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STATE-OF-THE-ART FOR ASSESSING EARTHQUAKE HAZARDS IN THE UNITED STATES

Report 21

SEISMIC SOURCE ZONES OF THE EASTERN UNITED STATES AND SEISMIC ZONING OF THE ATLANTIC SEABOARD AND APPALACHIAN REGIONS

by

Patrick J. Barosh

Weston Observatory
Boston College
Weston, Massachusetts 02193

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Seismic and geologic data indicate that the eastern United States has particular seismic source zones which can be delineated in terms of maximum expected epicentral intensity. Historic and recorded seismicities demonstrate that significant earthquakes arise only from specific source zones. These zones are associated with major fracture zones that trend northwest, an important exception being a northeast-trending zone extending from the central Mississippi Valley to the Gulf of St. Lawrence. Many fracture zones are related to those cutting the Gulf and North Atlantic oceanic basins, and their present development is related to the formation of these basins. Sags or embayments developed over their fracture zones during the Late Cretaceous where they intersected the Cretaceous coast. These embayments appear to have been relatively subsiding, and the major earthquakes occur in their inland portion. Rising uplands adjacent to the Atlantic coast also produce earthquakes as well as some inland extensional fault zones that extend northerly from the northwest-trending fracture zones. Earthquakes do not appear related to movements along large (Continued)

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faults as on the West Coast, but rather to adjustments on short segments of
faults at structural intersections.

Earthquake zonation is accomplished by estimating the maximum credible
earthquake for each source zone. Most zones probably have not experienced this
during the historic record, but some may have, such as Charleston, S.C. (Mod-
ified Mercalli, MM, Intensity X), Giles Co., Va. (MM Intensity VIII) and New
Madrid, Mo. (MM Intensity XII). Zonation in terms of Modified Mercalli Inten-
sity is less subjective than values of acceleration. Acceleration should be
determined from the intensity for specific sites.

PREFACE

This report was prepared by Dr. Patrick J. Barosh, Weston Observatory, Boston College, Weston, Massachusetts, under Purchase Order CW-82-M-4181. It is part of ongoing work at the US Army Engineer Waterways Experiment Station (WES) in the Civil Works Investigation Study, "Methodologies for Selecting Design Earthquakes," sponsored by the Office, Chief of Engineers (OCE), US Army. Technical Monitor for OCE was Mr. Paul R. Fisher.

Preparation of this report was under the direct supervision of Dr. E. L. Krinitzsky, Engineering Geology and Rock Mechanics Division (EGRMD), Geotechnical Laboratory (GL), WES, and the general supervision of Dr. D. C. Banks, Chief, EGRMD, and Dr. W. F. Marcuson III, Chief, GL. The report was edited by Ms. Odell F. Allen, Information Products Division.

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STATE-OF-THE-ART FOR ASSESSING EARTHQUAKE HAZARDS

IN THE UNITED STATES

Report 21

SEISMIC SOURCE ZONES OF THE EASTERN UNITED STATES AND

SEISMIC ZONING OF THE ATLANTIC SEABOARD AND

APPALACHIAN REGIONS

PART I: INTRODUCTION

A great amount of research has been carried out in recent years on
the causes of seismicity in the eastern United States and criteria with
which to delineate seismotectonic areas. Most of this research has been
sponsored by the US Nuclear Regulatory Commission, and has mainly been
directed along two divisions: study of present day seismicity and
study of geologic controls of the seismicity. Thus, the number of
seismic stations monitoring earthquakes in the region has been greatly
increased and seismotectonic studies have been made of the New England, New
Madrid, Nemaha, Mid-Continent Gravity Anomaly, and Charleston regions--
the first four by cooperative groups, the last by the US Geological
Survey (Fig. 1). In addition, utility companies have sponsored a number
of local studies and the Earth Physics Branch of the Department of

Energy, Mines and Resources of Canada has conducted many investigations of seismicity in southern Canada. The author has drawn heavily upon this work during preparation of this report.

