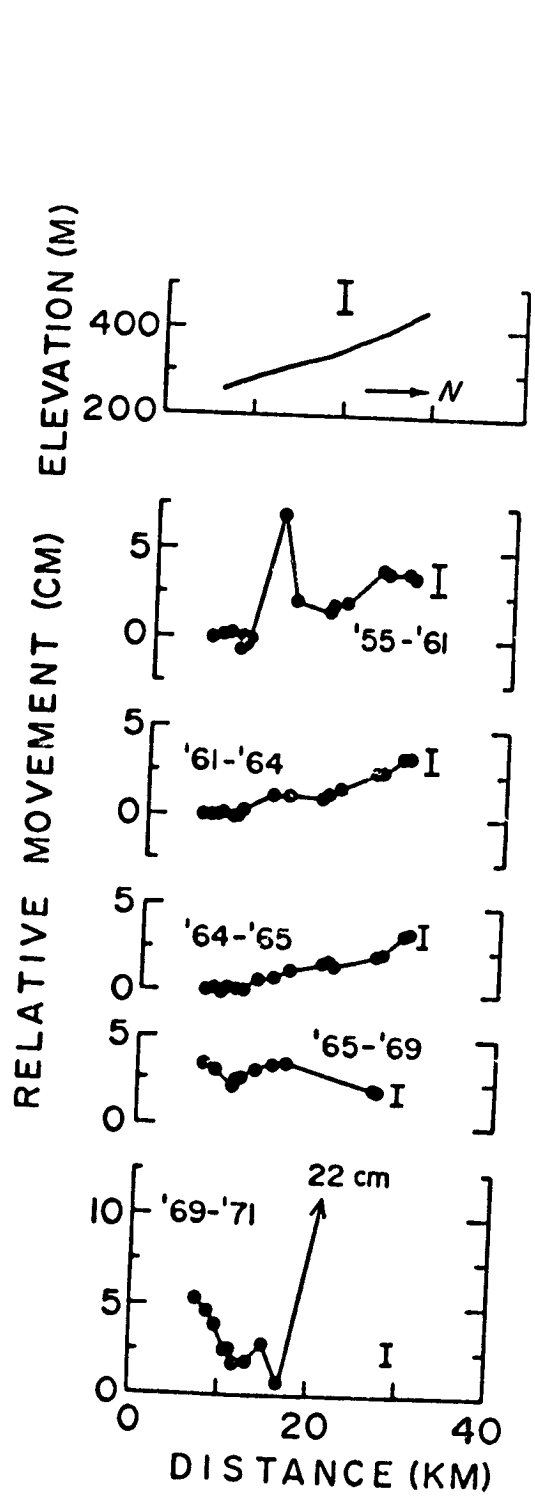
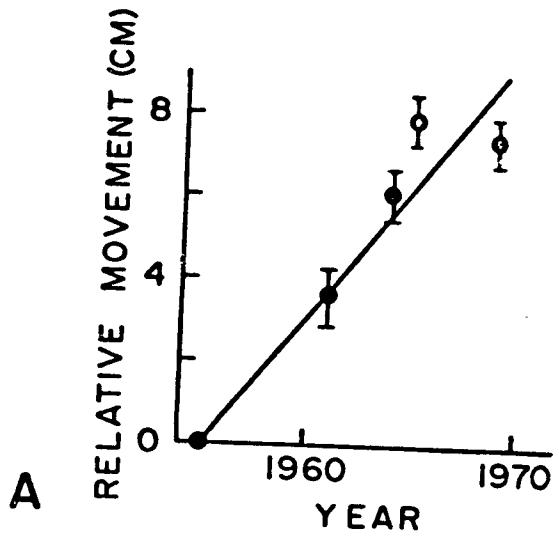

ELEVATION (M)
400
200
I
N
RELATIVE MOVEMENT (CM)
5
0
'55-'61
I
'61-'64
'64-'65
'65-'69
'69-'71
22 cm
10
0
20
40
DISTANCE (KM)

Fig 13


A
RELATIVE MOVEMENT (CM)
8
4
0
1960
1970
YEAR

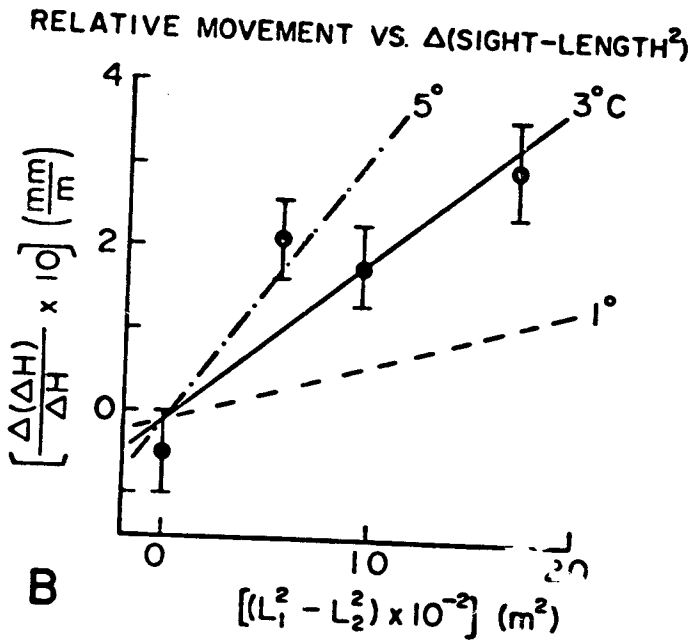

RELATIVE MOVEMENT VS. Δ(SIGHT-LENGTH²)
4
2
0
5°
3°C
1°
B
0
10
[(L₁² − L₂²) x 10⁻²] (m²)

Fig 14A + B

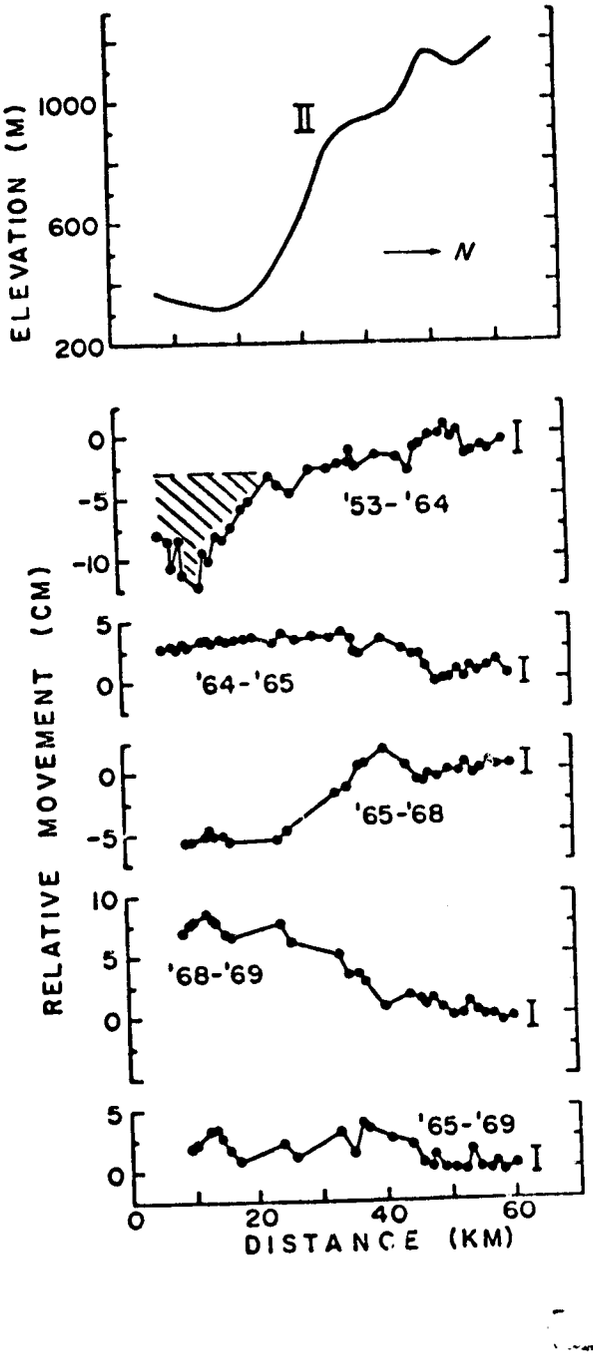
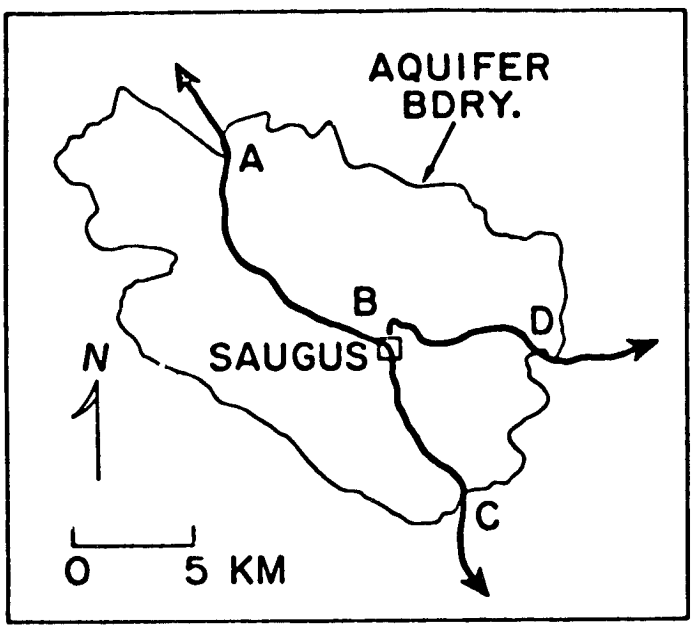

ELEVATION (M)
1000
600
200
II
N
RELATIVE MOVEMENT (CM)
0
-5
-10
I
'53-'64
5
0
'64-'65
0
-5
'65-'68
10
5
0
'68-'69
5
0
'65-'69
0
20
40
60
DISTANCE (KM)

Fig 15


AQUIFER
BDRY.
A
B
D
N
SAUGUS
C
0
5 KM

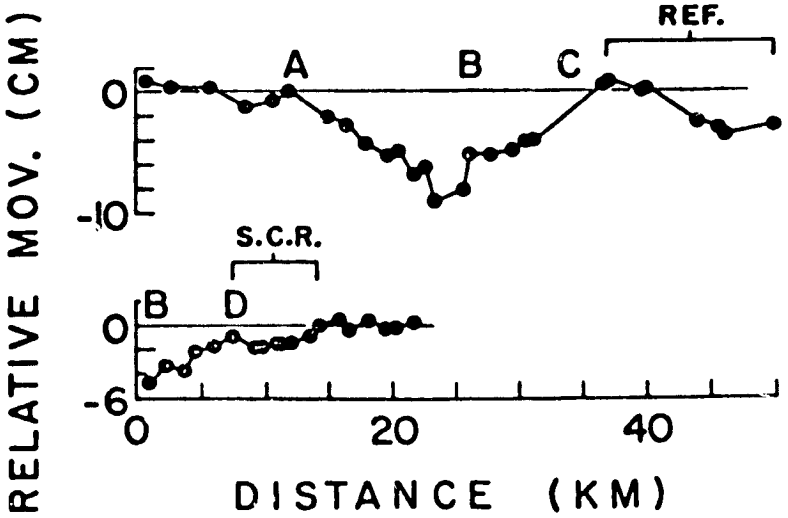

REF.
A
B
C
0
-10
S.C.R.
B
D
0
-6
0
20
40
RELATIVE MOV. (CM)
DISTANCE (KM)

