Regenerate faults of small Cenozoic offset as probable earthquake sources in the Southeastern United States

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CONTENTS

ILLUSTRATIONS

TABLES

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Abstract

The principal style of Cenozoic faults and earthquake focal-mechanism solutions known along the eastern seaboard suggests that a domain undergoing northwest-southeast compression extends along the eastern seaboard between the continental margin and the front of the Appalachian Mountains. In the southeast, several mapped, northeast-trending zones of high-angle reverse faults, and numerous faults in isolated exposures, offset Coastal Plain deposits as much as 100 m; the youngest recognized offset is 0.35 m in probable Pliocene or Pleistocene surficial gravels. Movement has been progressive from at least late Early Cretaceous into the Cenozoic, with average offset rates of about 1 m/m.y. Northeast-striking reverse source mechanisms of various recent earthquakes indicate that reverse faulting is still active, and that such faults can be seismogenic.

The extent of the domain is inferred from (1) reverse faults known in South Carolina and along the Fall Line from Georgia to New Jersey, (2) earthquake source mechanisms, particularly in coastal New England, and (3) the broad distribution of early Mesozoic normal faults in the exposed Piedmont terrain and beneath the Coastal Plain and offshore. We suggest that, to a large extent, the reverse faults reuse parts of these older faults. We characterize the domain from the known fault histories, the existence of geometrically compatible earthquakes, and our expectation that seismogenic reverse faults are scattered essentially throughout the region no more than a few tens of kilometers apart. If correct, this expectation implies that the historic seismic pattern is not stationary over the long term.

The intensity X Charleston earthquake of 1886 occurred within the best-documented part of the Atlantic Coast domain, close to a Cretaceous-Cenozoic reverse fault and an inferred early Mesozoic fault zone with east-northeast trends. Some recent small aftershocks near these faults indicate similar reverse faulting, and the 1886 intensity pattern suggests a northeast strike for the main-shock source. We thus conclude that northeasterly-trending reverse faulting produced the 1886 earthquake. Many of the recent aftershocks seem to be occurring on northwest-trending structure, however, probably in response to temporary changes in the local stress field resulting from the strain release of the 1886 main shock. If this conclusion and our concept of the Atlantic Coast domain are correct, then earthquakes at least as large as 1886 Charleston should be possible in most parts of the domain.

The Atlantic margin was placed in compression approximately normal to its length sometime after successful rifting: certainly by latest Early Cretaceous, probably in the Jurassic, and possibly soon after extrusion of the 184-m.y.-basalt at Charleston. The source of the compression is presumed related to the continental margin, and may be related to the duration of Atlantic spreading, based on declining rates of reverse faulting.

Introduction

The generation of crustal earthquakes involves rupture and fault offset. Unless each seismic event is spatially unique, geologically recognizable fault offsets should develop through time as deformation proceeds. This argument must be qualified with considerations of rates of faulting, adequacy of the stratigraphic record, depths of activity and levels of exposure. It represents the geologic view, however, that seismicity represents deformation, and that to some degree the history of this deformation should be evident in the rock record. It should be worthwhile then, in considering the sources of earthquakes along the eastern seaboard, to attend to the recent geologic history of the region.

In the western third of the United States many large historic earthquakes can be attributed to movement on particular faults, especially where there is accompanying surface faulting or where detailed earthquake locations and focal mechanisms are obtained. Such conclusions are reinforced by confidence that continuing movement of the faults is reasonable, based on large Cenozoic offsets and convincing evidence in the topography and surficial geology of late Quaternary fault movement. It does not necessarily follow, however, that all seismogenic faults exhibit these traits. In considering seismicity along the eastern seaboard, it may be important to avoid drawing too close a parallel with such impressive western stereotypes as the strike-slip San Andreas fault, the normal faults of the Basin and Range, or the reverse faults of the California Transverse Ranges. The intraplate structures involved in the generation of earthquakes along the eastern seaboard may be much more modest features, at least in their Cenozoic incarnations. The fact that major fault scarps and other fault-generated topographic features have not been found suggests that this is true, although the role of Cenozoic tectonics in shaping the topography has yet to be resolved.

The MM X earthquake of 1886 at Charleston, South Carolina and the scattering of historic events along the Atlantic seaboard have not been accompanied by recognition of direct geologic evidence of the causative faults. The earthquakes along the eastern seaboard for which documentation exists have been as shallow as those expressed at the surface in the western U.S., so that the apparent absence of Quaternary indications of cumulative surface deformation may well result from low deformation rates. Despite recent improvements, available information is not sufficient to characterize fully the tectonic regime now in force along the eastern seaboard. Faults with modest Cretaceous and Cenozoic offsets do exist in the southeast, however, and it is important to examine their implications concerning the pattern of deformation and the generation of earthquakes, both there and along the whole of the Atlantic seaboard. From this viewpoint, we propose an hypothesis involving a consistent pattern of deformation and associated seismicity along the Atlantic seaboard, describe the bases for this hypothesis, and discuss the relation of the 1886 Charleston earthquake to it.

This paper is a synthesis of data from many sources, and involves a conviction founded in California experience that most earthquakes are related to specific faults. A major debt is due the work of, and many discussions with, W. L. Newell, R. B. Mixon, and D. C. Prowell of the U.S. Geological Survey, whose work on the Stafford and Belair fault zones stimulated the present discussion.

The hypothesis

The available geologic and seismologic evidence leads us to the following hypothesis. An Atlantic Coast domain of northwest-southeast compression extends from Georgia to Canada between the Appalachian Mountains and the edge of the continent. Within this domain, movement on scattered, northeast-trending reverse faults has been underway for at least 100 m.y., but in that time has accumulated offsets on individual faults of no more than about 100 m. Movement of these faults is a reasonable source of seismicity in the domain. Such faults exist in the Charleston meizoseismal area, and are a likely source of the 1886 earthquake. The reverse faults tend to reuse preexisting faults and other discontinuities, particularly the normal faults formed during early Mesozoic rifting. This association and earthquake focal-mechanism solutions permit extropolation of the domain beyond the presently recognized extent of Cenozoic reverse faulting in the southeast. Given the inferred extent of the domain and occurrence of the 1886 earthquake on a northeast-trending reverse fault near Charleston, Charleston-type earthquakes should be possible in most parts of the Atlantic seaboard.

With the exception of some Cretaceous-Cenozoic reverse faults and earthquake focal-mechanism solutions indicating northeast-trending reverse faulting, none of the components of this hypothesis are strictly provable with present information. Together, however, they form a coherent and reasonable whole that incorporates the principal style of Cenozoic faulting known to have occurred in the region.

Cenozoic reverse faults in the southeast

Recent work in the southeastern U.S. has demonstrated that northeast-trending reverse faults with modest Late Cretaceous and Cenozoic displacements exist in the Atlantic Coastal Plain and Piedmont. In the most carefully studied examples, the Stafford fault zone in Virginia (Mixon and Newell, 1977) and the Belair fault zone in Georgia (Prowell and O'Connor, 1978), it has been demonstrated that Late Cretaceous and Cenozoic sediments of the Coastal Plain are offset by faults that can be mapped for many kilometers. Compilations and field checks of isolated fault exposures (York and Oliver, 1976; and Prowell, 1981) and recent work near Charleston, South Carolina (Behrendt and others, 1981) suggest that such faults are widespread in the southeast, and indicate a general consistency of fault geometries and movement histories throughout the region.

The geology of the southeast is not ideally suited to the recognition of Cenozoic faults. In the Piedmont the Paleozoic crystalline rocks of the Appalachian orogen and the inset early Mesozoic redbeds and basalts of the initial phase of Atlantic rifting are too old to record the timing of Cenozoic faulting. Recognition of Cenozoic age from the ruptures themselves may be impossible; even the early Mesozoic extensional faults are generally recognized only where red beds and basalts record their presence. Geomorphology has not yet proved fruitful in recognizing Cenozoic faulting, and outliers of Coastal Plain sediment and scattered patches of alluvium and coluvium of uncertain age are of only limited value. In contrast, the nearly flat-lying Upper Cretaceous and Cenozoic sediments of the Coastal Plain that unconformably overlap the crystalline rocks, and especially the unconformity at the base of the sedimentary section, provide excellent control on post-unconformity offsets[1]. However, the poor exposure and subtle stratigraphic differences within the Coastal Plain require careful and deliberate search to find and map those offsets. Where the Coastal Plain section is thick, it is likely that faulting at its base may degenerate upward into folding, so that no abrupt offset may exist near the surface.

Distribution and geometry

Isolated exposures of faults that cut Coastal Plain and surficial sediments are scattered within the Piedmont and Coastal Plain, particularly near the Fall Line (plate 1; Prowell, 1981). In the past, these post-unconformity faults were largely encountered and reported in the course of other investigations. They were almost always found exposed in cross section in isolated steep cuts that had been excavated by streams or man. Some of these exposures have since been obliterated by continued excavation, such as a small graben found in a clay pit in South Carolina (Inden and Zupan, 1975), or concealed by subsequent construction, such as a fault near the Calvert Street Bridge in Washington, D. C. (Carr, 1950).

These isolated faults are probably parts of more extensive faults and fault zones, as indicated by the careful mapping of two zones of post-unconformity faults near the Fall Line. In Virginia, a narrow, anomalously steep gradient of northeast trend on the unconformity beneath Potamac Formation sediments, discovered by R. B. Mixon during mapping of the Quantico Quadrangle (Mixon, Southwick, and Reed, 1972), led to recognition of the Stafford fault zone (plate 1). Surface mapping and shallow drilling documented the existence of several northeast-trending faults as long as 35 km, with vertical offsets on the unconformity of 15 to 100 m, down to the east. Together these faults form a northeast-trending zone at least 67 km long that passes through Fredericksburg, Virginia and reaches within 30 km of, and possibly through, Washington, D. C., (Mixon and Newell, 1977, 1978a and 1978b; Seiders and Mixon, 1981; Carr, 1950). Natural exposures of crystalline rock faulted against Coastal Plain sediments found in the course of field mapping, and relations in a trench excavated across the trace of the Dumfries fault, demonstrate that the faults have reverse offsets (Newell, Prowell, and Mixon, 1976; Mixon and Newell, 1978a).