A consistent relationship of seismicity to geologic structure and tectonic movement now seems to be emerging from a synthesis of these data. Relations that were found from the study of the seismic areas in New England over the past six years appear to apply to the other active regions in the eastern United States as well. The great variety of tectonic features and local structures apparently related to earthquake generation seems bewildering, but nonetheless may fit into an overall pattern and heirachy of features. It is the purpose of this report to present the preliminary results of assembling tectonic features to identify and define boundaries of earthquake source zones in the eastern United States. In addition, these data along with the available information on earthquake intensity are combined with estimates on the earthquake potential of source zones to produce an earthquake zonation map of the eastern seaboard and Appalachian regions of the eastern United States in terms of maximum expected epicentral intensity in firm ground.

PART II: DISTRIBUTION OF EARTHQUAKES
IN THE EASTERN UNITED STATES AND ADJACENT CANADA

Earthquakes in the eastern United States and adjacent Canada occur in scattered centers of activity that, nevertheless, form two very broad northeast-trending alignments (Figs. 2 and 3). A review of the altitudes of these active areas reveals that most earthquakes, and

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almost all destructive ones, occur in lowland areas below an altitude of 300 m (Barosh, 1979, 1980a). Exceptions to this are in the Great Smoky Mountains-Blue Ridge area, the Adirondack Mountains, and a few in the White Mountains.

Two general alignments that are readily apparent are one that trends southwest from the lower St. Lawrence River to northeastern Arkansas and a parallel one that trends through New England to northern Alabama (Fig. 3). These trends were first noted by Hobbs in 1904. The Arkansas-St. Lawrence alignment follows a remarkably straight lowland belt, below 200 m, from the lower St. Lawrence through the St. Lawrence lowlands and Lake Ontario to the southwest end of Lake Erie (Figs. 3 and 4). The lowland continues southwest of Lake Erie at less than 300-m altitude and would be lower if not for a thick mantle of glacial deposits that has buried the older valleys. The filled glacial Lake Maumee extends from Lake Erie across northwestern Ohio and over several preglacial river valleys that cross central Indiana and western Ohio (Fig. 5). The Anna, Ohio, earthquakes are located above this buried lowland near the junction of the buried Old Kentucky and Teays Rivers (the recent Sharpsburg earthquake occurred in northern Kentucky on the upper reaches of the Old Kentucky). The lowland continues along the Wabash tributaries and the lower Ohio River into the New Madrid region on the Mississippi River (Fig. 3). The principal seismic activity in the New Madrid region is concentrated in the river valley below an altitude of 200 m. Seismically active zones normal to this belt extend northwestward along the northeast side of the Ottawa River valley in Canada and along the Mississippi River where it forms the border between Illinois and Missouri.

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The pattern in the more eastern Atlantic belt of seismic activity is somewhat more complex and formed of both lowland and upland seismicity. The great majority of earthquakes in New England occurs below an altitude of 300 m, mostly below 200 m (Fig. 4). They are concentrated along the coast at bays and the mouths of major rivers and extend inland along north- and northwest-trending major river valleys. The activity in southeastern New York, along the Hudson River Valley, extends northward and merges with that of the St. Lawrence-Arkansas alignment (Fig. 3). The area of coastal lowland earthquakes extends from New England southwestward to Virginia as a broad, less distinct belt of activity centered over the Fall Line. The southeastern half of this belt stretches over very low land along the Fall Line, and across the northern Coastal Plain. A high concentration of activity also occurs farther to the southeast in the lowland around the Charleston, North Carolina, area, site of one of the largest earthquakes along the eastern seaboard. A northeast-trending upland belt of diffuse activity extends along the Great Smoky Mountains and Blue Ridge into the edge of New York (Fig. 3). Upland seismicity also occurs between the coastal activity and the St. Lawerence-Arkansas belt in the Adirondack Mountains, the southern edge of the White Mountains, and central New Brunswick (Fig. 3).

The distribution of earthquakes along the interior lowland St. Lawrence-Arkansas zone, the lowland along the Atlantic Coast, and adjacent interrupted upland zone apparently marks major neotectonic zones of the eastern U.S. The distinction between lowland and upland seismicity appears tectonically significant.