Fig 16

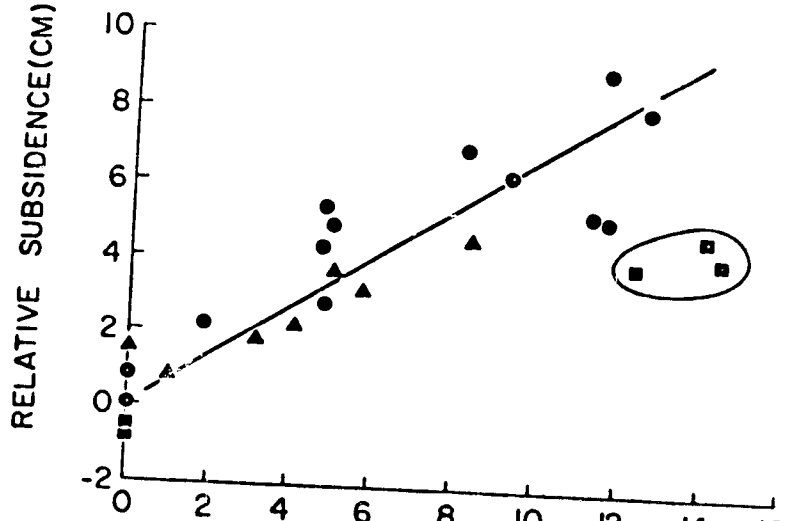


Fig 17

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OF POOR QUALITY

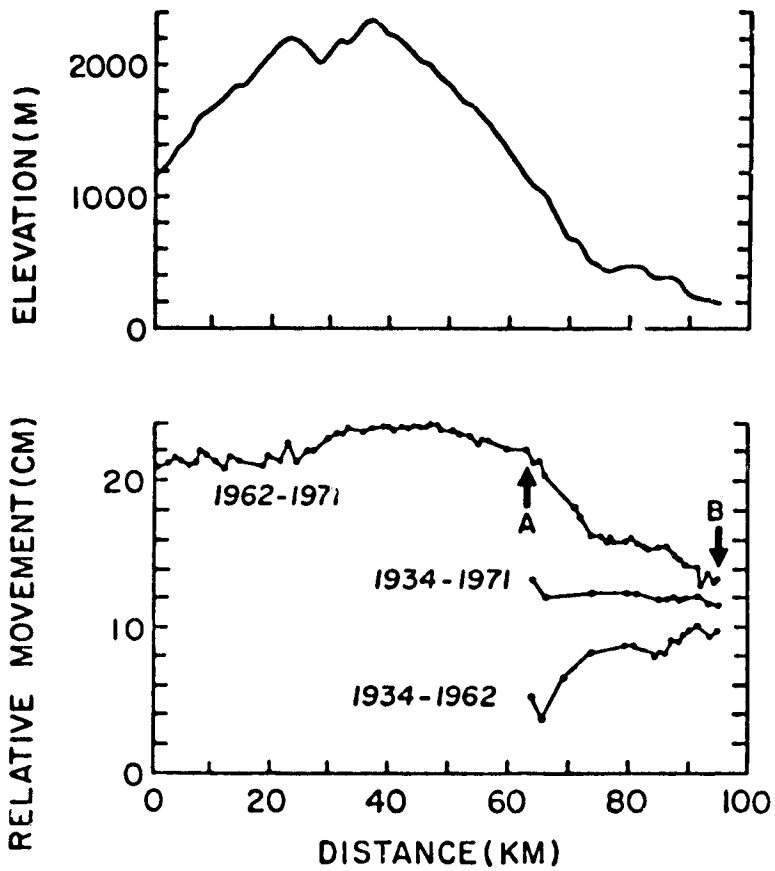
ELEVATION (M)
2000
1000
0
RELATIVE MOVEMENT (CM)
20
10
0
1962-1971
A
B
1934-1971
1934-1962
0
20
40
60
80
100
DISTANCE (KM)

Fig 18

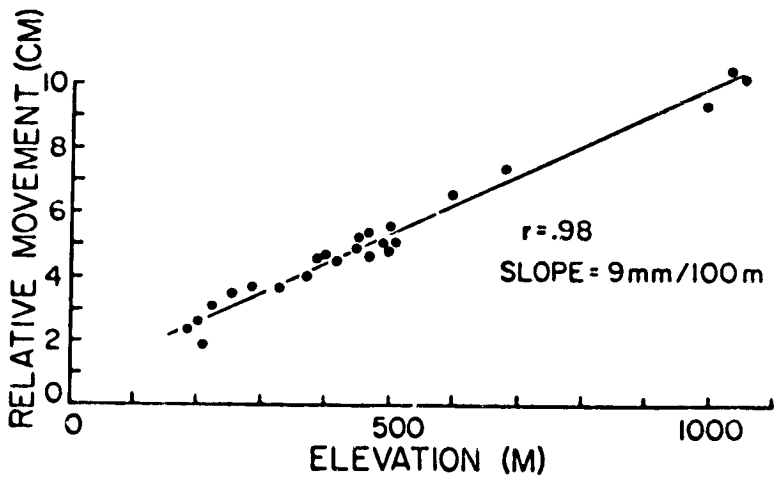
RELATIVE MOVEMENT (CM)
0
2
4
6
8
10
0
500
1000
ELEVATION (M)
r=.98
SLOPE = 9mm/100m

Fig 19

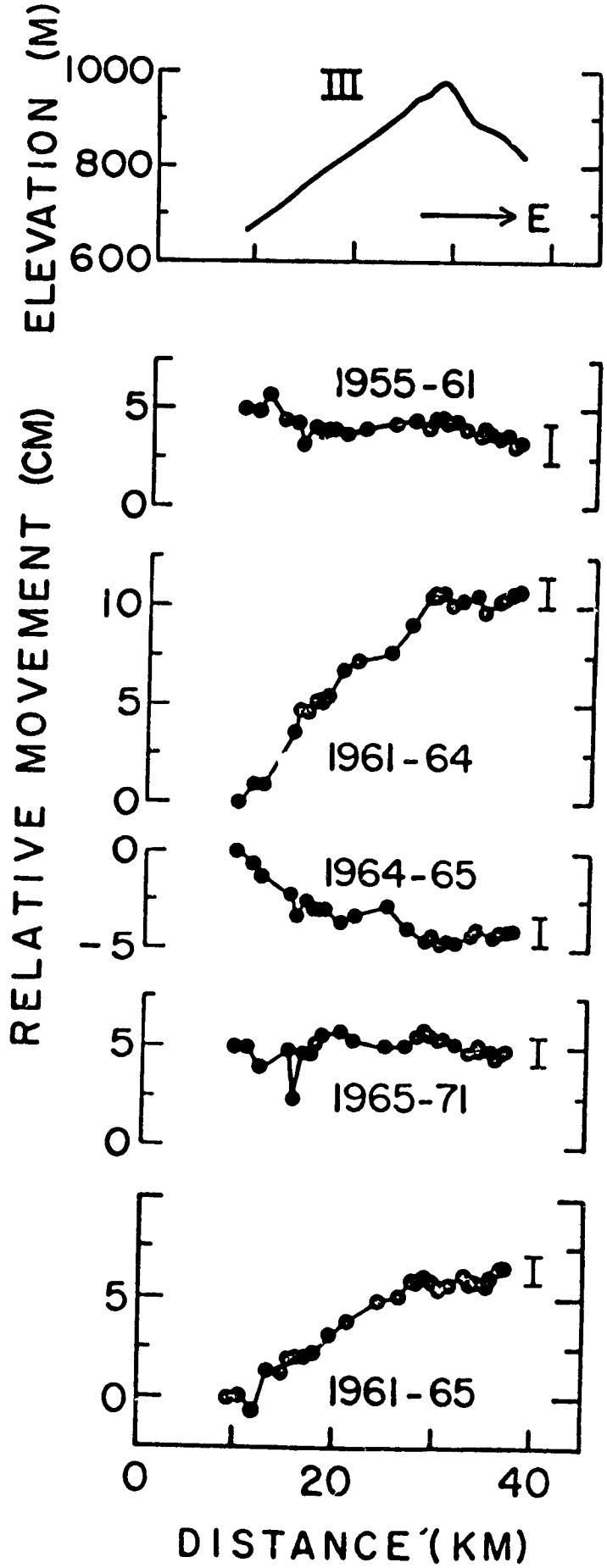
ELEVATION (M)
1000
800
600
III
E
RELATIVE MOVEMENT (CM)
1955-61
5
0
I
10
5
0
1961-64
I
0
-5
1964-65
I
5
0
1965-71
I
5
0
1961-65
I
0
20
40
DISTANCE (KM)

Fig 20

APPENDIX III

GEOPHYSICAL RESEARCH LETTERS, VOL. 7, NO. 11, PAGES 1017-1019, NOVEMBER 1980

ELEVATION CHANGES NEAR THE SAN GABRIEL FAULT, SOUTHERN CALIFORNIA

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Abstract. Analysis of repeated leveling observations in the vicinity of the San Gabriel fault in Southern California indicate subsidence immediately south of the fault relative to points to the north, south, and east. These observations were previously interpreted as reflecting tectonic motions associated with either the "Palmdale Bulge" or with preseismic effects of the San Fernando earthquake. Relative subsidence between 1953 and 1964 reaches approximately 9 cm and extends over a distance of more than 20 km. Subsidence occurs directly above the Saugus aquifer and shows a temporal correlation with the history of water level decline within the aquifer. The degree of subsidence of individual benchmarks is roughly proportional to the product of aquifer thickness and water level decline at the location of the benchmarks. These observations strongly suggest that movements of the surface near the San Gabriel Fault, previously inferred to be of tectonic origin, actually result from near surface sediment compaction within the Saugus basin.

Introduction

Releveling estimates of crustal movement may reflect tectonic deformation, nontectonic motion, or systematic measurement errors. Proper interpretation of releveling measurements entails descriminating between these effects. The releveling measurements in Southern California which are presented here were initially interpreted as representing tectonic movements associated with the "Palmdale Bulge" (Castle et al., 1976). More recently, Strange (1980) has sug-

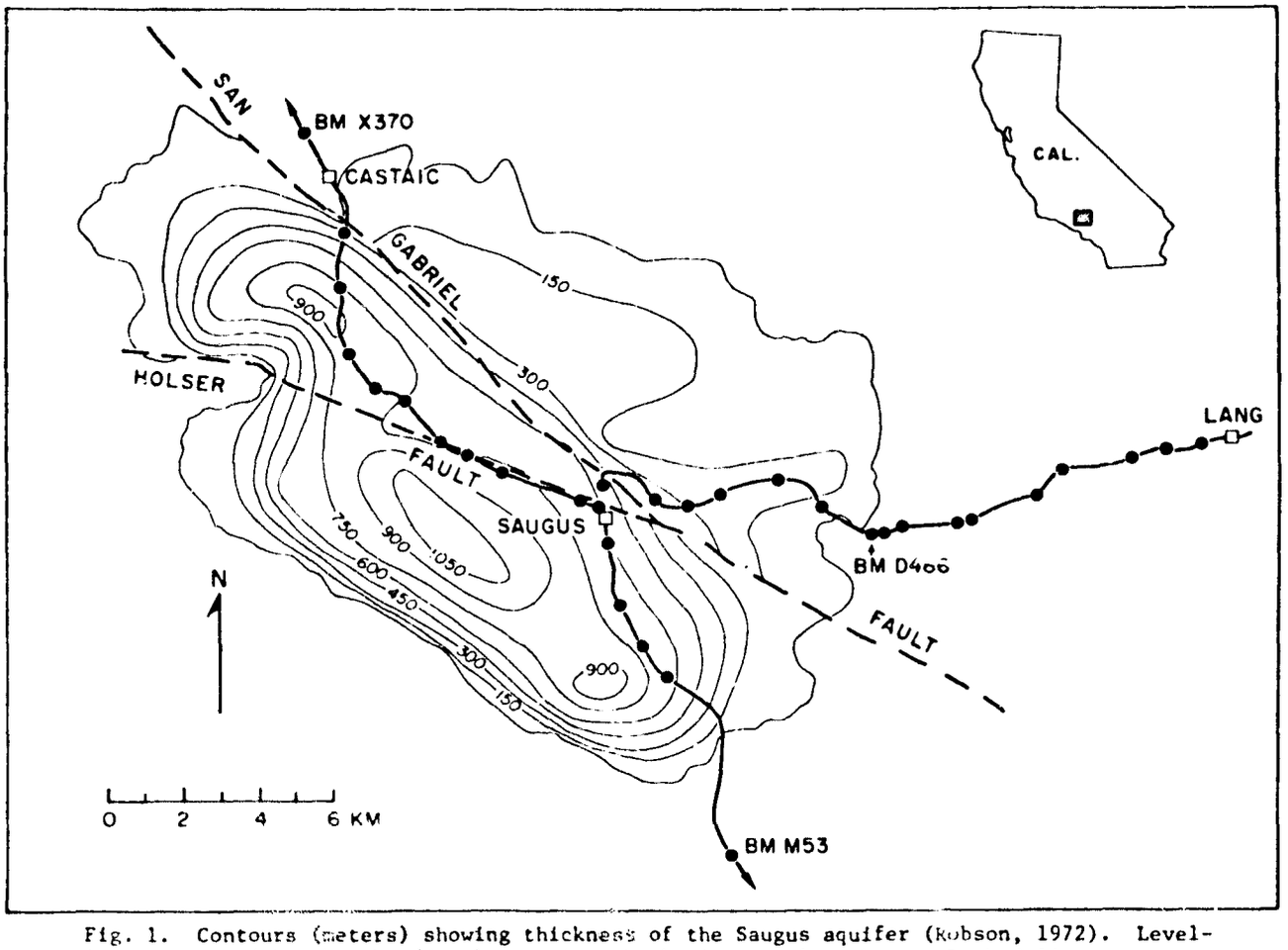

Fig. 1. Contours (meters) showing thickness of the Saugus aquifer (Robson, 1972). Leveling routes and benchmarks (dots) crossing the basin are also shown. Dashed lines are faults.