Seven hundred kilometers to the southwest, near Augusta, Georgia, a clay-pit exposure of saprolitized phyllite faulted against the basal Coastal Plain sediments (Middendorf Formation) was discovered during work on the State geologic map (O'Connor and others, 1974). This stimulated regional mapping, shallow drilling and trenching that documented the Belair fault zone (O'Connor and Prowell, 1976; U.S. Geological Survey, 1977; Prowell and O'Connor, 1978; and Plate 1). The unconformity was shown to be cut by several northeast-trending reverse faults with individual lengths of 2 to 5 km and offsets on

[1] The Coastal Plain section is everywhere underlain by a major unconformity. For brevity, we refer to this, and to the surface of crystalline rocks in the Piedmont, as the unconformity, and refer to faults that postdate it as post-unconformity faults. In a general way this unconformity is equivalent to the post-rift unconformity offshore (Dillon, Klitgord, and Paull, 1981), however that lies below the basalt at Charleston and offshore, rather than above it (table 1).

the unconformity of 5 to 30 m, down to the west. Together these faults form a northeast-trending zone at least 24 km long.

The work on these two widely separated fault zones shows that faults with significant length and modest Late Cretaceous and Cenozoic displacements do exist in the southeast. Even for these well mapped examples, however, the full lengths of the fault zones remain unknown. The Stafford and Belair fault zones have been mapped principally in the vicinity of the Fall Line, which they cross obliquely, because there they involve a thin Coastal Plain section. Along strike in either direction the limitations of the Piedmont and the thicker Coastal Plain section make recognizing and mapping the faults more difficult.

In thicker Coastal Plain section about 25 km southeast of the Stafford zone in Maryland, Jacobeen (1972) has mapped the Brandywine fault zone using seismic profiling (plate 1). During exploration for a gas storage site, reflection profiling across a local 30-m structural high on the lower Tertiary Aquia formation revealed two en enchelon reverse faults. These trend northeastward, and offset the unconformity, west-side-down, along a mapped length of 17 km. The vertical offsets across the two faults are 30 m and 50-60 m at the unconformity. These offsets decrease upward through the lower part of the Coastal Plain section, and are expressed largely as folding in the overlying Eocene and Miocene sediments.

Seismic profiling designed to search for this kind of structure has also revealed reverse faults of modest Cretaceous and Cenozoic offset in the Charleston meizoseismal area and off the South Carolina coast (plate 1). Onshore, the eastnortheast-trending Cooke fault offsets the base of the Coastal Plain section 50 m, down to the southeast, and shows decreasing offset upward into the lower Tertiary section (Behrendt and others, 1981a). Offshore, the Helena Banks fault has been traced for 35 km, and probably 70 km, on an east-northeast trend (Behrendt and others, 1981b). It offsets the base of the sedimentary section about 20 m, down to the southeast, and shows decreasing offset upward, with the shallowest recognized horizons being warped rather than sharply faulted.

The mapped faults and isolated fault exposures throughout the Coastal Plain and Piedmont have generally consistent geometries (Prowell, 1976, 1981, and written communication, 1979; Howard and others, 1977). The fault zones are probably complex in detail (Mixon and Newell, 1978a, p. 19; Prowell, written communication, 1979), and isolated parts may not be representative of the geometry of the zone as a whole. In a chance exposure it is impossible to determine whether a main or subordinate fault is involved; only where the fault zone is mapped for some length can the overall geometry be confidently established. Despite this, the faults, as now known, are generally northeast trending, high-angle reverse faults of small displacement. The dominant strike of individual faults is northeastward, as is the trend of the mapped zones, although the strikes range from northwest through east (fig. 1). Angles of dip range from 34 to 90°, but are generally high. Directions of dip are variably northwest and southeast, with no regular pattern evident. Vertical separations across individual faults are small, ranging from 0.2 to 100 m, and depend in part on the age of the offset horizon.

Those faults with upthrown hanging walls are considered reverse faults, although some may have strike-slip components as well. Evidence that the vertical separations represent actual displacement, or at least the principal component of fault slip, is of two kinds. Almost all the known northeast-trending faults show reverse separation (fig. 1), which is not likely for the dip components of strike-slip faults. Secondly, where well studied, the faults show local evidence of dip slip. Slickensides on both the Belair fault (Prowell and O'Conner, 1978) and the Dumfries fault of the Stafford zone (Mixon and Newell, 1978) reflect primarily dip slip. Faults of both the Stafford and Belair zones exhibit drag of Coastal Plain sediments up against the upthrown block (Newell, Prowell, and Mixon, 1976; U.S. Geological Survey, 1977).

At least a minor component of strike slip exists on some of the faults. On the Belair fault, Prowell and O'Conner (1978) report both slickensides at about 26 degrees to the dip direction and subsidiary splay faults that slightly offset kaolin clasts in the Tuscalosa Formation in a left-lateral sense. In the Stafford zone, some slickensides rake steeply southwest in the fault plane, and structural analysis of a trench exposure indicates that reverse and normal faults subordinate to the Dumfries fault are rotated progressively clockwise with increasing relative age (Newell, Prowell and Mixon, 1976). In contrast to the Belair fault, these imply a right-lateral component of slip across the Dumfries fault, at least in Potomac Formation time (latest Early Cretaceous).

Dominant strike slip seems unreasonable, however, although evidence is elusive because of the small displacements and the lack of piercing points or steeply dipping planes in the Coastal Plain section to use as control. Lateral offsets in the older Piedmont crystalline rocks, such as the 23-km left-lateral offset of the Augusta fault across the Belair fault described by Prowell and O'Conner (1978), have no necessary bearing on the Cretaceous-Cenozoic faulting. The Cretaceous-Cenozoic faults in the Belair and Stafford zones do not show the branching and anastomosing pattern characteristic of strike-slip faults, but form, instead, an open-spaced, en echelon pattern of parallel faults.

The known distribution of Cretaceous and Cenozoic reverse faults extends along the Fall Line from Georgia to Virginia, and about 200 km southeast of the Fall Line on and offshore near Charleston, South Carolina (plate 1). The principal concentration near the Fall Line is probably apparent rather than real, however, a result of the geologic opportunity provided by the thin Coastal Plain section there. The discovery of reverse faults where looked for on- and offshore near Charleston, and their existence well east of the Fall Line in Virginia, suggest that more will be found elsewhere on the Coastal Plain and shelf as the search proceeds.

History of fault movement

Offsets of the several post-unconformity faults shown in figure 1 are recorded by sediments that range in age from latest Early Cretaceous to Pliocene or Quaternary (Prowell, 1981), a period of 100 million years. The Stafford zone displaces sediments that record much of this time range, but at most of the fault localities the sediments probably represent only a very limited age range. Most of the ages are poorly constrained, however, because of the difficulties in determining the age of isolated patches of sediment, particularly unfossiliferous alluvium and colluvium.

In the more complete and well studied records, the offsets across particular faults decrease with decreasing age of correlative horizons, indicating progressive movement through time (fig. 2): the Stafford zone from late Early Cretaceous through probably Pliocene or Pleistocene, the Brandywine zone from late Cretaceous through Miocene or younger, the Belair zone from late Cretaceous through at least Eocene, and the Cooke fault from pre-Late Cretaceous through at least Paleocene. The Helena Banks fault also shows progressive movement, but the identity of the offset sedimentary horizons is less certain. The isolated faults are not inconsistent with progressive movement through time, particularly given the uncertainties of age assignments and the likelihood that both main and subordinate faults are involved. There is no good indication of geographic variation in the timing of faulting, although the data are much too limited to prove uniformity.

The greatest detail about the history of fault movement is provided by stratigraphic relations across the Stafford fault zone. As indicated by detailed structures exposed in a trench across the Dumfries fault (Newell, Prowell and Mixon, 1976), offset was underway during accumulation of the uppermost Lower Cretaceous Potomac Formation. Later movement progressively offset the Paleocene Aquia Formation, the Eocene Marlboro Clay, possibly the Miocene Calvert Formation, and, in one gutter exposure on the Fall Hill fault, upland gravel of probable Pliocene or Pleistocene age (Mixon and Newell, 1978a, fig. 7).

More specifically, using vertical separations across the Brooke structure (Mixon and Newell, 1978a, fig. 4, p. 9), the movement history to achieve a cumulative 105 m of offset (including drag) is approximately 83 m in the 35 m.y. from the base of the Cretaceous Potomac Formation to the base of the Paleocene Aquia Formation (2.4 m/m.y.), 13.5 m in the 11 m.y. from then to the base of the Eocene Marlboro Clay (1.2 m/m.y.), and 8.5 m in the 54 m.y. since then (0.2 m/m.y.), assuming movement to the present (fig. 2). The data are insufficient to determine the exact configuration of the Miocene Calvert Formation, so that any offset is unrecognized. If the Fall Hill gravel is considered representative of Brooke offset as well, then 0.35 m of offset can be added in perhaps the past 2 m.y. The average offset rate through the whole of the Late Cretaceous and Cenozoic is approximately 1 m/m.y., but the data indicate that the rate has decreased over time.

Expectable Quaternary offset across an individual fault is so small, perhaps 1/2 to 1 m at most, that it would be difficult to resolve in most situations where Quaternary control is actually available, and essentially impossible elsewhere. The youngest displacement reported in the Stafford zone is of this general amount, 0.35 m on the base of a gravel of probable Pliocene or Pleistocene age. The 0.6 m of reverse offset at the base of high-level terrace gravels over the Brandywine zone (Prowell, 1981) may represent recent fault movement of similar amount there. The tectonic origin of even smaller offsets that have been seen in gravels along the Stafford zone is equivocal, simply because the offsets are so small. Great difficulty is thus presented where there is need to test directly whether or not any movement has occurred across individual faults in the past several hundred thousand years.

Seismicity

Available seismologic information indicates that some reverse faults are moving and that reverse faults are capable of generating at least moderate earthquakes. Although the presence and geometry of earthquakes can indicate that a fault is moving, an absence of historic earthquakes does not mean that a fault is geologically inactive, but only that it lacks historic movement. This fact, the relatively low frequency of earthquakes along the eastern seaboard, the scarcity of detailed studies, and inconsistencies in the data greatly limit application of seismologic evidence to the hypothesis.