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Paper number 80L1439.
0094-8276/80/0080L-1439$01.00

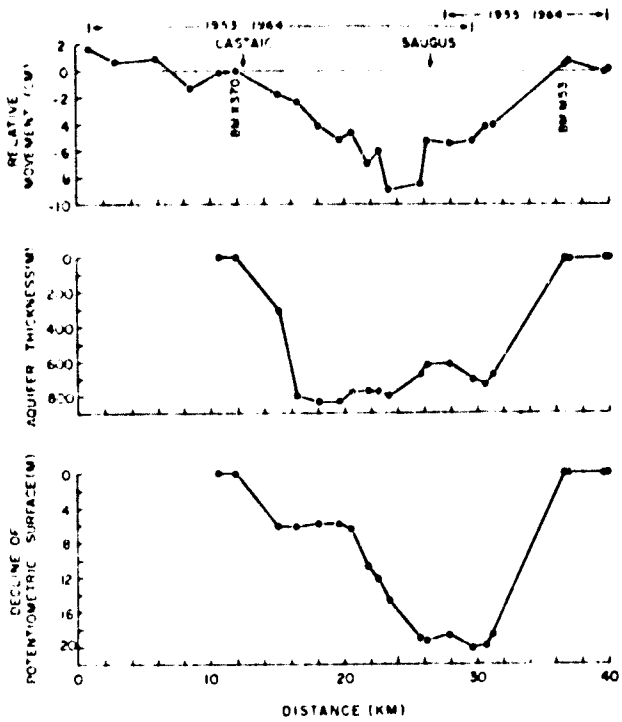
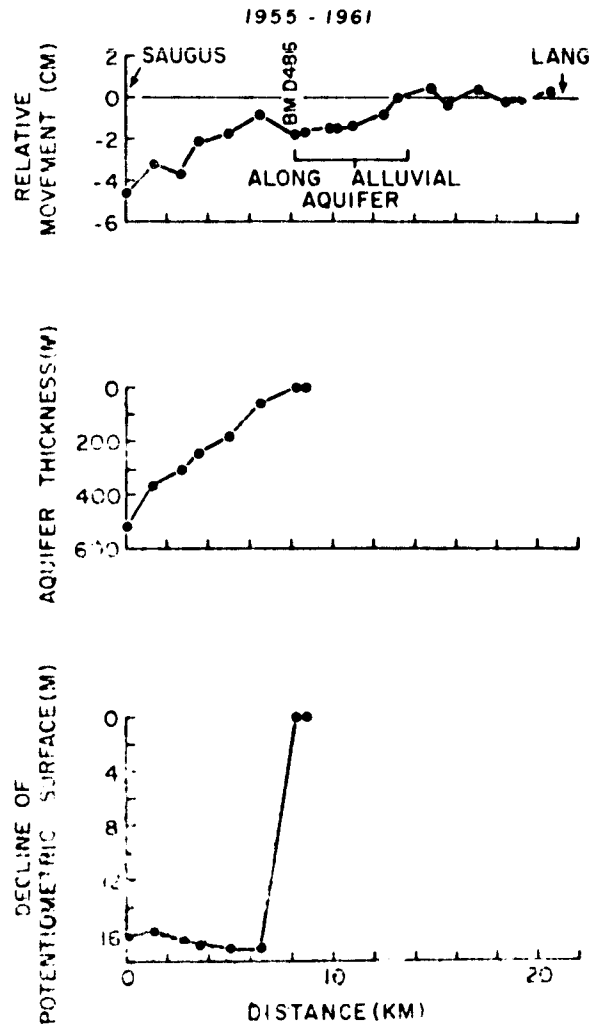


Fig. 2. A) Top: Movement of benchmarks along approximately North-South route plotted versus distance along route. BMX370 on northern periphery of aquifer assumed stable. Years of leveling surveys shown at top. Middle: Profile of thickness of Saugus aquifer along leveling route (from Robson, 1972). Bottom: Profile of decline in potentiometric surface (water level) as of 1963 relative to 1945 surface (from Robson, 1972). B) Profiles along route from north of Saugus to Lang (see Fig. 1). Same format as A.

gested that these measurements reflect preseismic deformation associated with the 1971 San Fernando earthquake. In this study, evidence is presented which strongly suggests that movements of the ground surface south of the San Gabriel fault in Southern California result, for the most part, from near surface sediment compaction due to fluctuations of the water level within the Saugus aquifer and not from tectonic activity. Although these crustal movements most likely are not tectonic in origin, the do represent real surface movements and in this sense further demonstrate the ability of precise leveling to detect relatively subtle deformation.

Data

Figure 1 shows contours of the thickness of the confined Saugus aquifer (Robson, 1972) with leveling lines and benchmarks superimposed. Profiles of elevation change, aquifer thickness and estimated change in potentiometric surface (water level) along the survey routes are shown in Figure 2a,b. The elevation change profile from north of Castaic through Saugus is plotted assuming stability for benchmark X370 immediately adjacent to the aquifer. This benchmark is chosen as a reference since it lies outside of the aquifer and appears generally stable relative to benchmarks further north. The elevation change profile from Saugus to Lang is plotted assuming stability near Lang since the benchmarks immediately adjacent to the Saugus aquifer occur within the alluvial aquifer of the Santa Clara River and appear to have subsided themselves. The change in water level shown in Figure 2a,b represents the calculated decline as of 1963 relative to the steady state level estimated to have existed around 1945 (Robson, 1972). This decline was due to pumping from the Saugus aquifer as well as overlying alluvial aquifers, and to an extended drought (Robson, 1972, pp. 8, 40). Because the water level was already below the steady state at the time of the first leveling survey (1953), the decline in



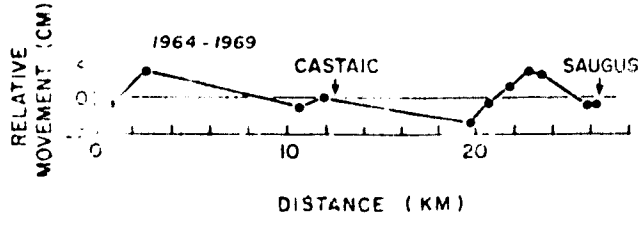


Fig. 3. Profile of elevation change for period 1964 to 1969 along route from north of Castaic to Saugus (see Fig. 1 for location).

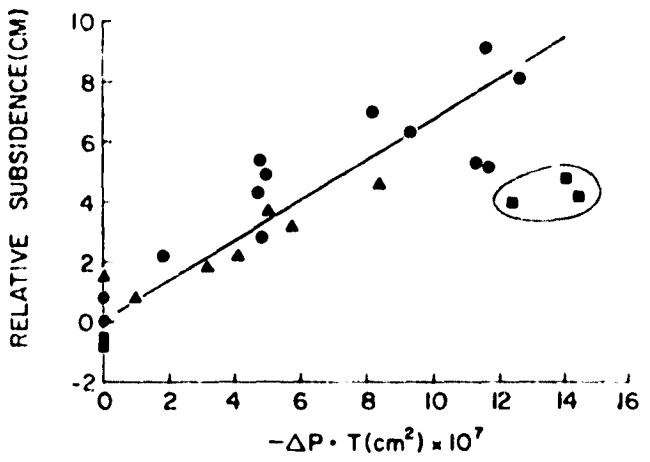


Fig. 4. Plot of subsidence versus change in potentiometric surface times aquifer thickness for benchmarks in and immediately adjacent to aquifer. Different symbols refer to different releveled segments: circles-north of Castaic to Saugus (1953-1964); squares-Saugus to South (1955-1964); triangles-Saugus to Lang (1955-1961).

water level shown in Figure 2a,b is somewhat larger than the actual change between leveling surveys. However, water levels declined more or less continuously up to 1963 (Robson, 1972).

Figure 2a,b indicate a reasonable spatial correlation between subsidence, aquifer thickness and change in water level. In addition, the water level within the Saugus aquifer stabilized after 1963 and began to recover around 1968 (Robson, 1972). Examination of post-1964 elevation changes across the northern part of the basin indicate a general absence of subsidence of the basin (Figure 3). Thus, the observed elevation changes correlate spatially with aquifer geometry and temporally with the history of water level decline.

In the ideal case, the amount of compression (subsidence) of a sediment layer is proportional to the thickness of the layer and the increase in effective pressure (grain to grain pressure) on the sediments (Terzaghi and Peck, 1967; p. 2[illegible]). Lowering the water level produces a proportional reduction in the buoyant force between grains and hence causes an increase in the effective pressure on the underlying sediments (Terzaghi and Peck, 1967; p. 384). Assuming uniform sediment characteristics throughout the aquifer (constant compressibility), to a first approximation the amount of compaction (subsidence) at a particular location will be roughly proportional to the product of water level decline and aquifer thickness. Figure 4 gives a plot of relative subsidence versus the product of the change in water level and aquifer thickness for benchmarks in and immediately adjacent to the Saugus aquifer. Considering the uncertainties in aquifer parameters and probable spatial variations in aquifer compressibility, the observations fit the simple linear relationship [illegible]. The three circled points on the right side of the plot lie in the southeastern portion of the aquifer (Figure 1) and may reflect either different sediment characteristics or possibly tectonic movement in this area. The slope of the straight line in Figure 4 is an estimate of the average compressibility of the Saugus aquifer. Although the compressibility can vary through at least two orders of magnitude depending on sediment type, stress level, and stress history, the value obtained from Figure 4 ($\approx 7 \times 10^{-8} cm^{-1}$) is quite comparable to those reported for other confined aquifers in California (Lofgren, 1979, p. 35).

Conclusions

Repeated leveling surveys conducted between 1953 and 1964 indicate subsidence of the Saugus Basin relative to benchmarks on its periphery. Subsidence extends over a distance of about 20 km with maximum relative subsidence near the center of the basin reaching 9 cm. The spatial correlation between the area of relative subsidence and the Saugus aquifer, the temporal correlation with the history of water level decline, and the parameter dependence indicated in Figure 4, strongly suggest that movements in the Saugus Basin, previously inferred to be associated with either the "Palmdale Bulge" or preseismic effects of the 1971 San Fernando earthquake, result from near-surface sediment compaction. This study demonstrates the need for caution when applying tectonic interpretations to releveling observations in areas of unconsolidated sediments. On the other hand, the fact that the leveling measurements accurately detected real surface movements under actual field conditions demonstrates the potential of this technique for monitoring crustal deformation.

Acknowledgements. I am grateful to the National Geodetic Survey for supplying the leveling data used in this study, and to Jack Oliver and Larry Brown for helpful comments. This research was supported in part by U.S. Geological Survey Grant 14-08-0001-17625, and N.A.S.A. Grant NAG5-40. Cornell Department of Geological Sciences Contribution No. 679.

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Strange, W.E., The impact of refraction correction on leveling interpretations in California, Eos, Trans. Am. Geophys. Union, 61, 367-368, 1980.

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APPENDIX IV

VERTICAL CRUSTAL MOVEMENTS IN THE VICINITY OF THE 1931 VALENTINE, TEXAS, EARTHQUAKE

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Ithaca, New York 14853

ABSTRACT

Deformation possibly related to an earthquake in Texas has been revealed from analysis of precise leveling data collected by the National Geodetic Survey. Relative elevation changes deduced from first-order leveling surveys conducted between 1910 and 1957 show that at least two areas in western Texas near the epicenter of the 1931 Valentine earthquake (M = 6.4) subsided relative to their surroundings. Apparent subsidence east of Van Horn, Texas is attributed primarily to the combined effects of groundwater withdrawal and topography-related survey errors. In contrast, relative subsidence near Valentine, Texas, of 11.2 $\pm$ 1.0 cm extending over a distance of about 20 km does not appear to be due to either near-surface effects or leveling errors and thus may represent coseismic deformation of the Valentine earthquake. The predominance of subsidence over uplift indicated by the leveling data suggests a normal fault mechanism for the earthquake. This is consistent with the Cenozoic structure of the region which is characterized by Basin and Range morphology. If the observed subsidence was caused by the 1931 earthquake, the leveling data suggest that the epicenter for this event lies considerably closer to Valentine than originally reported. This new location is supported by a recent relocation for the Valentine earthquake, and by local intensity reports.