The most compelling relation is presented by earthquakes along the Ramapo fault system, located west of New York City about 400 km northeast of the Stafford fault zone. This 100-km-long fault system borders the early Mesozoic Newark basin on the northwest, strikes northeastward, and dips 60° to the southeast (Ratcliffe, 1980). No Cretaceous or Tertiary sediments are present to record its behavior during the past 100 m.y., and the Quaternary sediment that is present is very young. The current behavior of the fault system is indicated by the source mechanisms of numerous small earthquakes in the area and by the pattern of their hypocenters (Aggarwal and Sykes, 1978; Yang and Aggarwal, 1981). The focal-mechanism solutions show subhorizontal to gently plunging pressure axes and at least one north- to northeast-striking reverse focal plane, and the hypocenters define a southeast-dipping zone. Thus, contrary to its extensional movement in the early Mesozoic, the fault system is now moving in reverse fashion, and is generating earthquakes at least as large as magnitude 3.3. Aggerwal and Sykes (1978) suggest that three MM VI and three MM VII earthquakes reported in the greater New York area in the past 250 years may well have occurred on the same system, although the rather broad distribution of recent earthquakes recorded in the area (Chiburis, Ahner, and Graham, 1979) raises the possibility that other structures may be involved as well.

A magnitude 3.8, MM V-VI earthquake occurred in 1973 near Wilmington, Delaware in an area of previous minor earthquakes, including one of intensity VII (Sbar and others, 1975). The earthquake occurred at shallow depth beneath the northeast-trending lower reach of the Delaware River, and produced an isoseismal pattern that was elongate northeastward along the Fall Line. A composite focal-mechanism solution determined from the main shock and five aftershocks indicates northwest-southeast compression, although the computed pressure axis plunges about 40 degrees. One nodal plane indicates reverse dip slip on a nearly vertical plane striking north-northeast, northwest side up. This plane (plate 1) was preferred by Sbar and others over the alternate nearly horizontal plane because of consistency with the northeast-trending Fall Line and other similar geologic trends and the implied uplift of the Piedmont relative to the Coastal Plain.

In South Carolina, a small, shallow earthquake near Trenton in 1976 suggests northeasterly striking reverse faulting at a site 40 km northeast of the Belair fault zone (Talwani, 1977; Talwani and Tarr, written communication, 1979). A focal-mechanism solution yields a subhorizontal pressure axis and northeasterly striking reverse focal planes (plate 1).

Recent work in the epicentral area of the 1886 Charleston earthquake (Tarr, 1977, Tarr and Rhea, 1981) has produced focal-mechanism solutions suggesting reverse faulting on both northwest and northeasterly trends (plate 1). A magnitude 3.8 earthquake occurred there on November 22, 1974, the largest event since the South Carolina seismic network was begun in 1973. This yielded a focal-mechanism solution indicating northeast-southwest compression with the pressure axis plunging 40-50 degrees (Tarr, 1977). Focal planes are subhorizontal and steeply southwest dipping, both with northwest strikes. Tarr and Rhea (1981) have divided the well-located earthquakes that have been recorded through October, 1979 into three groups (fig. 6) with internally consistent first-motion patterns. Composite focal-mechanism solutions for two of the groups (A and B) are similar to the 1974 solution. The steep, west-dipping focal planes are preferred by Tarr and Rhea because of the vertical pattern of hypocenters in northeast-southwest section. In contrast, the third mechanism (group C) has a pressure axis plunging 35° to the north, and a subhorizontal and a steep, north northwest-dipping focal plane.

Microseismicity has been observed at four large reservoirs in the southeast: North Anna reservoir, Virginia (Dames and Moore, 1976) and, in South Carolina, Monticello (Talwani, Restogi and Stevenson, 1980), Clark Hill (Talwani, Secor and Scheffler, 1975; Guinn, 1980), and Jocassee (Talwani, Rastogi,and Stevenson, 1980). For the most part, however, this seismicity is exceedingly shallow or geometrically variable. The microseismicity at North Anna (Dames and Moore, 1976), extends to depths of at least 4 km, and indicates compression oriented east-west to northwest-southeast. Composite focal-mechanism solutions have subhorizontal pressure axes and one focal plane orientation in common, which strikes northeasterly and dips to the northwest (an average of N25E, 60NW). Assuming that only one fault orientation is involved, this geometry is consistent with that of the westernmost strand of the Stafford fault zone (Dumfries fault), 35 km along strike to the northeast.

The coastal New England region southeast of the Appalachian Mountain highland also experiences small earthquakes that indicate generally northeasterly trending reverse faulting. Of the 18 earthquakes ranging from eastern Connecticut to Maine studied by Graham and Chiburis (1980), 5 of the 7 with reasonably good focal-mechanism solutions have subhorizontal pressure axes and north- to northeast-striking reverse focal mechanisms. For the two exceptions, in Rhode Island and southern New Hampshire, Yang and Aggarwal (1981) present focal-mechanism solutions with subhorizontal pressure axes and northeast-striking reverse focal mechanisms.

Reuse of old discontinuities

The Cretaceous-Cenozoic reverse faults may well reuse preexisting discontinuities in the crust. This possibility is raised by the general alinement of the reverse faults with the structural grain of the Appalachian orogen and the early Mesozoic basins. The early Mesozoic normal faults offer the best opportunity for reuse: they are properly oriented, common, the most recent preceeding major faults in the region, and, as products of crustal extension during Atlantic rifting, probably extend well into the crust.

The strongest evidence for reuse of early Mesozoic faults is the reverse movement on the Ramapo fault system indicated by recent earthquakes, because there the existence of the older faults is well documented (Ratcliffe, 1971 and 1980). The corollary demonstration from field relations that these faults have moved during the Cretaceous and Cenozoic is prevented by the absence of stratigraphic control.

The Cretaceous-Cenozoic reverse faulting of the Brandywine zone in Virginia also seems to reuse an early Mesozoic border-fault system, as discussed by Mixon and Newell (1977). The Brandywine zone is alined with faults of the Richmond basin (plate 1), and continuity of faults across the intervening 80 km is suggested by an alined east-dipping gravity gradient. Red beds of the Richmond basin probably extend northeastward beneath the Coastal Plain cover, as indicated by drill hole and geophysical evidence (Mixon and Newell, 1977), and were encountered in several holes near the Brandywine faults. There, the faults separate largely granitic basement on the northwest from largely redbeds on the southeast (Jacobeen, 1972). This suggests local presence of an early Mesozoic border-fault system with offset down to the southeast, as well as continuity with the western boundary of the Richmond basin. The Stafford zone is similarly alined with the Farmville Basin (plate 1; Mixon and Newell, 1977), but local evidence for the presence and dip direction of an early Mesozoic fault along the Stafford zone is lacking.

About 30 km southeast of the Brandywine trend, Newell (written communication, 1979) reports another northeast-trending structure between the North Anna and Potomac Rivers, across which the shallow Coastal Plain section is monoclinally folded, down to the southeast. This structure lies close to the buried southeastern margin of the Richmond-Brandywine tongue of early Mesozoic sediments inferred from aeromagnetics (Klitgord and Behrendt, 1979; Zietz and others, 1977; and plate 1). This relation could also reflect reuse of an early Mesozoic border-fault system. Similarly, the Cooke fault is adjacent to a northeast-trending early Mesozoic fault inferred principally from geophysical data, as discussed below.

The possibility of two other similar associations between buried early Mesozoic border faults and Coastal Plain faults has been raised by Daniels and Zietz as an adjunct to their considerations of basement structure. They suggest (1981) the presence of a major Piedmont-early Mesozoic boundary that trends westward beneath the Coastal Plain in western Georgia, and point out its possible association with the nearby Andersonville fault (plate 1). This fault and nearby possible faults also trend westward and vertically offset lower Tertiary sediments as much as about 30 m, south side up (Owen, 1963; Zapp, 1965).

Daniels and Zietz (1978) had earlier suggested that faults reported in Coastal Plain sediments near Kinston, North Carolina were related to the nearby northwest boundary of a buried early Mesozoic basin inferred from aeromagnetics (plate 1). These faults, proposed by Brown and others (1977), have vertical separations in Upper Cretaceous and lower Tertiary rocks that are largely up on the west, and form a northeast-trending en echelon pattern (the Graingers fault zone) with an inferred aggregate left slip of about 6.4 km.

A crude test of the significance of these spatial associations between Cretaceous-Cenozoic faults and early Mesozoic normal faults can be made by comparing the sense of vertical separation across the younger structures with the dip direction predicted from the older normal-fault relations. The early Mesozoic normal faults should dip toward the adjacent redbed basins, so that subsequent reverse movement on the same structures should be up on the basin side, the relation demonstrated along the Ramapo fault. This is the case for the Brandywine, Cooke, and Andersonville faults and the monocline in Virginia, for all of which there a is local basis for inferring the direction of dip of the early Mesozoic faults. Thus, if the structures are simple enough for this test to be applied, these four meet it, whereas the Stafford and Graingers zone do not. The Graingers zone is not considered to be dip slip by Brown and others (1977), so with present information perhaps it should not be included. Although the Stafford zone lies on strike with faults at the northwest end of the Farmville basin, there is no local control on the dip direction of any early Mesozoic faulting.

It is not necessary that reuse of old discontinuities be limited to early Mesozoic faults, nor have all early Mesozoic faults necessarily been undergoing reverse faulting. None of the exposed border faults can, with available evidence, be demonstrated to have undergone displacement similar to that of the Stafford zone, and many of the isolated fault exposures lack evidence concerning older faulting. Although the early Mesozoic faults may be preferred for reuse by the reverse faulting, other structural elements of the Appalachian orogen are also properly oriented and may be used as well.

Assuming that reuse of early Mesozoic normal faults is common, the general abundance of Cretaceous-Cenozoic reverse faults can be estimated. Early Mesozoic faults are known along the border of the exposed redbed basins, and similar faults presumably occur in association with the buried basins as well (plate 1). Klitgord and Behrendt (1979) find northeast-trending horst and graben structure to characterize the whole of the offshore continental margin beneath the Jurassic post-rift unconformity, based on interpretation of magnetic-depth estimates. These faults and basin boundaries, together with the locations of known Cretaceous-Cenozoic faults, have a typical lateral spacing of about 25 to 50 km, and show few longitudinal gaps larger than 100 km. Early Mesozoic faults may also exist in areas now devoid of redbeds. As a result, we would expect the distribution of Cretaceous-Cenozoic faults to be approximated by a system of northeasterly-trending lines spaced 25 to 50 km apart, with each line containing scattered gaps, some of which may be as long as 100 km.

Sierran Foothill Analogy

The western foothills of the Sierra Nevada in California (fig. 3) show similarities to the southeastern Piedmont in framework geology and Cenozoic tectonics that, by analogy, reinforce parts of the reverse fault hypothesis. Recent attention has been drawn to the foothills largely by the magnitude 5.7 Oroville earthquake in 1975, which involved normal faulting of several centimeters along a surface trace 3.8 km long, and concurrent plans for a major thin-arch dam in the region. In contrast to the recognized tectonic activity around the Sierran structural block (fig. 3), no active faults or seismic hazard had been anticipated in the foothills. The result has been much new work on both the earthquakes and late Cenozoic tectonics of the region, on which the present discussion is based (Akers and McQuilken, 1975; Bartow, 1980; Bufe and others, 1976; Cloud, 1976; Harwood, Helley, and Doukas, 1981; Lahr and others, 1976; Langston and Butler, 1976; U.S. Geological Survey Staff, 1978; and Woodward-Clyde Consultants, 1977).