INTRODUCTION

Western Texas, (Figure 1) structurally a part of the Basin and Range Province, is considered by some to be the eastern offset of the Rio Grande Rift (Woodward et al., 1975; Muehlberger et al., 1978). Faults of Tertiary and Quaternary age shape the landscape into alternating horsts and grabens. The grabens are commonly filled with large thicknesses of unconsolidated alluvial, lacustrine clay, silt. sand and gravel. These deposits are the major sources of groundwater in western Texas. Evidence of recent tectonic activity in this area includes north- and northwest-trending faults offsetting Quaternary geomorphic surfaces (King, 1965; Muehlberger et al., 1978), historic seismicity (Coffman and von Hake, 1973; Sanford and Toppozada, 1974), and recent uplift in the Diablo Plateau area (Brown et al., 1978; Reilinger et al., 1980).

Since very little is known about the long-term earthquake history of western Texas, it is difficult to evaluate seismic risk from analysis of earthquake statistics. The Valentine, Texas, earthquake of August 16, 1931 (M = 6.4), was the strongest reported earthquake to occur in the region (Sellards, 1932). Byerly (1934) located the epicenter at 30.9°N and 104.2°W, in the Davis Mountains, 38 km northeast of Valentine, Texas (Figure 1). More recently, Dumas et al. (1980) have used regional and local teleseismic P-wave arrival time anomalies from the Gnome underground nuclear explosion to determine station corrections and relocate the 1931 event at 30.69°N, 104.57°W (Figure 1). This relocation is considerably closer to Valentine than the previously determined epicenter. First P motions analyzed by Byerly (1934) and later reexamined by Sanford and Toppozada (1974) suggest predominantly normal faulting, striking N40°W and

dipping 74° SW. Dumas et al. (1980), using the same first P motion observations, have determined a new fault plane solution and suggest that the earthquake occurred on a predominantly strike-slip fault striking N59° W and dipping 70° NE.

The inconsistency of fault planes determined by seismic data and the lack of evidence of surface faulting in the Valentine region make it difficult to assign this event to a specific fault. A very poorly located earthquake on 1 August 1975 (M_L 4.8), was felt in Valentine, however no focal mechanism was determined for this event. Nevertheless, the Valentine region is seismically active at present (Dumas et al., 1980).

In this study we present elevation changes deduced from precise leveling surveys in the vicinity of the Valentine earthquake. Because leveling measurements can be influenced by near surface movement, systematic errors, and tectonic deformation, specific criteria are applied to separate these effects. Anomalous relative subsidence near Valentine, Texas suggests the existence of an active buried fault that may have been the source of the 1931 Valentine earthquake. If so, this is the first documented case of contemporary earthquake deformation in Texas.

DATA ANALYSIS

The primary data used in this study are estimates of crustal movement derived from repeated leveling surveys conducted by the National Geodetic Survey and water-well levels reported by the U.S. Geological Survey and the Texas Water Commission.

The details of the leveling procedure and the method used to derive crustal movement information from repeated leveling surveys have been described elsewhere (Bomford, 1971; Brown and Oliver, 1976). Briefly,

leveling is an optical technique used to determine the differences in elevation between closely spaced ($\sim$ 1 km) points (benchmarks) on the earth's surface. Repeating a survey at some later time (Δt = tens of years) results in new elevation observations that can be compared to the original observations to estimate relative movement.

Figure 1 shows the locations of the leveling routes used in this study, and Figures 2a and 3a, elevation changes along these routes. All of the surveys used to determine elevation changes were performed between 1910 and 1957 by the National Geodetic Survey according to first-order standards. There is no evidence of abnormal benchmark instability in their published descriptions or from field inspection (J. Dorman, 1979, Personal communication). The observed elevation changes in Figures 2a and 3a have been plotted with respect to benchmark Y17 (Sierra Blanca) as a reference. This does not restrict the data since only relative movements are significant.

Apparent crustal movements between Sierra Blanca and Sanderson, Texas, are shown by the profile in Figure 2a, which reveals, among other things, a broad regional down-tilting to the southeast. The magnitude of this tilt is about 0.76 μrad. It is difficult to ascertain whether this tilting is due to systematic leveling errors or to a regional tectonic effect (Brown et al., 1980). For example, since the route is a partially north-south line, tidal and unequal lighting errors (Bomford, 1971; Balazs, 1975) could have some influence on the data. In either case, this signal is almost certainly not directly related to the Valentine earthquake. Because of this, the data were adjusted to remove the regional tilt. Figure 2a shows the adjusted elevation change profile along with the unadjusted profile. Both profiles indicate anomalous subsidence near Valentine, Texas, although

with slightly different magnitudes. The significance of this feature and its implication for the 1931 Valentine earthquake will be discussed in the following section.

The profile of movements between Sierra Blanca and Monahans (Figure 3a) shows no regional tilting, possibly a consequence of its predominantly east-west orientation. However, this profile does show a negative correlation with topography between benchmark BM H17 (near Van Horn) and BM S16 (near Pecos). There are a number of known systematic leveling errors that accumulate with elevation (unequal atmospheric refraction; rod miscalibration; Bomford, 1971). Although a correlation with topography does not in itself invalidate the leveling observations, it does constitute a negative factor as far as credibility is concerned (Brown et al., 1980). In an attempt to remove the effects of possible systematic errors, a plot of relative movement versus elevation was constructed for the segment of the movement profile that correlates with topography (Figure 4). The slope of the regression line fit to this plot is 25 mm/100 m, corresponding to the average magnitude of the elevation-correlation. Removing the part of the signal which correlates with topography results in the adjusted profile segment shown in Figure 3b. The broad subsidence features between BM H17 and BM S16 is no longer apparent. The adjusted profile shows few outstanding features except for a small relative subsidence between BM 3826 and BM H17, and a large relative subsidence centered near Pecos, Texas. Subsidence near Pecos has been shown to be the effect of sediment compaction due to groundwater withdrawal (Rosepiler and Reilinger, 1977). Subsidence between BM 3826 and BM H17 also appears to be due to near surface movements and will be discussed further in the next section.

The other data used in this study are water levels in wells located

near the leveling routes. The geographical distribution and thickness of the principal aquifers in the vicinity of Van Horn and Lobo, Texas, are given by Gates et al (1978) and Hood and Scalapino (1951) . Pumping for irrigation from these aquifers began in 1949 (Gates et al., 1978), increased rapidly in the early 1950's and has remained fairly stable to the present. As water withdrawal is larger than natural recharge, well levels have declined in these basins. Water level declines reached 15 meters between 1949 and 1957 near Lobo, Texas (Figure 2b). The decline in water levels for wells located near the leveling routes are shown in Figures 2b and 3b (data from Gates et al. (1978) and Hood and Scalapino (1951)).

DISCUSSION AND INTERPRETATION

Elevation Change Due to Water Level Decline

Whenever the water table is lowered, the effective pressure on the underlying sediments is increased. The increase in effective pressure, or grain-to-grain loading, causes the sediments to compact which results in subsidence of the land surface.

Land subsidence due to withdrawal of groundwater has been identified in the Pecos region (Rosepiler and Reilinger, 1977) and in various other parts of North America where large volumes of water are being removed from underground aquifers (Poland and Davis, 1969). Therefore, care must be taken to avoid misinterpreting subsidence due to fluid withdrawal as tectonic motion. Inspection of Figure 2b shows a strong correlation between water-level decline and land subsidence between BM L23 and BM Q23. This strongly suggests that the land subsidence in the Lobo Flat region is due to compaction of the aquifer in response to decline in hydrostatic

pressure. Compaction probably occurs chiefly in the intrabedded clay and reworked volcanic clastics. The ratio of land subsidence to decline in water level is about 1 to 400, equivalent to 25 mm subsidence for 10 m of water-level decline, a value comparable to that observed in other areas in the United States (Poland and Davis, 1969). It is important to note that the apparent subsidence south of Lobo Flat, i.e., near Valentine, is not correlated with water-level decline and therefore is most likely not due to pumping (Figure 2b).

A less-extensive land subsidence due to water withdrawal is indicated west of Van Horn along the Sierra Blanca-Pecos profile (Figure 3b) by a similar correlation of subsidence and water-level decline. The primary effect is confined between BM 3826 and BM H17. The ratio of land subsidence to water-level decline at BM 3826 is 1:700.

Implications for the 1931 Valentine Earthquake

As noted in the previous section, the anomalous relative subsidence centered near Valentine is not easily attributed to near-surface effects, and the magnitude of this subsidence is considerably greater than random error. It is therefore suggested that the observed relative subsidence represents tectonic deformation associated with the Valentine earthquake. The predominance of subsidence over uplift indicated by the leveling data suggests a normal fault mechanism for the earthquake (Savage and Hastie,1966).

Both fault plane solutions for the 1931 Valentine earthquake (Sanford and Topozada, 1974; Dumas et al., 1980) suggest a Northwest striking fault. This trend is in good agreement with both the regional structural grain and the long axis of the isoseismal contours for the 1931 earthquake given by Sellards (1932). The leveling route traversing the Valentine area is also oriented in a

generally northwest direction. Therefore, if the observed subsidence is due to fault slip as suggested, the areal extent of subsidence ($\sim$30 km) is a rough estimate of the length of the earthquake fault. Kanamori and Anderson (1975) give an empirical relation between fault dimensions and surface wave magnitudes. Using these relationships and assuming a fault width of $\sim$10 km suggests a magnitude in the range 6.3 - 7.9 for the Valentine earthquake. Although not unreasonable, this is somewhat larger than the magnitude estimated for this event (Mb=5.8 to 6.4, Gutenburg and Richter, 1949; Dumas et al., 1980). However, the fault length determined from the zone of subsidence is probably overestimated for the following reasons: 1) surface deformation will extend beyond the ends of the fault; and 2) observed subsidence could include the effects of preseismic and/or postseismic slip on the fault or on extensions of the fault.

Although the details of the faulting responsible for the Valentine earthquake are poorly constrained by the available data, the observed pattern of surface deformation is at least consistent with a northwest-striking normal fault dipping southwest and located just north of Valentine, Texas. This location is in agreement with the area of maximum intensity for this event (Sellards, 1932), and with the recomputed epicenter reported by Dumas et al. (1980).

CONCLUSIONS

Relative elevation changes deduced from repeated leveling surveys in the Van Horn-Valentine region of western Trans-Pecos, Texas, indicate significant vertical movements between 1910 and 1957. Although some of the movements appear to be due to water withdrawal (e.g., near Van Horn and

Lobo, Texas), the most anomalous movements occur near Valentine and do not appear to be associated with groundwater withdrawal or leveling errors. Subsidence near Valentine, reaching 11.2 ± 1.0 cm relative to points to the northwest and southeast, is thus most likely of tectonic origin. The pattern of deformation near Valentine is consistent with normal faulting on a northwest-striking, southwest-dipping fault, in agreement with one interpretation of the focal mechanism inferred for this event. This earthquake is the largest recorded in Texas and the observations presented here may represent the first case of earthquake deformation in this region.

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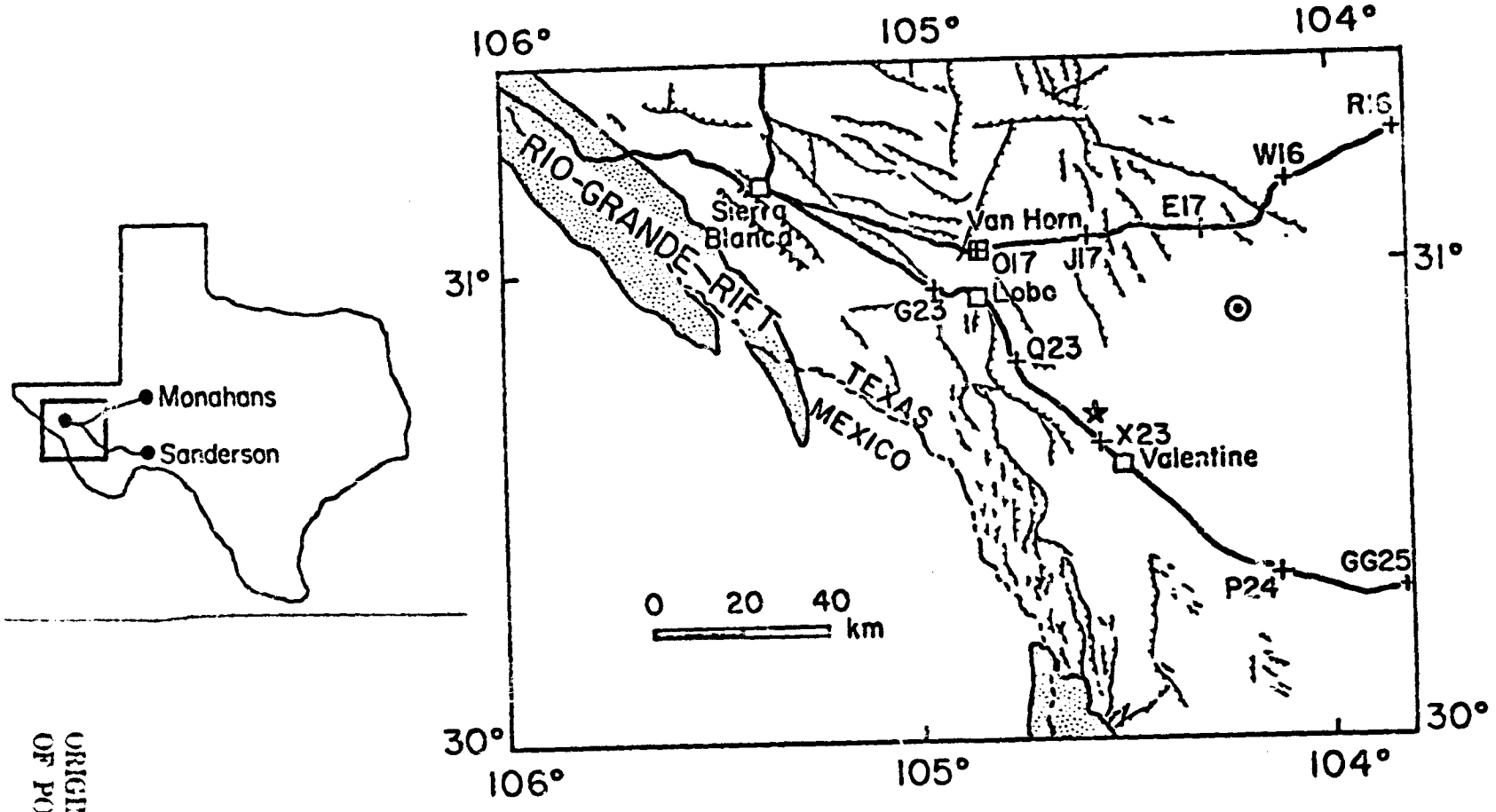
FIGURE CAPTIONS