Like the Piedmont, the Sierran foothills consist of old metamorphic and intrusive rocks cut by Mesozoic faults, in this case thrusts of the Foothills fault system. In contrast to the Piedmont, however, Eocene and younger sedimentary and volcanic rocks in the foothills are locally preserved along old, west-draining valleys. Evidence of Cenozoic faulting is retained by these younger rocks, whereas in the rest of the crystalline terrain it has been obliterated by erosion. Offsets in the younger rocks indicate that normal faulting along pre-existing faults has been underway since the Miocene, but has accumulated relatively small offsets. The Cenozoic offsets occur across thin gouge and shear zones that follow the older faults in the crystalline rocks. This implies that such zones elsewhere in the crystalline terrain may also represent Cenozoic faulting. Not all of the older faults have been reused as normal faults, however, nor is the whole of any one older fault necessarily involved in the younger faulting.

The late Cenozoic offsets along the Foothills system range from 1 to 200 meters, but away from the edge of the Basin and Range on the north the offsets do not exceed several tens of meters. Evidence that offset has been cumulative since the Miocene exists locally, but in some places movement seems to have begun later. Locally preserved, buried soils in colluvium record fault movements within the past 100,000 years. These paleo B soils, exposed in many exploratory trenches in various parts of the foothills, show offsets that range from barely perceptible to as much as 0.6 m. As an example of fault history in the Foothills system, offsets across the Poorman Gulch fault are shown in figure 2: 60-75 m in the past 23 m.y., 27 m in the past 4-12 m.y., and, in addition, apparent truncation of a paleo B soil.

The Foothills fault system has been marked by a scattering of small earthquakes and perhaps seven MM VI and three MM VII events in the past 125 years, including the MM VII Oroville earthquake (Cloud, 1976; Woodward-Clyde Consultants, 1977). In the context of the design and safety of the thin-arch Auburn dam, the fault system is considered capable of generating an earthquake as large as magnitude 6 1/2 to 7 (U.S. Geological Survey Staff, 1978).

The Sierran Foothills and southeastern Piedmont are far from identical, but there is enough similarity to warrant drawing instructive analogies about regenerate fault behavior, recognition of Cenozoic faults and seismogenic potential. The Foothills example suggests that much more evidence of Cenozoic faulting may be found in the scattered surficial deposits in the Piedmont, but that trenching probably will be necessary. Little other direct evidence of Cenozoic faulting may exist, because scarps cannot be expected to survive, and Cenozoic and early Mesozoic shear and gouge zones in the Piedmont may be indistinguishable. In the Foothills, and probably in the Piedmont as well, regenerate faulting occurs selectively along pre-existing faults, and despite relatively low average rates of offset, can be seismogenic.

Compressional domain along the Atlantic margin

The Cretaceous-Cenozoic reverse faults indicate that parts of the southeast have been undergoing northwest-southeast compression for at least 100 m.y. The youngest stratigraphic evidence indicates probable Pliocene or Quaternary movement, and earthquakes indicative of similarly oriented compression show that this regime is still in force. The dip of the faults to both the northwest and southeast requires that on a regional scale this compression be horizontal.

The extent of the domain subject to northwest-southeast compression, based on the reverse faults themselves, includes at least South Carolina and a band along the Fall Line from Georgia to Virginia. Earthquakes, together with the inferred reuse of early Mesozoic faults, suggest that much of the Atlantic margin from Georgia northeast through coastal New England is involved. Zoback and Zoback (1981), using in-situ stress measurements as well as earthquake focal mechanisms and Cenozoic faulting, define an Atlantic Coast stress province under northwest-southeast compression that has a similar extent. The apparent axis of compression is approximately perpendicular to the continental margin and to Appalachian structure and topography (Diment and Urban, 1981), and is approximately parallel to the early Atlantic spreading direction represented by the near-shore oceanic fracture zones (plate 1). Thus there is a geometric coherence along the whole of the passive U.S. continental margin that implies a single tectonic system.

The lateral extent of the domain is difficult to estimate. Given the early Mesozoic horst-and-graben structure (Klitgord and Behrendt, 1979) that characterizes the offshore continental margin, and the recent demonstration that reverse faults can be found offshore, we postulate that most or all of the offshore continental margin is involved. Northwest of the Fall Line, the distribution of the exposed early Mesozoic basins suggests that the domain extends at least as far as the eastern front of the Appalachian Mountains. West of the Appalachian front (and Hudson-Champlain lowland in the northeast) Zoback and Zoback (1981) find east-west to southwest-northeast compression, although control on the location of the boundary between the two provinces is poor.

To characterize the Atlantic Coast domain of northwest-southeast compression, we combine the history of reverse faulting in the southeast, the source geometries of some modern earthquakes there and in the northeast, and the expectation that reverse faults are scattered diffusely throughout the domain. This yields a domain of scattered faults that have been moving in principally reverse fashion for at least 100 m.y. At least a modest strike slip component is present on some faults as well, probably depending on fault orientation, and this makes inference of the trend of the compression axis only approximate. The Cretaceous-Cenozoic faulting probably follows early Mesozoic faults to a considerable extent. As suggested by analogy with the Foothills fault system, however, the regenerate faulting is probably selective, involving only some of the pre-existing faults, and then probably only parts of those. The faulting has proceeded at low average rates of only about 1 m/m.y., which over 100 m.y. has produced cumulative vertical offsets on individual faults of only about 100 m or less. Evidence from the Stafford zone and from the Cooke fault (discussed below) suggests a decreasing rate of offset with time, so that Quaternary rates may be even lower.

Reverse faults as earthquake sources

The reverse faults in the Atlantic Coast domain can be seismogenic, as indicated by the Ramapo activity and the existence of other earthquakes with northeasterly striking, reverse source mechanisms. Numerous earthquakes of MM V to VII and greater have occurred in the Atlantic Coast domain (Coffman and von Hake, 1973) for which structural explanation is needed, although sources other than the northeast-trending reverse faulting could be involved as well. How large the earthquakes from the reverse faults can be is unknown. The faults are small by western U.S. standards, almost inconsequential, although no larger faults with a history of Cenozoic movement have been found. Analogy with the Foothills fault system in California suggests, however, that relatively small faults can generate earthquakes at least as large as MM VII. Aggarwal and Sykes (1978) suggest that the Ramapo system is capable of generating at least MM VII earthquakes, and we suggest that the MM X Charleston earthquake occurred on a northeast-trending reverse fault, as discussed below.

Not all early Mesozoic faults need be potential earthquake sources. However the probable abundance of such faults, and the relatively close spacing of reverse faults demonstrated by recent work in Virginia, suggest that seismogenic reverse faults occur in most parts of the Atlantic Coast domain. The historic absence of earthquakes in some parts of the domain can simply indicate that 200 to 300 years is too short a period to sample the long term seismic behavior of the domain. The occurrence of Cretaceous-Cenozoic faults in some areas lacking historic seismicity, such as North Carolina, suggests this.

Of the alternative sources of earthquakes that have been proposed, the Appalachian decollement has the most direct relation to the reverse-fault hypothesis (Behrendt and others, 1981a; Seeber and Armbruster, 1980). If the implication of recent proposals (Cook and others, 1979; Harris and Bayer, 1979) is carried to the limit, such a decollement could extend throughout the Atlantic Coast domain. Although there is no geologic evidence that such a decollement has moved at all since the Paleozoic, some of the early Mesozoic extension could have reused such a decollement as a sole, and the reverse faulting might be a shallow expression of more recent decollement movement in a compressional regime. It is also possible, however, that the early Mesozoic faulting and/or the reverse faulting cut across and extended below any such decollement. Regardless, the existence of Cenozoic reverse faults and earthquakes with similar reverse source mechanisms and subhorizontal pressure axes indicate that reverse faults are producing earthquakes along the eastern seaboard.

1886 Charleston earthquake

The origin of the Charleston earthquake of 1886 is still uncertain, despite its relatively large size (Bollinger, 1977, and 1981). In the careful field investigation following the earthquake, no surface offset indicative of a source fault was found on the low swampy ground of the meisoseismal area (Dutton, 1889). The shallowest rocks in which the earthquake could have originated are buried beneath a kilometer-thick layer of Cretaceous and Cenozoic Coastal Plain sediments, which are nearly flat lying and lack major faults or other prominent local structure. As a result, information bearing on the specific structural origin of the earthquake must come largely from drilling and geophysics, viewed in the context of the regional setting.

Various origins have been proposed for the earthquake. The relative isolation of Charleston seismicity has tempted consideration of some unique, but unspecified, tectonic source. Charleston seismicity can be viewed as part of a northwest-trending zone of earthquakes that extends from Charleston to the Blue Ridge (Bollinger, 1973). Sykes (1978) argued that the earthquake was related to an old, northwest-trending crustal feature that had localized the development of the Blake Spur fracture zone when Atlantic spreading began, although no such feature is known. The Charleston earthquake has been attributed to stress concentration around or within mafic or high density rock masses, which are inferred from gravity and magnetics to occur beneath the Coastal Plain near Charleston (Kane, 1977; Long and Champion, 1977; McKeown, 1978). Taber (1914) proposed a buried, northeast-trending fault based on 1886 intensity data, and Bollinger (1981) reaches a similar conclusion. Northeast-trending Paleozoic and early Mesozoic faults in the exposed Piedmont 150 km to the northwest have led to suggestions that similar faults beneath the Charleston epicentral area may have been responsible. The recently proposed decollement beneath the Appalachian orogen has been called upon (Behrendt and others, 1981a Seeber and Armbruster, 1980). Tarr (1977), and Tarr and Rhea (1981), using focal-mechanism solutions for recent small earthquakes in the 1886 meizoseismal area, suggest that the Charleston earthquake resulted from reverse faulting in a regional stress field with maximum horizontal compressive stress oriented northeast-southwest.

Most of these hypotheses involve little or no information about local structure and the Cenozoic geologic history of the region. The presence of the 1886 epicenter within the best documented part of the Atlantic Coast domain suggests northeast-trending reverse faulting as the source of the 1886 earthquake. Evidence in the meizoseismal area for an early Mesozoic fault system and recent discovery of Cretaceous-Cenozoic reverse faulting there reinforce this view.