Figure 1. Regional geography and Cenozoic faults of the study area. Heavy lines indicate leveling routes. Open squares indicate cities. Crosses represent location of benchmarks. Epicenter locations of the 1931 Valentine, Texas, earthquake reported by Byerly (1974), and Dumas et al. (1980) are also represented.

Figure 2. (a) Profiles of relative vertical crustal movements from Sierra Blanca to Sanderson, Texas, derived from N.G.S. surveys conducted in 1917 and 1957. Unadjusted profile is shown by crosses; adjusted profile, by solid circles. The upper curve reflects the topography along the profile. Locations of some cities and benchmarks discussed in the text are indicated. (b) Expanded display of recent crustal movements and water-level decline (open circles) in the vicinity of Lobo, Texas.

Figure 3. (a) Profiles of relative vertical crustal movements between Sierra Blanca and Monahans, Texas, derived from N.G.S. surveys conducted in 1910 and 1956. Topography is shown between the unadjusted profile (crosses) and adjusted profile (solid circles). Location of some cities and benchmarks discussed in the text are indicated. (b) Expanded profile of recent crustal movements and water-level decline (open circles) near Van Horn, Texas.

Figure 4. Plot of net elevation change versus elevation.



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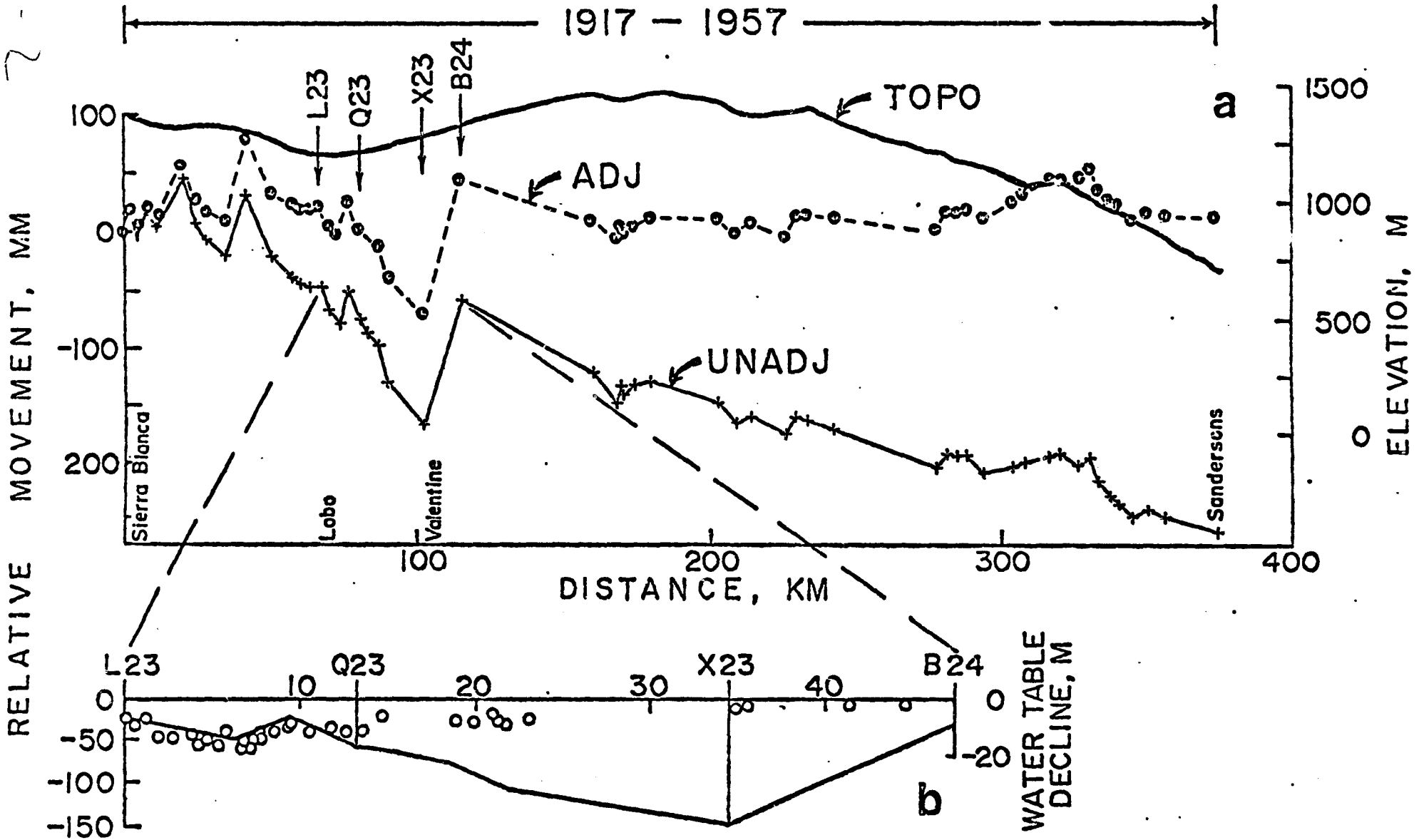
1917 — 1957
L23
Q23
X23
B24
TOPO
ADJ
UNADJ
a
b
100
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200
1500
1000
500
0
RELATIVE MOVEMENT, MM
ELEVATION, M
Sierra Blanca
Lobo
Valentine
Sandersons
100
200
300
400
DISTANCE, KM
L23
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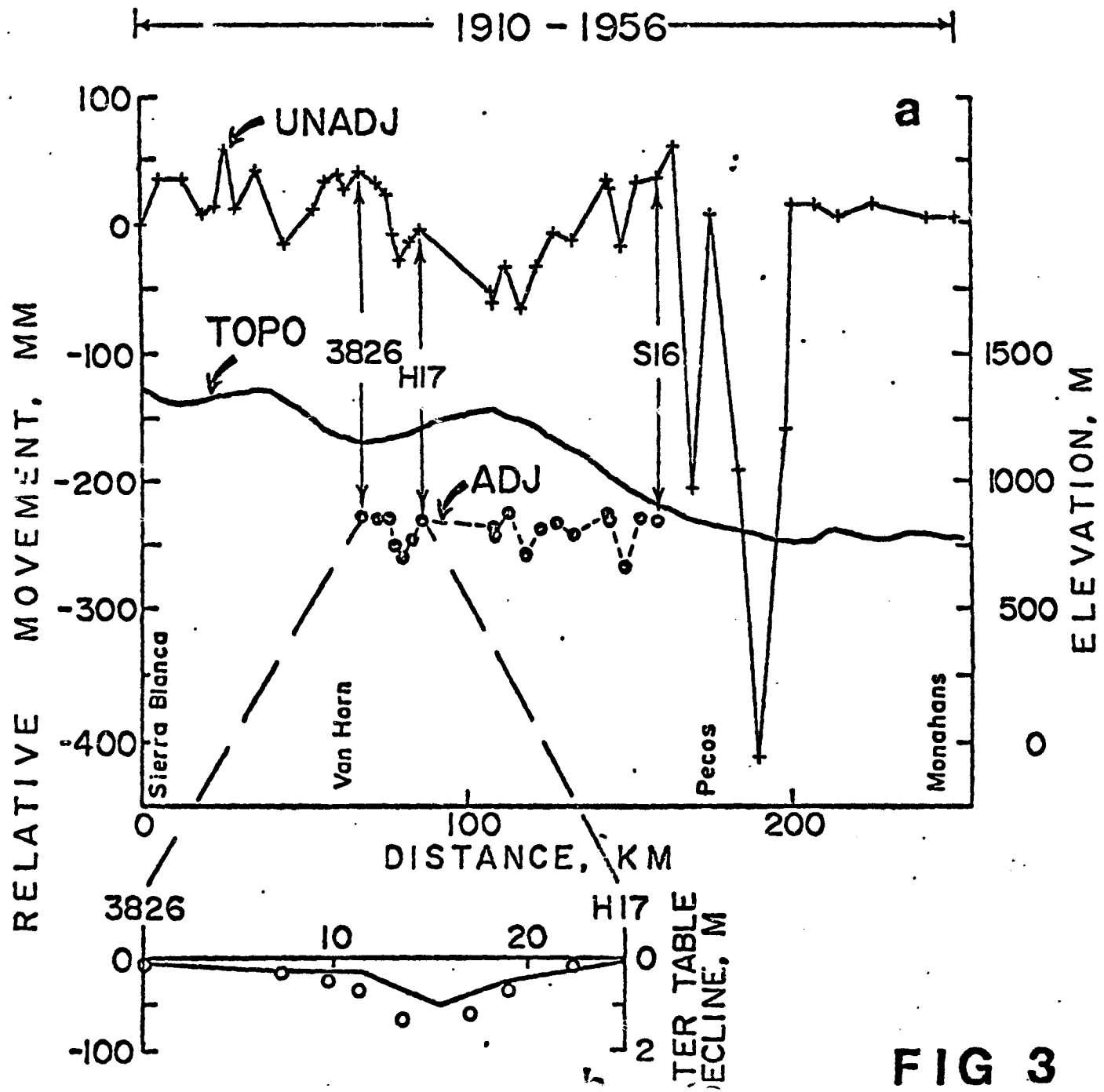
WATER TABLE DECLINE, M

FIG 2


1910 - 1956
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-200
-300
-400
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1000
500
0
RELATIVE MOVEMENT, MM
ELEVATION, M
Sierra Blanca
Van Horn
Pecos
Monahans
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DISTANCE, KM
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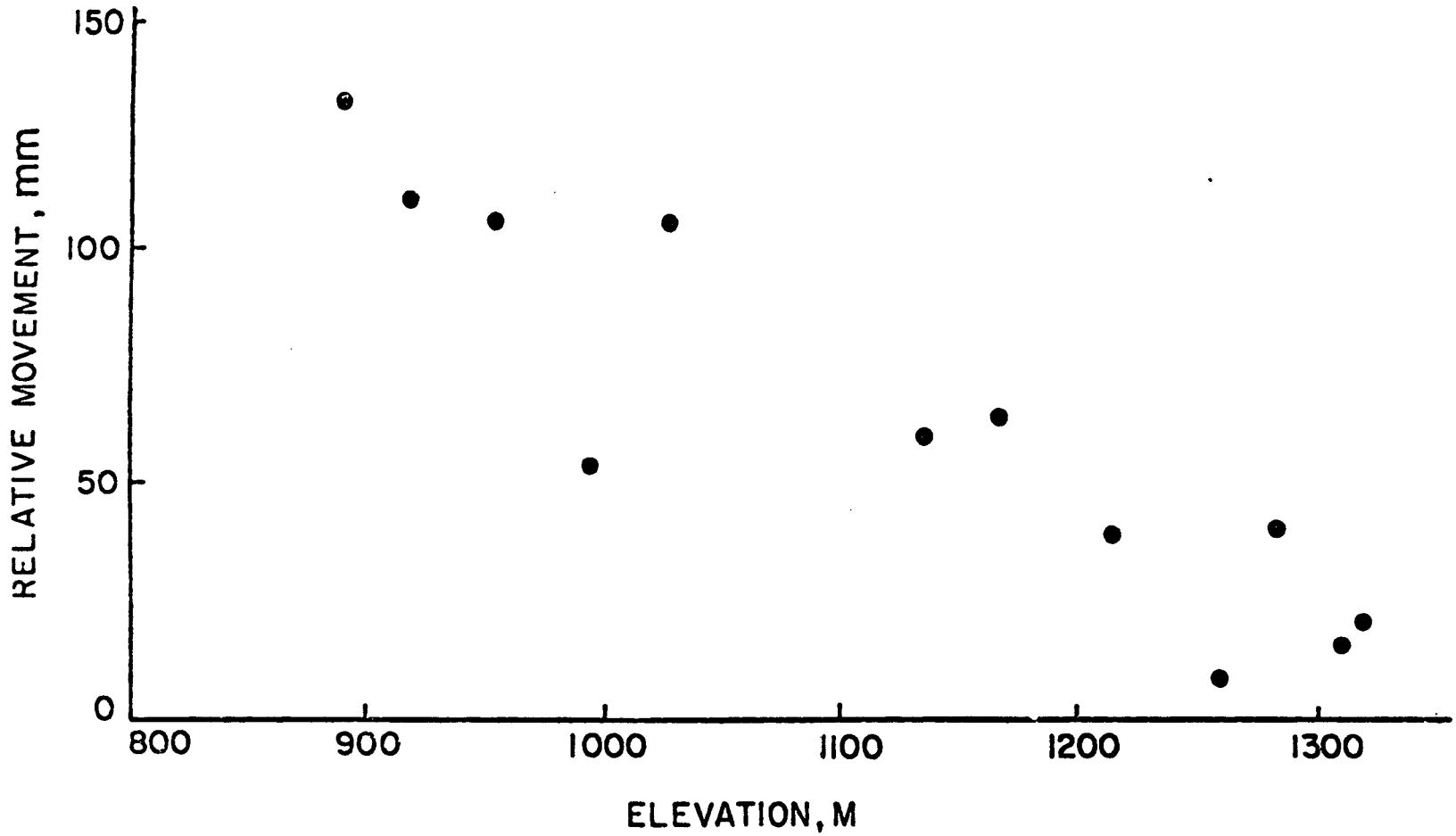
FIG 3


150
100
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0
RELATIVE MOVEMENT, mm
800
900
1000
1100
1200
1300
ELEVATION, M

APPENDIX V

DETECTION OF VERTICAL FAULT DISPLACEMENTS BY PRECISE LEVELING: A CASE STUDY IN WESTERN KENTUCKY*

ABSTRACT

Relative vertical displacements of bench marks in extreme western Kentucky have been determined by comparison of successive leveling surveys in 1947 and 1968. The resulting pattern of apparent surface deformation shows a steep offset which can be closely modeled by a normal fault buried in an elastic half-space. The offset is located near the northern boundary of the Mississipoi Embayment and the New Madrid seismic zone, an area where faults have previously been inferred on the basis of both geological and geophysical evidence. If the apparent movement is due to slip along a fault, several lines of evidence (regional structure, earthquake data, and lineations) suggest that the postulated fault trends north-northeast. Thirteen earthquakes were recorded in this area between the times of leveling; focal mechanisms exist for three of these. The nearest of these three focal mechanisms to the leveling offset implies normal faulting. The magnitude of the earthquake, however, appears to be too small to account for the amount of slip required by the fault model. Thus the apparent deformation may have accumulated with several undetected small

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* By F. Steve Schilt and Robert E. Reilinger

earthquakes, or gradually as aseismic creep.

INTRODUCTION

The purpose of this study is to examine the possibility that anomalous surface deformation near the northern end of the Mississippi Embayment, as measured by repeated precise leveling surveys, was caused by movement along a buried fault. The area studied is between Wickliffe and Paducah, Kentucky, and the leveling data, when supplemented with other geological and geophysical information, suggest dip-slip displacement along an unmapped, concealed fault.

Brown and Oliver (1976) have compared regional structural features of the eastern United States with relative rates of vertical crustal movement as determined from multiple first-order leveling surveys. They found a significant correlation between the velocity patterns and large-scale (hundreds of kilometers) geological structure. A portion of their data for the central United States is examined here on a more local scale, with interest in shorter wavelength behavior.

Coseismic surface deformation has been studied using leveling and other techniques for many large, dip-slip earthquakes (e.g., Savage and Hastie, 1966; Reilinger and Brown, 1981). In most of these examples, movements are large and are easily identified by their close temporal and spatial association with earthquakes, and in some cases, surface faulting. This study involves a more subtle degree of deformation and many kinds of evidence are considered in order to determine if the deformation is tectonic and if so, how it occurred.

REGIONAL STRUCTURE AND GEOLOGY

The Mississippi Embayment (Figure 2.1) began to form in the Late Cretaceous when the Pascola Arch (then connecting the Ozark and Nashville domes) was downwarped. The cause of this subsidence is not well understood, but it has been viewed as an isostatic adjustment subsequent to Precambrian rifting, which extended into the continent as the failed arm of a triple-junction (Burke and Dewey, 1973; Ervin and McGinnis, 1975). Depression continued as a linear trough until late Eocene, with an accumulation of about 1000 meters of sediment in the central portion of the embayment. A cover of alluvial deposits is Late Tertiary in age, and is locally overlain by Pleistocene loess.

Nearly all of the observed surface faults in this region occur in the Paleozoic rocks surrounding the Cretaceous and younger fill of the Mississippi Embayment (Heyl, 1965; Hook, 1974). The major displacements on these Paleozoic faults probably occurred in Pennsylvanian and Mississippian time (King, 1969). Far fewer faults are mapped within the Cretaceous and younger rocks of the embayment (York and Oliver, 1976; Russ, 1979; Zoback, 1979). Recent seismicity (Figure 2.1) defines several linear NE-trending zones approximately parallel to the embayment axis, and some zones trending NW (Stauder and others, 1976; Herrmann and Canas, 1978).

The difficulty of finding mappable faults which can be clearly related to this seismicity may be largely due to a masking effect of the unconsolidated alluvial cover. In an attempt to map faulting within the embayment, Fisk (1944) made a detailed study of lineations, and obtained a grid-like pattern of orthogonal faults for the entire

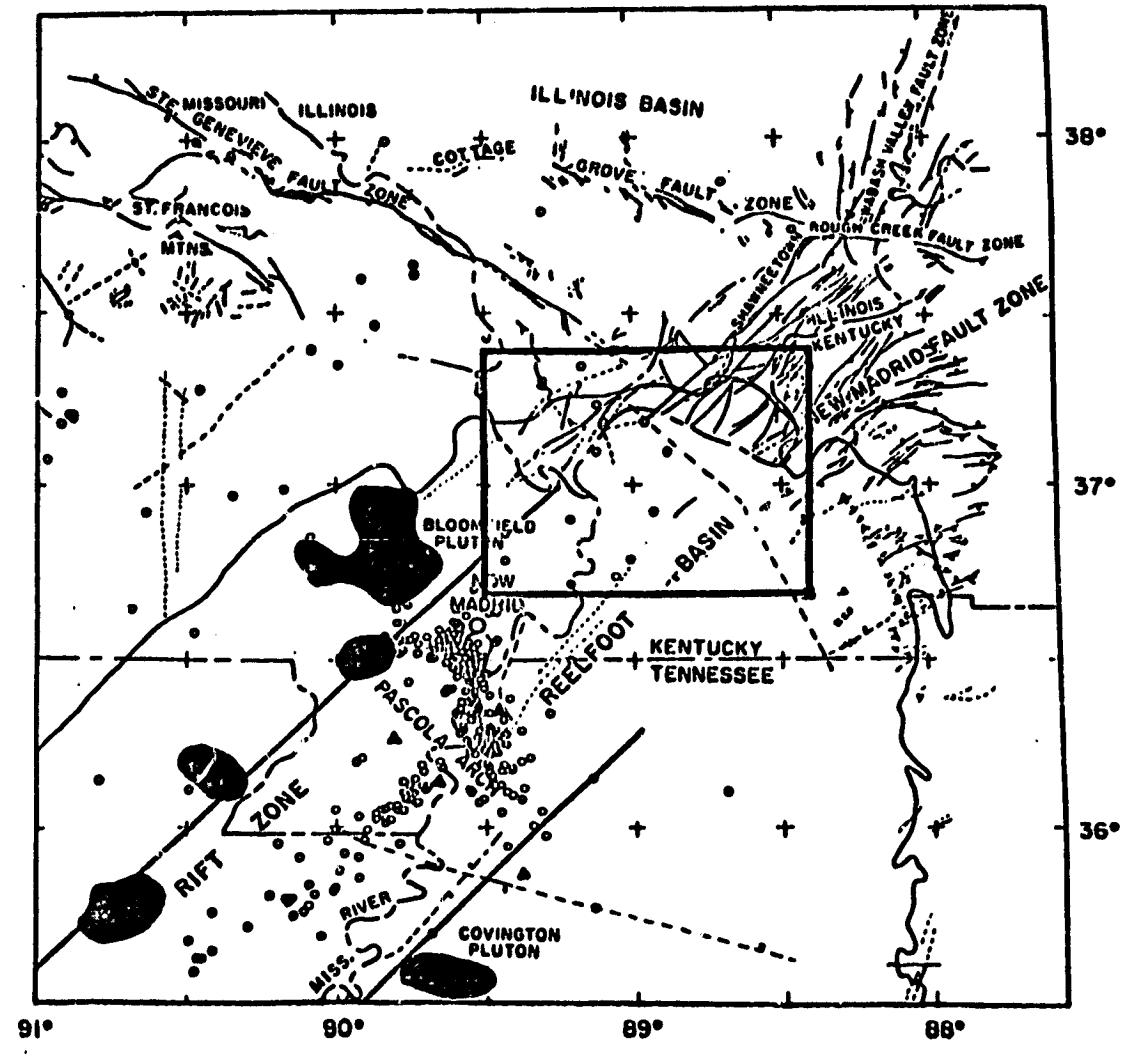


FIGURE 2.1. Seismotectonic map of the northern Mississippi embayment and surrounding regions (Hildenbrand and others, 1978). Study area is indicated by rectangle.

lower Mississippi River Valley. The overall pattern of Fisk (1944) is not reflected in the seismicity, and more recently Stearns and Zurawski (1975) have suggested that lineations may have often been interpreted as faults when they are actually joints or joint swarms. Stearns and Zurawski (1975) therefore used subsurface structural contours based on well data in conjunction with lineations, and their inferred fault pattern is considerably different from that of Fisk (1944), and shows somewhat better correlation with seismicity.

York and Oliver (1976) reviewed the evidence for Cretaceous and Cenezoic faulting in the central United States, and found all known faults of this age to lie in the northern tip of the embayment. The inferred fault pattern of Stearns and Zurawski (1975) likewise indicates numerous faults just inside the northern margin of the embayment. Zoback (1979) delineated two fault systems by seismic reflection profiling near Reelfoot Lake in northwestern Tennessee, about 75 km southeast of our study. Thus, the leveling data in this region deserve careful examination for evidence of modern fault activity.

LOCAL STRUCTURE

The dense Paleozoic faulting of the western Kentucky faulted area disappears at the northern edge of the embayment (Figures 2.1 and 2.2). This area has been mapped in detail and shown to be a northeast trending horst and graben fault complex with vertical displacements of 150-500 meters, associated with northwest-southeast lateral extension

of approximately 1.5 km (Hook, 1974). Normal faulting (with dips averaging 70-75 degrees) is predominant, although minor strike-slip motion is evident. The fault zone is now apparently quiescent, as almost no seismic activity has been detected there.

Considerable evidence, however, suggests the continuation of basement faulting beneath and/or into the embayment. In southernmost Illinois, subsurface mapping has revealed a complex of buried grabens and faults just within the northern boundary of the embayment. At least a part of these displacements occurred after the formation of the sub-Cretaceous erosional surface, with episodic activity in the Late Cretaceous, post Eocene, Pliocene, and Pleistocene (Ross, 1963a; Ross, 1963b).

SEISMOTECTONICS

The pattern of seismicity from 1974 to 1976 is shown in Figure 2.1 (Hildenbrand and others, 1978; Stauder and others, 1976). The major characteristics are the broadly-defined northeast trend within the embayment and a more diffuse pattern outside it. The existence of smaller-scale northeast- and northwest-trending lineaments is evident. Street and others (1974) indicate hypocentral depths for earthquakes in the New Madrid seismic zone ranging from 5 to 38 km, but the majority of the events shown in Figure 2.1 are probably between 2 and 10 km deep (Stauder and others, 1976).