Meizoseismal area

The meizoseismal area of the 1886 earthquake lies immediately northwest of the city of Charleston, South Carolina, and is bisected by the Ashley River (fig. 4). It covers a slightly elongate area about 50 km long and 30 km wide, with the long axis oriented northeast-southwest. Bollinger (1977) draws a simple egg-shaped figure around the intensity X effects. The original isoseismal patterns of Dutton and of Sloan, although more intricate, are also elongate along a northeast-trending axis that is approximately coincident with a line connecting the two points of highest intensity (epicentrums of Dutton, 1889). The principal difference between Dutton's and Sloan's patterns--the reentrant at the northern end --results from their differing conclusions concerning whether the general absence of higher intensity indicators there was real or simply due to the few "objects" available to express the intensity (Dutton, 1889). Much of the meizoseismal area is low swampy ground, whereas a Pleistocene barrier beach in the northwest part causes that area to stand about 50 feet higher. Because much of the intensity evidence consisted of sand boils and damage related to liquefaction, local ground conditions may have distorted the intensity pattern.

Bollinger (1981) sought indications of the orientation of the causative fault for the 1886 earthquake from the intensity data. He concluded that the northeast elongation of the inner 1886 isoseismals suggests a northeast fault strike, inasmuch as the elongation of at least the innermost isoseismals for an earthquake tends to mimic the fault trend. Comparison of the broader 1886 isoseismal pattern with those of the 1954 Fairview Peak and 1971 San Fernando earthquakes led him to the same conclusion.

Structure

The Charleston meizoseismal area lies within a region, distinct from exposed Piedmont to the northwest, which is considered to represent an early Mesozoic extensional terrain (Gohn and others, 1978; Popenoe and Zeitz, 1977; and Daniels and Zeitz, 1978). Some local basins within this buried early Mesozoic terrain (Daniels and Zeitz, 1981), the exposed early Mesozoic basins in the Piedmont (King and Beikman, 1974), and similar structures offshore (Klitgord and Behrendt, 1979) are characterized by northeast-trending boundaries between the basin fill and adjacent basement.

Northeast-trending basement structure is also present in the meizoseismal area, based on the refraction results of Ackermann (1981). The surface of crystalline basement in the Charleston area (figs. 4 and 5 and table 1) defines a northeast-trending high with a moderately sloping southeast flank and an abrupt, 900 m downward step along its northwest margin. A redbed section is inferred to fill the deep on the northwest, the upper part of which extends across the high to be encountered in corehole CCC 3. The steep northwest margin of the basement high would thus represent an early Mesozoic normal fault or fault system, with displacement down to the northwest.

Above the lower Mesozoic redbeds and basalt, both refraction and reflection seismic profiling show the J horizon (table 1) in the area of figure 4 to be a nearly featureless surface that slopes seaward at about 7 1/2 m/km (Ackermann, 1981; Behrendt and others, 1981a). However, this surface is slightly offset, down to the east, in several places within the meizoseismal area (fig. 4), as demonstrated by a reflection survey designed to achieve high resolution in the upper few kilometers of section (Behrendt and others, 1981). Three principal reflectors are involved, B, J and K (table 1). The offsets are expressed in the seismic records largely as irregular monoclinal bends as wide as a kilometer or more, but these should represent faulting, at least at greater depth. Some of the offsets extend with decreasing magnitude up into the overlying Coastal Plain section, indicating continued movement at least into the early Tertiary, whereas others do not (fig. 4).

The east-northeasterly striking Cooke fault (fig. 4) appears in the seismic records as a steep reverse fault, down to the southeast, within a wider zone of associated deformation (Behrendt and others, 1981a; Hamilton, Behrendt, and Ackermann, 1981). This zone offsets B about 190 m, J about 50 m, K about 40 m, and a younger horizon about 30 m. This younger horizon lies in the uppermost Pee Dee formation (about 70 m.y.), based on Yantis and others (1981) and position relative to K (Hazel and others, 1977). The movement history to achieve a cumulative offset of 50 m at the top of the basalt is 10 m in the 12 m.y. between J and K (0.8 m/m.y.), 10 m in the 13 m.y. between K and the younger reflector (0.8 m/m.y.), and 30 m in the 70 m.y. since latest Pee Dee time (0.4 m/m.y.), assuming movement to the present (fig. 2). There is no documented stratigraphic evidence of offset in the Eocene and younger section, where irregular unconformities and near absence of post-Oligocene sediments complicate the record.

The 190 m offset of the B horizon poses a quandary, because 140 m of offset preceding extrusion of the 184-m.y.-old basalt places reverse faulting within the early Mesozoic extensional regime. A hiatus of about 90 m.y. separates the top of the basalt from the base of the overlying Coastal Plain section, during which time extensive erosion should have occurred. It is reasonable that the 140 m of fault displacement occurred during this hiatus, but that the resultant offset at the top of the basalt was planed off by erosion. The offset and 140 m difference in basalt thickness would now be evident only at the base of the basalt. Of the 50 m of offset now evident at the top of the basalt, 10 m predated the K horizon. This must largely or entirely postdate initiation of Coastal Plain sedimentation or it, too, should have been destroyed by erosion.

We can thus add to the movement history of the Cooke fault 140 m of reverse movement in the 90 m.y. hiatus above the basalt, which represents a minimum rate of 1.6 m/m.y. The long term average rate of offset for the fault is approximately 1 m/m.y., but like the Stafford fault, offsets across the several stratigraphic intervals indicate that the rate has decreased with time (fig. 2).

This structure in the Charleston meizoseismal area is the same kind as that characterizing the Atlantic Coast domain. The Cooke fault strikes east-northeastward, as indicated by Behrendt and others (1981a), lies close to an early Mesozoic fault of similar trend, and has a sense of offset consistent with reuse of the older structure. The offset is small, but has been cumulative through the Late Cretaceous and early Tertiary.

The interpretation that the 140 m offset of the B horizon across the Cooke fault postdates the basalt provides the closest control yet available on the timing of the reversal from extension in the early Mesozoic to the compression of the Atlantic Coast domain. The reverse faulting began sometime between 184 and 94 m.y. ago. Unless the early rate of movement was four times higher than that in the late Cretaceous, however, the shift must have occurred in the Jurassic, and possibly soon after extrusion of the basalt.

The spatial correlation between reverse faulting and the step in the basement surface in the meizoseismal area is only approximate (figs. 4 and 5). The two isolated fault localities with Cenozoic offset seem to lie directly above the basement step, rather than up dip from it, and the Cooke fault and the pre-Cenozoic fault lie southeast of any reasonable up-dip extension of a simple, steep border fault. If early Mesozoic faults are controlling the specific locations of the reverse faulting, then they must be present off the main basement boundary as well as along it. This is not unreasonable, as the border fault system may involve splays and subordinate faults that either are masked by the smoothing inherent in refraction work or involved post-redbed offsets too small to be resolved.

The basement fault and documented locations of Cenozoic faulting lie in the northwest third of the meizoseismal area, rather than near its center (fig. 4). This eccentricity could result from distortion by local ground conditions of the intensity pattern from a Cooke-fault or nearby source. It could also indicate that the source of the 1886 earthquake does not lie close to the Cooke fault, but about 10 km to the southeast instead. In this central part of the meizoseismal area, Ackermann's refraction work (1981) shows the basement surface to be smoothly inclined to the southeast at about 10 m/km. It is intriguing that 20 mm of differential vertical offset of the ground surface, east-side down, reported by Lyttle and others (1979) from comparison of 1963 and 1974 first-order level surveys, occurs along the railroad just

Table 1.--Major stratigraphic horizons and intervals near Charleston

K, J and B are principal seismic reflectors; J and B mark the tops of refraction intervals characterized by the velocities shown. Compiled from Behrendt and others, 1981a; Gohn and others, 1977; Hazel and others, 1977; Ackermann, 1981; Lanphere, 1981; Yantis, Costain, and Ackermann, 1981.

Age in m.y.	Horizon	Description
	--Surface--	Coastal Plain section, 1/2-1 km thick, principally Late Cretaceous to Oligocene, base at corehole CCC 1 is Cenomanian; K
82	--K--	correlated with a velocity contrast near the Campanian/Santonian boundary in CCC 1.
94		
	--Top of Basalt/J--	Major hiatus and unconformity
184		
	4.2-5.8 km/sec	Basalt over redbeds; at CCC 3 basalt is 257 m thick, underlain by at least 120 m of redbeds, which in places may reach J where basalt seems missing.
	--Top of Basement/B--	
	6.0-6.4 km/sec	Presumed crystalline basement predating Atlantic rifting; identity somewhat ambiguous, at least in places, as basalt within redbed section might produce similar reflections and velocities, and layered reflections are locally observed below B.

south of the southern 1886 epicentrum (fig. 4). This raises the possibility that surface deformation has been occurring along the axis of the 1886 meizoseismal area, perhaps from some form of afterslip or slow upward propagation of the main-shock offset through the Coastal Plain section.

Recent seismicity

Recent small earthquakes in the Charleston meizoseismal area, as relocated by Tarr and Rhea (1981), form a diffuse, north-trending zone about 20 km long and 10 km wide, and range in depth from near surface to about 15 km (figs. 4, 5 and 6). No single, simple source zone is evident in the three-dimensional pattern of hypocenters (fig. 6), although errors in location may have dispersed a tighter pattern of sources.

The earthquakes are considered aftershocks of the 1886 main shock (Tarr, 1977; Bollinger, 1981), and may indicate the depth of the main shock. Bollinger (1981) suggests that the 1886 intensity pattern permits a shallow focus, and there is no evidence requiring a deeper one. It is not necessary, however, that the recent earthquakes represent either the geometry or specific location of the 1886 source. Like the basement step and Cenozoic faults, the recent earthquakes are eccentric to the center of the meizoseismal area (fig. 4), and there is no means of tying them with certainty to the specific 1886 source.

The three composite focal-mechanism solutions determined by Tarr and Rhea (1981) yield conflicting source geometries, which prevent direct inference of a regional stress orientation. Two of the solutions yield pressure axes plunging 35° and 45° to the south-southwest, and one yields a pressure axis plunging 35° to the north. Combining the focal mechanisms and the three-dimensional pattern of hypocenters, Tarr and Rhea (1981) conclude that northwest-striking reverse faulting has occurred on two steeply dipping planes represented by hypocenter groups A and B (fig. 6). Group C, in contrast, represents east northeast-striking reverse faulting, down to the south, if the steep focal plane is selected. The latter is consistent with the geometry of the Cooke fault.

In order to further test the possible relation of these earthquakes to the northeast-trending basement step and the younger faults, a cross section (fig. 5) was drawn approximately normal to the structures and the axis of the meizoseismal zone. The C-group source (earthquakes 11, 14, and 27), in addition to yielding a source geometry consistent with the Cooke fault, lies about 15 km down dip from it. Behrendt and others (1981a) find a 70° dip for the Cooke fault above about 3 km; correlation of the C group source with the Cooke fault yields a similar, 60° dip.

Inspection of figure 5 and the stereographic presentation of hypocenters in figure 6 demonstrates that, within their location uncertainties, all the earthquakes except those in the southern group can lie on two northeasterly striking, northwest dipping surfaces, one passing through the C group and the Cooke fault, and one about 5 km above it. Although this organization would be consistent with northeast-trending structure, it would require ignoring the focal-mechanism results for group B.

The focal-mechanism solutions for earthquake groups A and B suggest that northwest-trending structures are present in the meizoseismal area. The hypocenters in these groups can be viewed as forming very steep, westnorthwest-trending zones (fig. 6). Northwest-trending mafic dikes are present in the region (Daniels and Zeitz, 1978), and Schilt and others (1981) found suggestion of basement structure where COCORP reflection-line 3 crosses the Ashley River. The configuration of the vibration lines does not permit a northeast strike for this feature if it has any length, whereas a northwest strike is possible.

One means of resolving the conflict in source mechanisms is to infer that the horizontal stresses are approximately equal (Zoback and Zoback, 1981) in the area in which the regional stress field has been perturbed by the strain release of the 1886 main shock. The northwest-southeast compression typical of the Atlantic Coast domain would not yet have been restored, and the geometries of continuing aftershock adjustments would be sensitive to both existing structures and the effects of preceeding adjustments.

Source of the main shock

The Cenozoic regional setting of the Charleston earthquake is one of scattered, northeast-trending reverse faults. Faults of this style have been documented west of Charleston near the Fall Line and to the southeast offshore, as well as in the 1886 meizoseismal area itself. This kind of fault in the Atlantic Coast domain can be seismogenic, and probably has produced earthquakes at least as large as MM VII. No other faults with a record of Cenozoic movement have been found in the area. Some of the recent small earthquakes in the meizoseismal area suggest that movement on east northeast-trending reverse faults is occurring, at least as part of the aftershock adjustments, and the 1886 intensity pattern suggests a northeast strike to the main-shock source. We thus conclude, in a fashion similar to that suggested by Bollinger (1981), that northeast-trending reverse faulting produced the 1886 earthquake, but that many of the recent aftershocks seem to be occurring on northwest-trending structure, probably in response to temporary changes in the local stress field resulting from the strain released in the main shock.

Exploration is still incomplete in the meizoseismal area, and the Cooke fault may only be representative of the structure there. The fact that the Stafford and Belair fault zones consist of several faults, and that other Cenozoic offsets exist in the meizoseismal area, make it reasonable that a system of northeast-trending reverse faults exists in the meizoseismal area. If the Cooke fault was not the 1886 source, a possibility raised by its position eccentric to the meizoseismal area, some similar nearby fault (or faults) probably was.

Discussion

Beginning with geologic structure and history, we conceive a tectonic behavior for the whole of the eastern seaboard based on the principal mode of Cenozoic faulting known in the region. We argue that an Atlantic Coast domain of northwest-southeast compression and resulting reverse faulting has existed since at least the early Cretaceous, and that reverse-fault movements in the domain can account for much of its seismicity, including the 1886 Charleston earthquake, although other mechanisms may be at work as well. In constructing the hypothesis, we have been guided by the expectation that geologic history is not capricious, and that the history of the passive Atlantic margin has been generally consistent in its details as well as in outline. Thus, for example, we have combined fault histories in Virginia and South Carolina with earthquakes in New Jersey and New England to obtain a coherent tectonic style and history for the whole domain.

The most important implication of the hypothesis is the possibility that earthquakes as large as the 1886 event are not limited to Charleston. If that earthquake did occur on a northeast-trending reverse fault, then similar earthquakes should be possible wherever such faults are present and still active. We infer that the reverse faults have a common history of movement extending from the Mesozoic through the present, and argue that they are scattered essentially throughout the region no more than a few tens of kilometers apart. The result is that Charleston-type earthquakes should be possible in most parts of the Atlantic Coast domain.

The potential for surface faulting in the domain is not clear, although the record along the Stafford zone, in particular, indicates that offset near or at the surface has occurred in the past. The thickness of sediment overlying hard rock is probably a key factor in determining whether or how the faulting will be expressed at the surface. Monoclinal folding rather than discrete faulting in the thick Coastal Plain section, for example, might account for the lack of surface faulting in the meizoseismal area at the time of the 1886 earthquake. Because the displacement expectable in a million years or less across a single fault is very small, it is unlikely that evidence of offset will be found along most faults that have moved in the late Quaternary. The absence of such evidence, however, except where the late Quaternary record is exceedingly clear, cannot demonstrate that offset has not occurred.

The northwest-southeast compression of the Atlantic Coast domain is an important element in the post-rift history of the Atlantic margin. Following the extension that resulted in Atlantic rifting about 190 m.y. ago., the trailing continental margin has sagged and accumulated a mantling wedge of sediment upon the depressed post-rift unconformity. In addition, at some time between successful rifting and the beginning of the Coastal Plain record about 100 m.y. ago., the continental margin was placed in compression approximately perpendicular to its length. The transition from extension to compression may have occurred soon after rifting was accomplished, based on our interpretation of Cooke-fault offsets. Ever since, the compression has driven small reverse faults, which we suggest are scattered throughout the Atlantic Coast domain.

The source of the compression is unknown, although the perpendicular relation between the compression axis and the Appalachian orogen, the continental margin, and the Jurassic axis of Atlantic spreading suggests that one or more of these is involved. A causal relation between the compression and something associated with the continental margin is implicit in our argument. The possibility that the compression began soon after rifting, and the decrease in rates of reverse faulting through time indicated by Stafford- and Cooke-fault offsets, suggest an inverse relation between the rate of faulting and the duration of Atlantic spreading.

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FIGURE CAPTIONS

Figure 1.--Strikes of post-unconformity faults in the southeastern Piedmont and Coastal Plain between eastern Georgia and Washington, D. C. Compiled from Prowell (1981), Behrendt and others (1981a), Mixon and Newell (1977), and Prowell and O'Conner (1978). The strikes of 41 individual faults are plotted, including those in the Stafford and Belair zones, and the strikes of the zones are shown by the zone names.

Figure 2.--Fault offset histories, showing age and displacement of marker horizons. Brooke structure of Stafford zone from Mixon and Newell (1978a), Belair fault zone from Prowell and O'Connor (1978), Poorman Gulch fault in the Foothill fault system from Woodward-Clyde Consultants (1977), Cooke fault from Behrendt and others (1981a), Hamilton and others, 1981, and see text discussions; dashed line shows minimum pre-Late Cretaceous rate of offset.

Figure 3.--Foothill fault system, points of known late Cenozoic offset, and setting within the westward tilted Sierran block. Details in Sierra Nevada foothills from Woodward-Clyde Consultants (1977, v. 1, fig. 34), and Jennings, 1977; eastern Sierran front and general setting from King and Beikman, 1974.

Figure 4.--Meizoseismal area of the 1886 Charleston earthquake. Isoseismals from Bollinger (1977) and Dutton (1889, plates 26 and 27): Dutton's and Sloan's isoseismals are their outermost complete lines, for which no values are given. Plotting on the modern base map is only approximate, and depends principally on distances along railroads and rivers. Basement surface from Ackermann (1981). Faults from Behrendt and others (1981a). Earthquakes through December, 1979 with location qualites A to C shown from Tarr and Rhea (1981), with numbers keyed to their table 1. Surface offset indicated by level surveys from Lyttle and others (1979). C - Charleston, AR - Adams Run, MP - Middleton Place, S- Summerville, J-Jedburg. See separate explanation.

Figure 5.--Northwest-southeast cross section through 1886 Charleston meizoseismal area. See figure 4 for section line. Basalt and basement surfaces from Ackermann (1981). Cooke fault from Behrendt and others, (1981a); J and B as in Table 1. Earthquake hypocenters from Tarr and Rhea (1981) with numbers keyed to their table 1; bars represent axes of location-error elipses projected into section.

Figure 6.--Stereogram of recent earthquake hypocenters (1974-1979) in the Middleton Place - Summerville zone, Charleston area, South Carolina. Shows hypocenters of earthquakes with location quality A to C through January 1979 and M2.9 event of December 7, 1979, from Tarr and Rhea (1981), with numbers keyed to their table 1; letters identify their three groups of hypocenters. Vertical view from altitude of 250 km; top of box at ground surface, bottom at depth of 10 km. Stereogram prepared by C. E. Johnson and P. T. German.