The inset rectangle seen in Figure 2.1 indicates the locality for which leveling data will be presented (Figure 2.2). The center of the

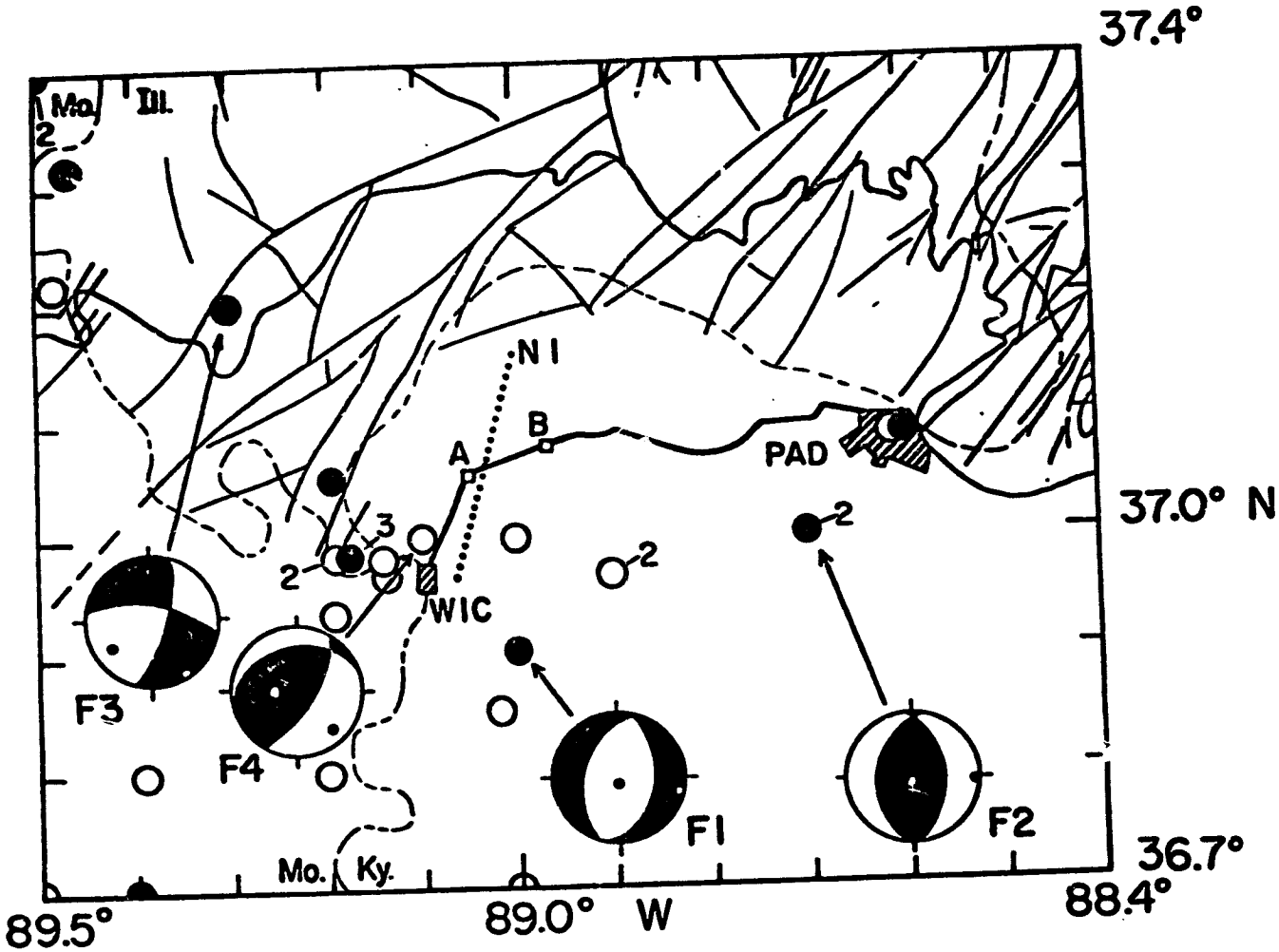


FIGURE 2.2. Seismotectonic map of the rectangular inset seen in Figure 2.1, showing the route of the two leveling surveys (1947 and 1968.7). The basemap with faulting is adapted from Stearns and Wilson (1972).

study area is approximately 75 km northeast of New Madrid, Missouri, the site of the large and damaging earthquakes of 1811-1812.

Figure 2.2 shows the seismicity in the study area for two different time periods. Solid circles represent events which occurred between 1947.5 and 1968.7 (listed in Table 2.1a), the dates of first-order leveling surveys along the route between Wickliffe and Paducah, Kentucky. Open circles represent events for 1928-1947.5 and 1968.7-1972 (listed in Table 2.1b), time periods outside the leveling interval. If non-instrumental historical seismicity were included in Figure 2.2, the greatest activity would be at Cairo, Illinois (37.0° N lat, 69.8° W long). For 1855-1972, 29 events have been reported at Cairo, including one damaging event in 1883 (intensity VI-VIII). The sharp clustering at Cairo since 1855, however, may reflect population bias, in absence of instrumental locations before 1928.

Although thrust, normal, and strike-slip focal mechanisms are all present in the area shown by Figure 2.2, six of the eight nodal planes (four focal mechanisms) strike north-northeast between 356 and 031, a span of 35 degrees. The majority of nearby focal mechanisms (within about 100 km of Figure 2.2) indicate thrusting, as noted by Street and others (1974). If the strike-slip event in southern Illinois can be associated with movement along the trend of the New Madrid fault zone, the sense of diplacement is consistent with east-west compression. Composite focal plane solutions in Arkansas and Missouri show a significant compoment of right-lateral fault motion indicative of EW compression in the central part of the embayment (Herman and Canas, 1978). The stress pattern indicated is not simple, but there is apparently a significant component of ESE-WNW compression in the

TABLE 2.1a. Earthquakes in area near Cairo, Illinois (Figure 3)
Time interval: 1947-1968.7

Date (yr.mo.da)	Lat (N)	Lon (W)	Depth (km)	Intensity	Magnitude	Reference
1947.01.16	36 59	89 11	-	II-III	(3.2)	a,b
1953.05.06	36 59	89 11	-	III	(3.4)	a,b
1953.05.15	36 59	89 11	-	III	(3.4)	a,b
1957.03.26	37 05	88 36	-	IV	(3.8)	a,b
1958.01.27	37 03	89 12	-	V	(4.2)	a,b
1962.02.16	37.0	88.7	25	III-IV	(3.6)	b,d
1963.01.31 (F1)	36.9	89.0	-	-	3.0	b,c
1963.05.02	36.7	89.4	-	-	3.1	a,b
1963.08.02 (F2)	37.0	88.7	18	V	4.0	b,c,d
1965.08.13	37 19	89 28	33	IV	3.2	a,b,d
1965.08.14 (F3)	37.2	89.3	38	VII	3.8	b,c,d
1965.08.15	37 22	89 28	16	V	3.5	a,b,d
1965.08.15	37 24	89 28	-	V	3.4	a,b

(F1): focal mechanism of Street and others (1974) (see Figure 3)

(3.2): magnitude estimated by Nuttli (1974)

a: Docekal (1970)
b: Nuttli (1974)
c: Street and others (1974)
d: Stearns and Wilson (1972)

TABLE 2.1b. Earthquakes in area near Cairo, Illinois (Figure 3)
Time interval: 1928-1947 and 1968.7-1972

Date (yr.mo.da)	Lat (N)	Lon (W)	Depth (km)	Intensity	Magnitude	Reference
1928.04.15	37 17	89 32		IV	(3.8)	a,b
1930.08.29	37.9	89.0		V	-	d
1930.09.03	36 58	88 54		III	(3.4)	a,b
1930.09.03	36 58	88 54		III	(3.4)	a,b
1931.04.06	36 51	89 01		IV	(3.8)	a,b
1933.10.24	37 17	89 32		III	(3.4)	a,b
1934.08.19	36 56	89 12		VI	(4.7)	a,b
1934.08.19	36 59	89 09		II-III	(3.2)	a,b
1936.08.02	36.7	89.0		III	(4.1)	a,b
1936.12.20	37 17	89 32		II	(3.0)	a,b
1939.04.15	36 48	89 24		III	(3.4)	a,b
1940.02.04	37 13	89 29		III	(3.4)	a,b
1940.05.31	37 05	88 36		V	(4.2)	a,b
1940.10.10	36.8	89.2		II-III	(3.2)	a,b
1941.10.21	36 58	89 08		IV	(3.8)	a,b
1941.11.22	37 17	89 32		II-III	(3.2)	a,b
1942.08.31	36 59	89 11		IV	(3.8)	a,b
1942.11.13	36 59	89 11		IV	(3.8)	a,b
1970.12.24	36.7	89.5		IV	3.6	b
1972.06.18 (F4)	37.0	89.1		III	3.2	c

see TABLE 1A. for explanation

embayment proper, in contrast to the Paleozoic SE-NW extension seen in the adjacent western Kentucky faulted area.

LEVELING DATA

The route between Paducah and Wickliffe (Figure 2.2) has been surveyed twice by the National Geodetic Survey using first order leveling. The first survey was made in 1947 and the second in 1968. Elevation changes computed from these leveling surveys (relative to Paducah) are shown in Figure 2.3. Also shown is the topography along the route. The standard error for the apparent elevation change (Brown and Oliver, 1976) at Wickliff relative to Paducah is $\pm$ 15 mm . About 60 percent of the bench marks along this route are in concrete posts.

The most striking feature in Figure 2.3 is the steep gradient between benchmarks A and B (31 $\pm$ 6 mm in 9 km) corresponds to a net tilt of 3.4 $\pm$ 0.7 X 10-6 rad. This tilt is about an order of magnitude larger than most regional tilts in the eastern United States (Brown and Oliver, 1976), and is definitely larger than random measurement error.

DISCUSSION AND INTERPRETATION

The segment of the route between A and B lies between the trends of the Big Creek fault zone (Fisk, 1944) and the western Kentucky faulted area (Hock, 1974)(Figure 2.2). It is possible, in light of the evidence presented thus far, that the movement indicated by Figure 2.3

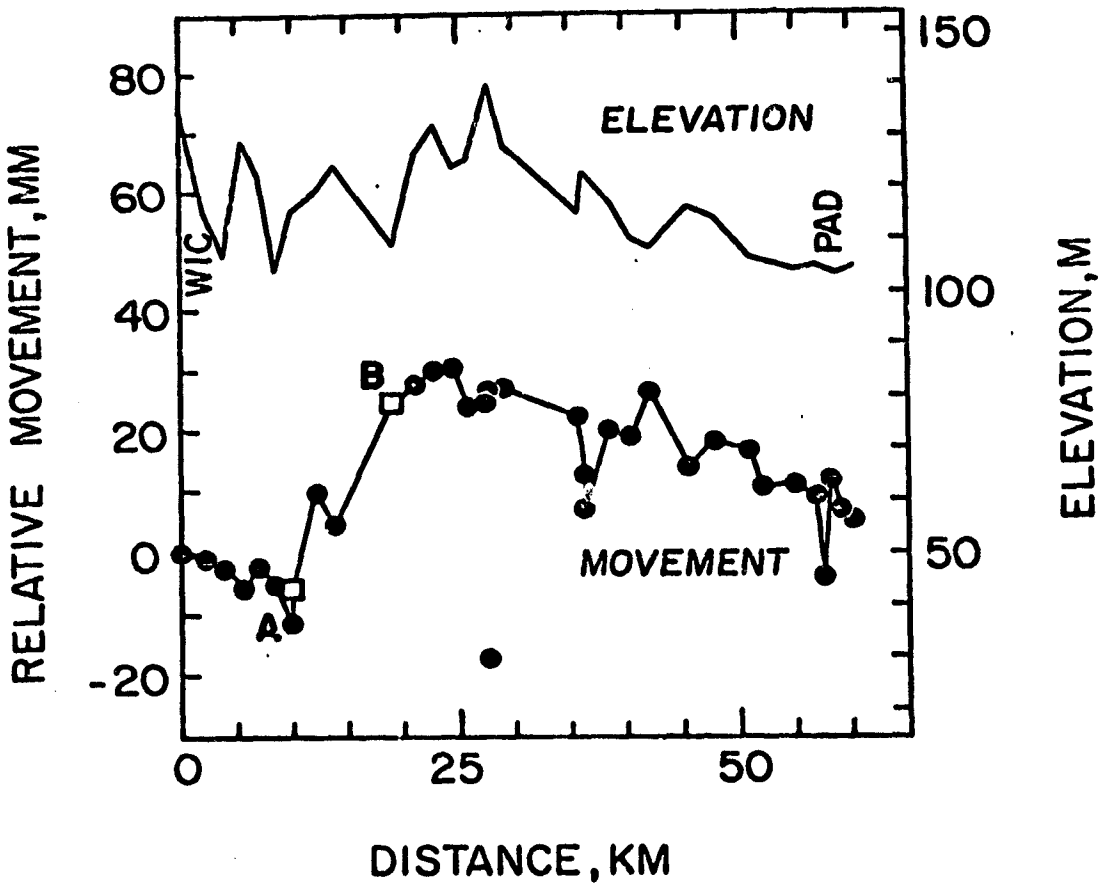


FIGURE 2.3. Plot of the relative vertical benchmark movement from Wickliffe, Kentucky (WIC) to Paducah, Kentucky (PAD). Elevation along the leveling route is also shown. Two bench marks of particular interest are labelled "A" and "B".

is fault related. However, possible errors and various alternative explanations for the leveling observations must be considered.