Explanation for figure 4

Earthquake epicenter
Isoseismals
    MM X of Bollinger
    Dutton
    Sloan
epicentrum
contour on basement surface
post-basalt faults
    fault correlated between seismic crossings
    fault crossings on seismic lines; solid where extends into
        Cenozoic rocks, open where does not

surface offset along railroad from comparison of level surveys
Clubhouse Crossroads core holes

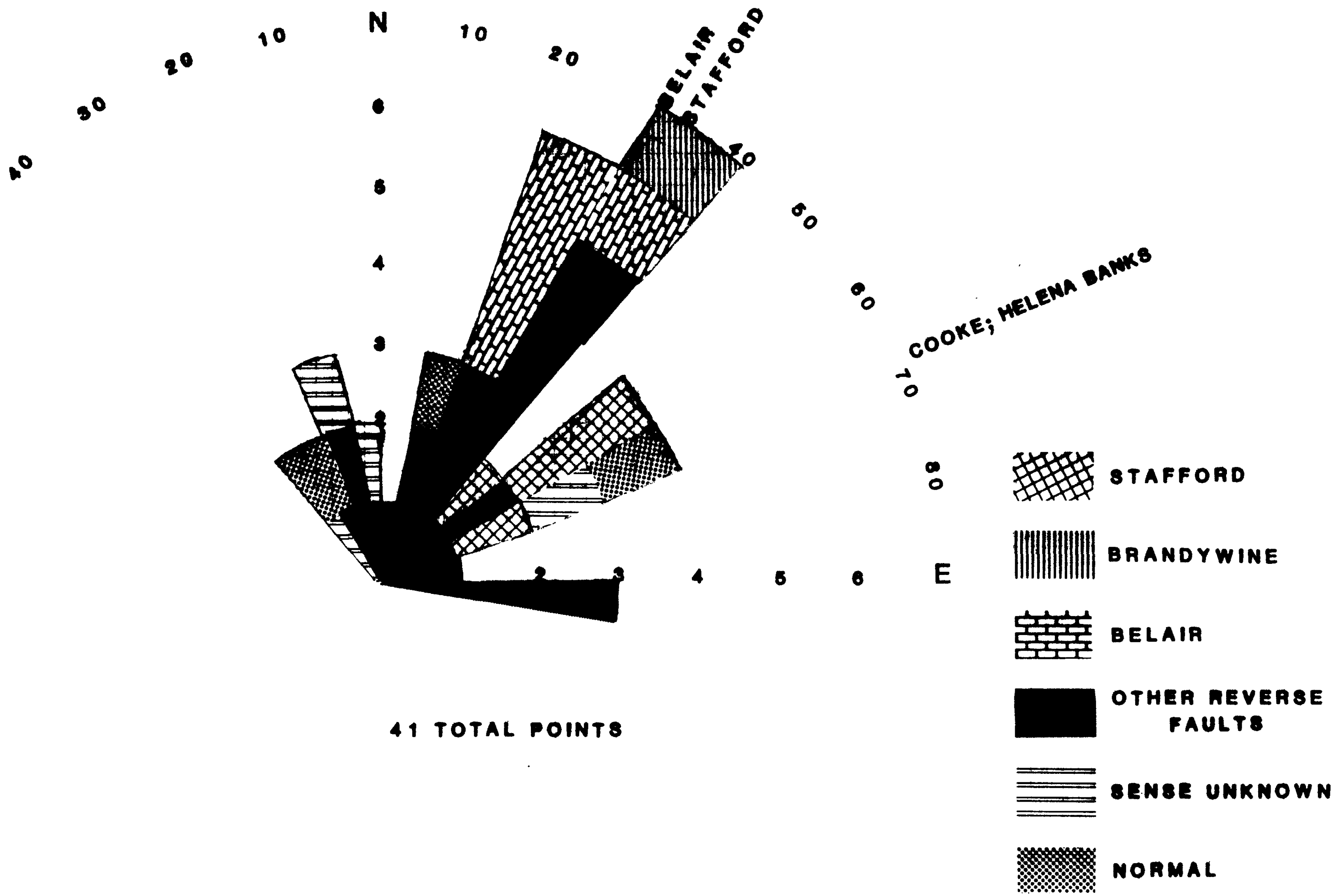

N
E
BELAIR
STAFFORD
COOKE; HELENA BANKS
41 TOTAL POINTS
STAFFORD
BRANDYWINE
BELAIR
OTHER REVERSE FAULTS
SENSE UNKNOWN
NORMAL

Figure 1

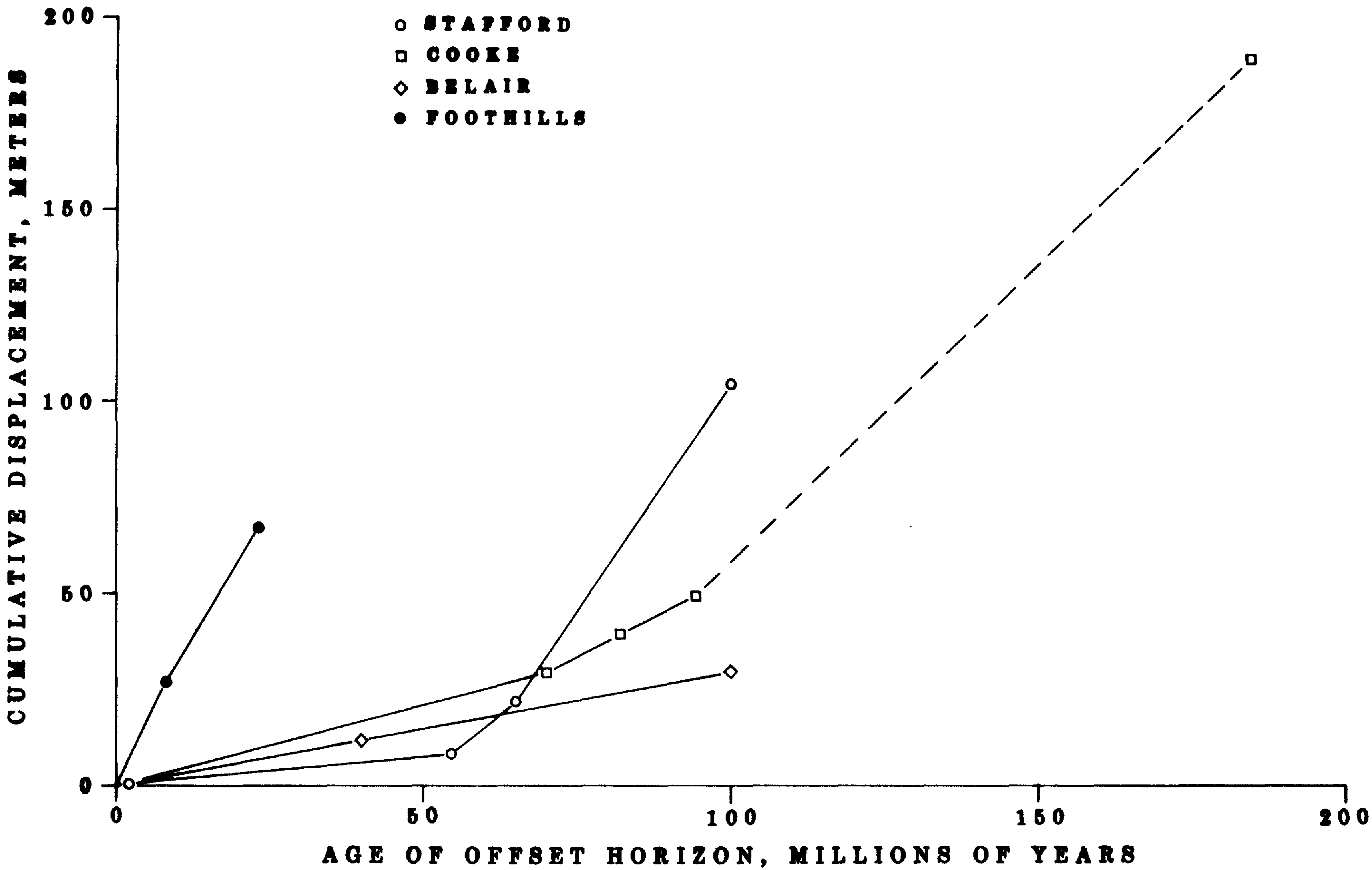

STAFFORD
COOKE
BELAIR
FOOTHILLS
200
150
100
50
0
CUMULATIVE DISPLACEMENT, METERS
0
50
100
150
200
AGE OF OFFSET HORIZON, MILLIONS OF YEARS

Figure 2

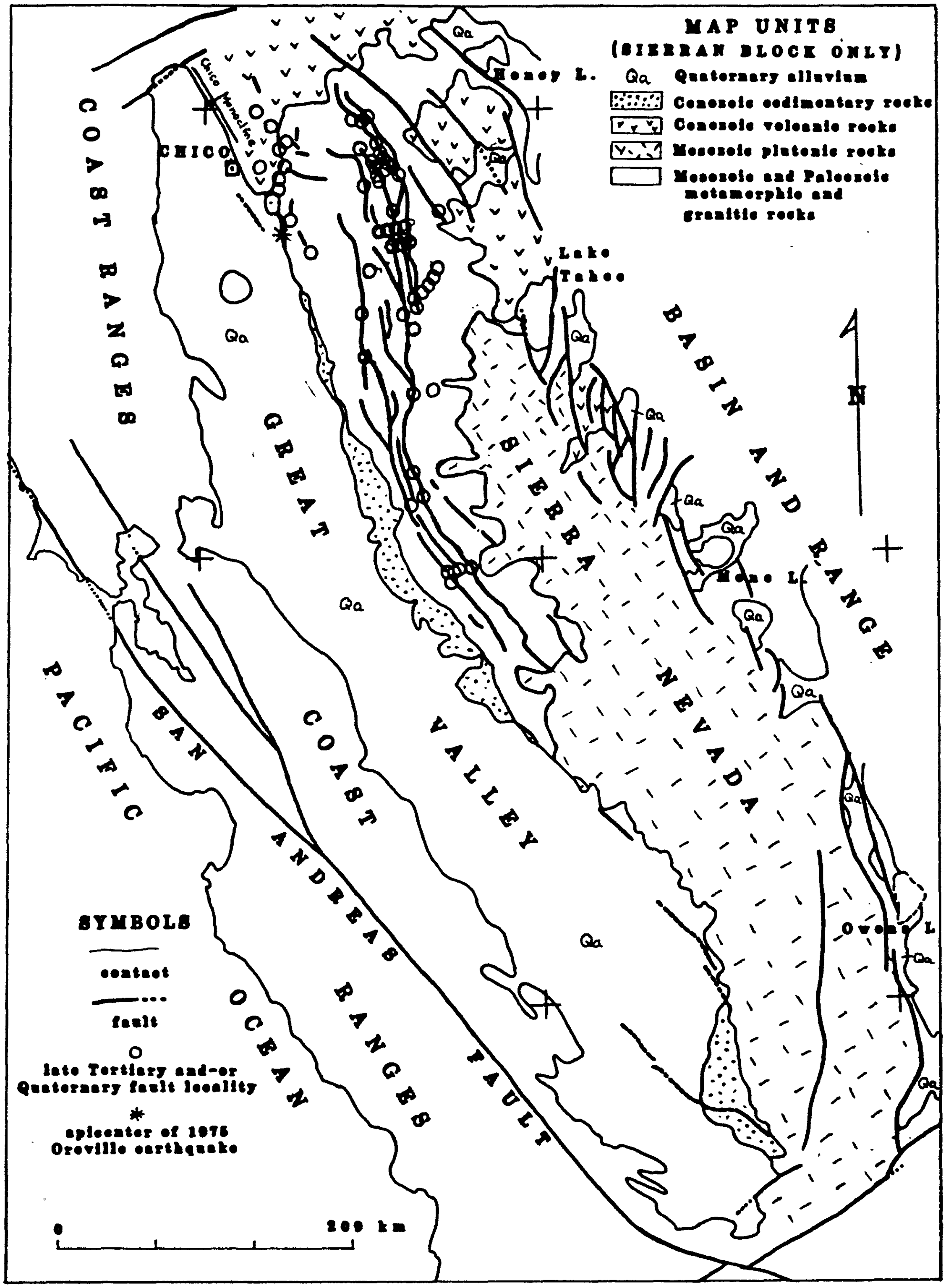

MAP UNITS
(SIERRAN BLOCK ONLY)
Qa Quaternary alluvium
Cenozoic sedimentary rocks
Cenozoic volcanic rocks
Mesozoic plutonic rocks
Mesozoic and Paleozoic metamorphic and granitic rocks
SYMBOLS
contact
fault
late Tertiary and-or Quaternary fault locality
epicenter of 1975 Oroville earthquake
0
200 km
N
COAST RANGES
CHICO
Chico Monocline
Honey L.
Lake Tahoe
Mono L.
Owens L
BASIN AND RANGE
SIERRA NEVADA
GREAT VALLEY
COAST RANGES
SAN ANDREAS FAULT
PACIFIC OCEAN
Qa