Since both leveling surveys were double run (elevation differences between successive benchmarks are measured twice for each survey), it is highly unlikely that the observed tilting is due to blunders (mistakes by the field party). The tilting occurs over too short a distance and in too flat an area to be attributed to many of the possibile systematic sources of error (e.g., unequal refraction, tidal effects), especially considering the tendency for such errors to cancel each other when computing the elevation change.

Near-surface non-tectonic processes also appear inadequate to explain the observed tilting. There is no evidence of significant fluid withdrawal large enough to cause sediment compaction in the area. Oil or gas wells are not present in this part of Kentucky, and the observed elevation changes show no relationship to either water-level fluctuation (seasonal or long-term) or aquifer geometry (Davis and others, 1973). The region studied is not karstland, arguing against surface slumping due to cavern collapse (Gerlach, 1970). There is little correlation between relative movements and type of monumentation used along this line, arguing against significant systematic trends due to benchmark instability. Crustal loading (e.g. reservoirs) has been shown to cause appreciable downwarps, but no such cases of large artificial loads were found near the offset.

If non-tectonic mechanisms are inadequate to explain the observed elevation changes, as seems to be the case, it is reasonable to hypothesize that they may be due to fault movement. The sense of surface deformation (down to the west) would be consistent with:

(1) a westward dipping normal fault

(2) an eastward dipping thrust fault, or

(3) a strike-slip fault with a significant component of dip-slip motion of type (1) or (2)

The closest focal mechanism (F4 in Figure 2.2; Street and others, 1974) is about 10 km southwest of bench mark A. It is a thrust type , but that earthquake occurred four years after the second leveling survey. The next nearest mechanism (F1 in Figure 2.2; Street and others, 1974) is about 20 km south of benchmark A. It is a normal type, and occurred within the time interval between surveys. It is therefore difficult to associate the offset seen in the leveling data with one or the other type of fault. In each case, however, one of the nodal planes would be consistent with the sense of inferred surface deformation.

From a single leveling route it is not possible to determine the areal trend of surface deformation. However, the trends of the nodal planes of the focal mechanisms, the trend of the structural trough formed by the top of the Paleozoic, the trend of the Mississippi River as it bounds western Kentucky, and the trend of lineations defined by local stream patterns (Figure 2.4) are all NNE. Thus it seems likely that if the observed movements are due to fault activity, the responsible fault strikes NNE.

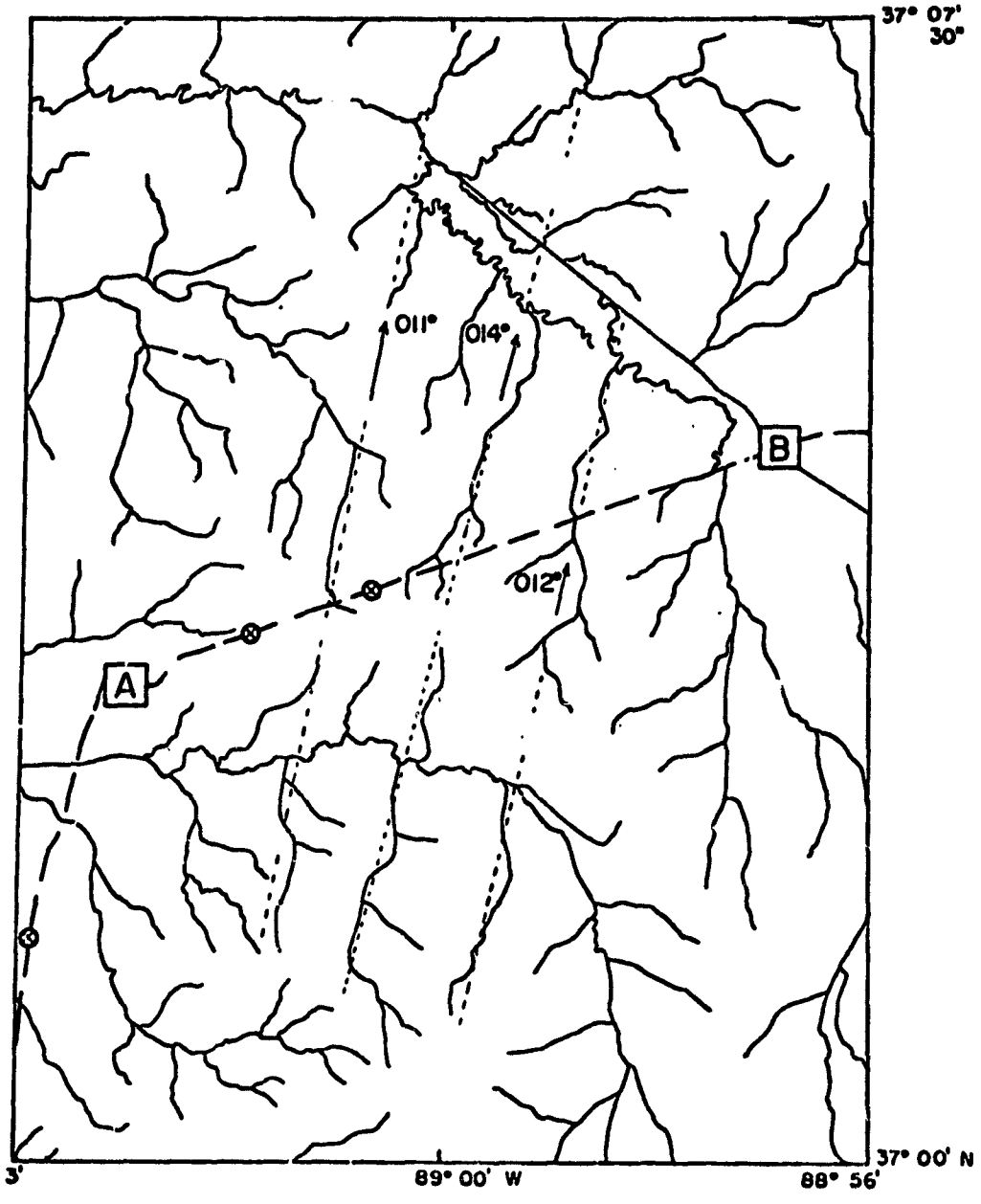


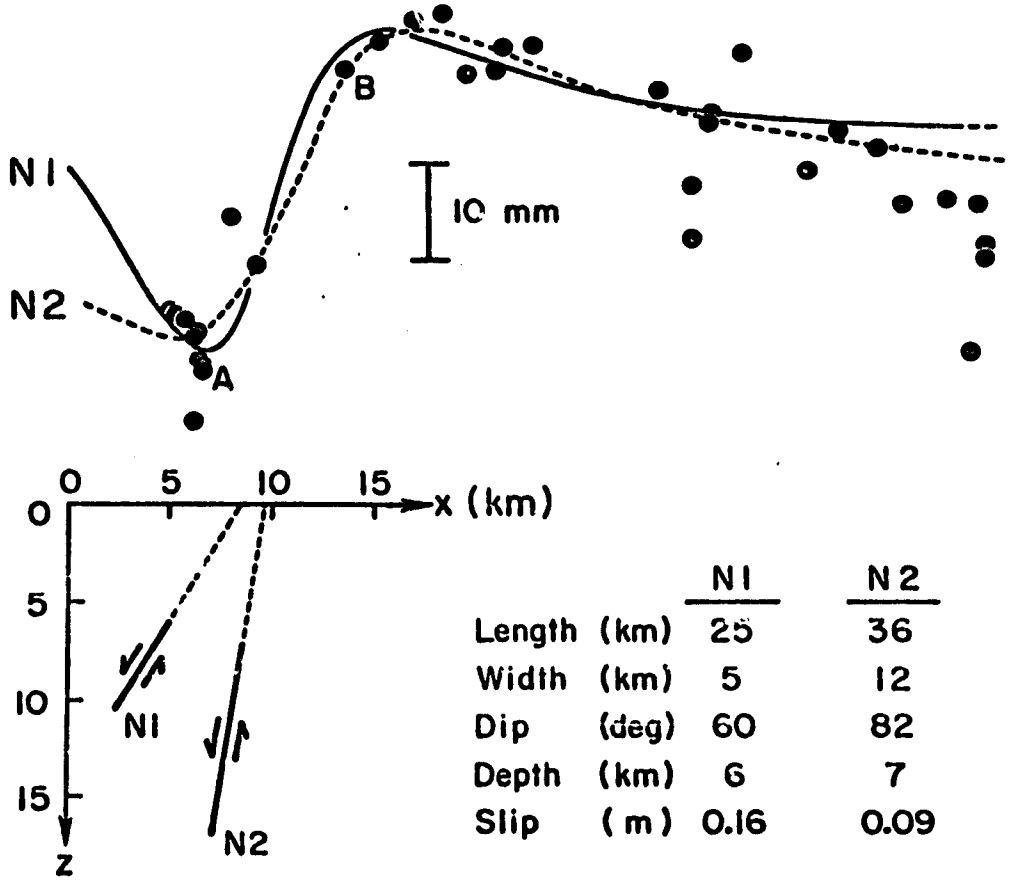
FIGURE 2.4. NNE-trending lineaments defined by stream pattern near bench marks "A" and "B" (other bench marks are shown by "X"). Streams were traced from U. S. Geological Survey 7 1/2' topographic quadrangle maps.

THEORETICAL FAULT MODEL

Savage and Hastie (1966) have given a convenient formulation for computing the surface deformation due to a dip-slip dislocation in an elastic half-space. By experimenting with various fault parameters, the pattern of apparent deformation was modelled with moderate slip along a westward dipping normal fault. Before making comparasons of the observed and modelled displacements, the bench mark positions were projected onto the profile P-P′ (Figures 2.2 and 2.5), on the assumption that the postulated fault trends at 015.

The pattern of deformation was best matched by normal faulting, although thrust faulting can also give a plausible fit. Two normal fault models which fit the data reasonably well are described and shown in Figure 2.5. In addition to the problem of non-uniqueness, there is the ambiguity introduced by the lack of data extending further to the west. However, with certain assumptions, order of magnitude estimates of fault parameters are possible. Assuming that the fault is in the upper crust and that its length is a few times its width, a slip of the order of 10 cm over an area of 100 km2 is required to match the magnitude of the observed movements.

If the deformation represents the strain release for one earthquake, the seismic moments implied by the two models in Figure 2.5 are 6 X 10+24 dyne-cm and 1 X 10+25 dyne-cm. Applying an empirical relationship between seismic moment and magnitude, a magnitude of 5.5-6.5 would be indicated (Kanamori and Anderson, 1975). The largest reported earthquake for this region between the times of leveling had an estimated magnitude of only 4.2. Thus, if the observations reflect



		N1	N2
Length	(km)	25	36
Width	(km)	5	12
Dip	(deg)	60	82
Depth	(km)	6	7
Slip	(m)	0.16	0.09

FIGURE 2.5. Surface displacement predicted by dip-slip on a buried normal fault. Dots indicate relative movements of bench marks (projected along the inferred fault strike shown in Figure 2.2) as measured by the leveling surveys. Displacement curves for two different normal fault models are shown. These two models give some idea of the tradeoffs possible in obtaining a reasonable fit to the observed movements.

fault motion, the deformation must have accumulated during multiple small events or as aseismic creep.

CONCLUSIONS

Elevation changes estimated from repeated leveling in western Kentucky may represent subsurface faulting. This conclusion, however, is based on a relatively limited data set, and the possibility of near-surface effects cannot be totally ruled out.

Modeling of a buried fault suggests a dip-slip of about 10 cm over a fault area of about 100 km^2. A normal fault is preferred, but thrusting is also possible. The equivalent magnitude for a single earthquake is larger than any earthquake experienced during the interval between surveys, suggesting incremental or aseismic displacement. It is recommended that these benchmarks be resurveyed as soon as possible to see if further changes have occurred since 1968.

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