Figure 3

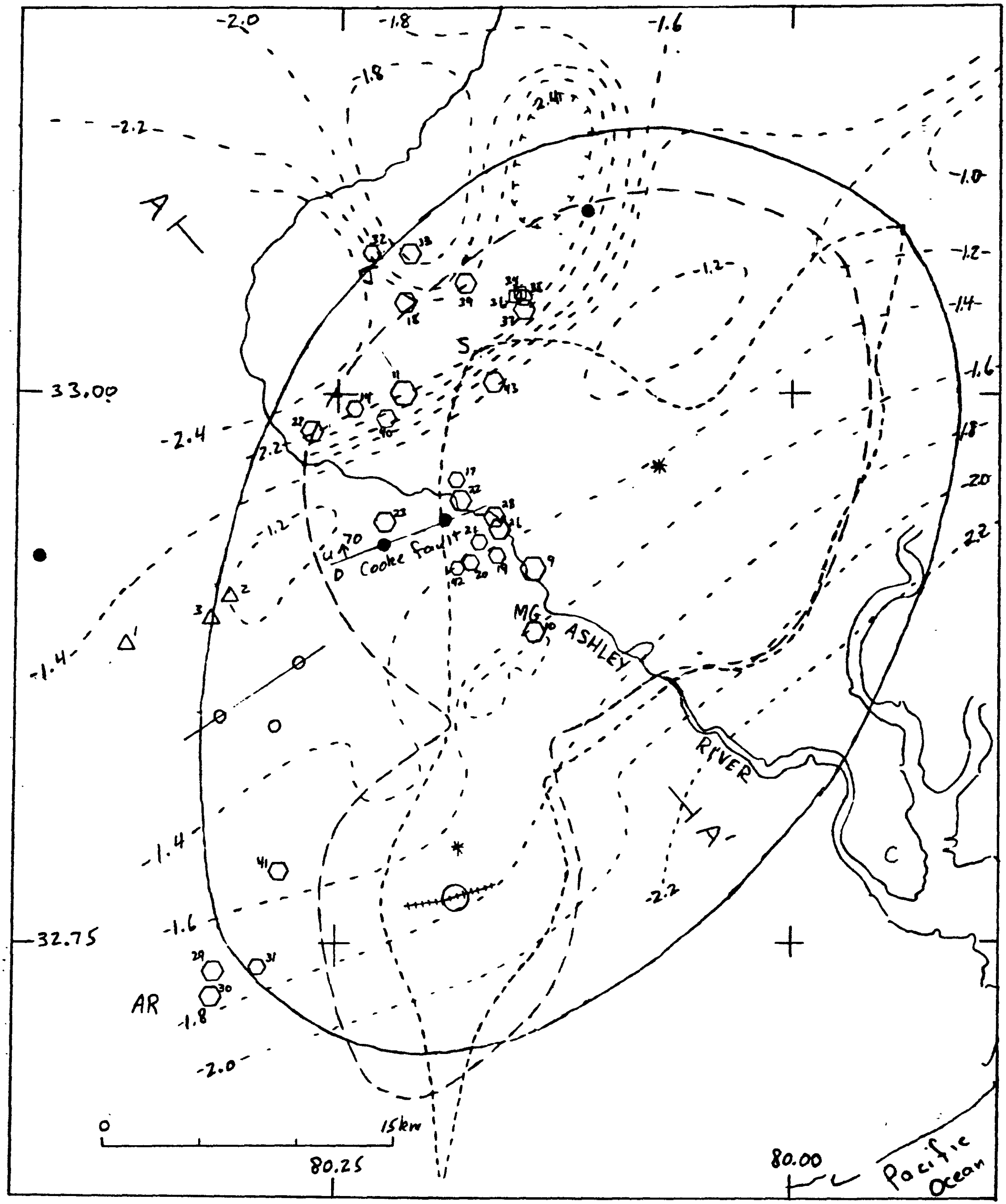

-2.0
-1.8
-1.6
-1.8
2.4
-2.2
-1.0
A
-1.2
-1.4
-1.6
-1.8
-2.0
-2.2
-1.2
S
33.00
-2.4
-2.2
-1.2
Cooke fault
U
D
70
MG
ASHLEY
RIVER
-1.4
A
-1.4
C
-2.2
-1.6
32.75
AR
-1.8
-2.0
0
15km
80.25
80.00
Pacific Ocean

Figure 4

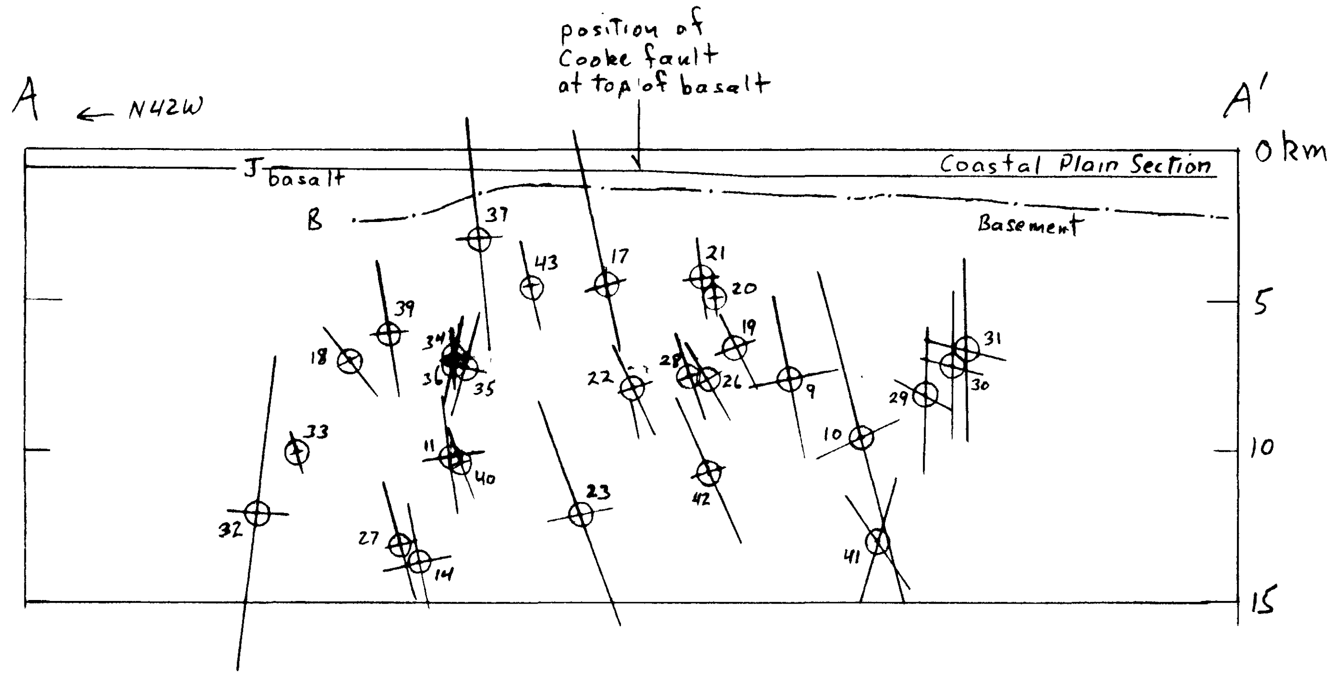
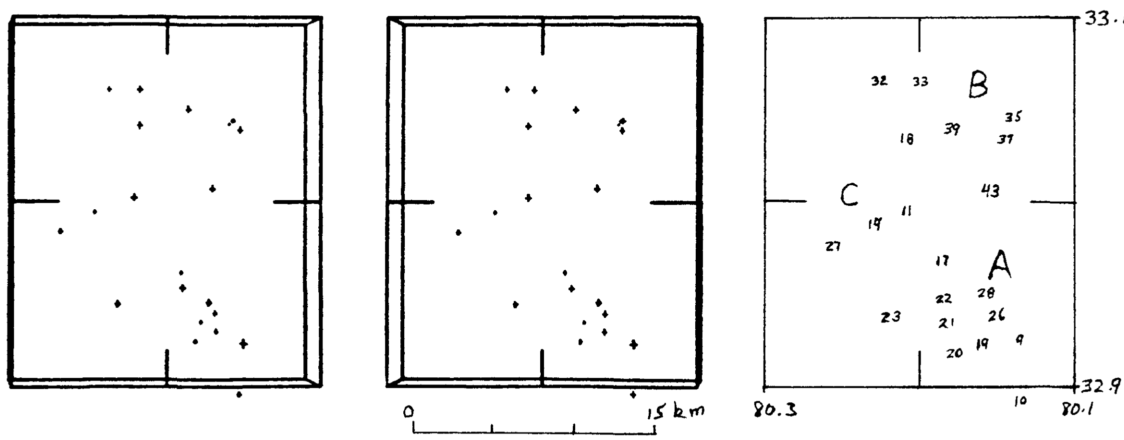

position of
Cooke fault
at top of basalt
A
← N42W
A'
0 km
5
10
15
J
basalt
Coastal Plain Section
B
Basement
37
43
17
21
20
19
39
18
34
36
35
22
28
26
9
31
30
29
10
33
11
40
23
42
32
27
14
41

Figure 5



0
15 km
33.1
32.9
80.3
80.1
B
C
A
32
33
35
39
18
37
43
11
19
27
17
22
28
23
21
26
20
19
9
10

Figure 6