Records of modern-day instrumentation, unbiased by subjective factors, indicate that the concentration of earthquake activity in lowland areas is truly a tectonic feature. In the historical past (pre-instrumentation) more earthquakes were reported from lowland areas because these were and still are the centers of population and because these are the areas of thickest deposits of unconsolidated sediments. Such thick deposits may amplify the intensity of earthquakes relative to adjacent areas that are shallowly underlain by bedrock. This effect could have resulted in reports of more small felt earthquakes from valleys and '.. assigning epicenters of larger earthquakes to valleys where the intensity was the greatest. But today's seismograph network, which records magnitudes rather than intensity and permits accurate location of epicenters, also indicates that earthquake activity is most common in valleys and bays. These lowlands are shown to be controlled by geologic structure where the geology is well known and the earthquake distribution is more likely controlled by these structures. In fact, recent leveling studies have shown that some seismically active lowland areas are presently subsiding. There is also evidence that indicates areas of upland activity are rising.

PART III: SEISMICITY AND VERTICAL MOVEMENTS

Vertical crustal movements, chiefly subsidence, appear to be occurring in several places in the northeastern United States and adjacent Canada (Barosh, 1980b, 1982a) (Fig.6), and possible movement is suggested in other areas. The areas where this movement is taking place are seismically active.

SUBSIDENCE

Geodetic work shows that the La Malbaie seismic area in Quebec (Fig. 6) is situated along the eastern side of an area that is subsiding (Frost and Lilly, 1966; Vanicek and Hamilton, 1972). Resurveys of first order level lines show recent subsidence in Passamaquoddy Bay, Maine-New Brunswick, along the southern Maine coast (Tyler, Ladd, and Borns, 1979; Tyler and Ladd, 1981) (Fig. 7) and at least relative subsidence across the southern end of the Connecticut Valley (Brown, 1978). The data suggest that Passamaquoddy Bay is subsiding at the very rapid rate of 9 mm per year relative to Bangor, Maine (Tyler, Ladd, and Borns, 1979). Subsidence in the bay has been corroborated by tidal-gage records, positions of historical construction relative to sea level (Smith and Bridges, 1981), salt marsh advance (Anderson and others, 1981), and archaeological studies (Sanger, 1981), and tidal gage data (Walcott, 1972). Tidal gage data also indicate that Sandy Hook, New Jersey, on the southeastern side of Raritan Bay, is subsiding at an unusual rate

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(Walcott, 1972) as do changes of ledges in respect to sea level at the south side of Cape Ann, Massachusetts (D.C. Smith, written comm., 1982)

The mid-Atlantic coast north of Cape Fear, North Carolina, has the general features of a subsiding coastline (Fenneman, 1938), and the very extensive flooding of the Atlantic Coastal Plain in Virginia by the waters of the lower Chesapeake Bay appear to confirm an area of present-day subsidence indicated by releveling studies (Holdahl and Morrison, 1974). The southeast Georgia coastal area, which flanks the southwestern side of the Charleston seismic area, is subsiding relative to the adjacent areas as indicated by the deformation of Pleistocene and Holocene shoreline features (Winkler and Howard, 1977) (Fig. 8). Areas also appear to have subsided relatively in the area of the Valantine, Texas, earthquake (Ni and others, 1981).

The New Madrid seismic area is located at the head of the Mississippi embayment, one of the major structural features of North America. This trough-like sag was subsiding during the Cretaceous and early Tertiary as shown by the sedimentary record (King and Beikman, 1974). The area appears to have subsided as a graben in the Quaternary (O'Leary and Hildenbrand, 1981) and may still be subsiding, as indicated by the fact that the river drainage system of the midcontinent flows into and down the axis of the embayment. Also, stretches of the Mississippi River in the upper embayment are indicated to have anomalous grades. These stretches, which lie adjacent to the positions where faults project across the river, are out of grade, the slope of the bottom being opposite the southward flow of the river (R.B. Winkley, written commun., and S.A. Schumm, oral commun., 1981). Recent subsidence could explain these anomalies. In addition, a large area on

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the east side of the Mississippi River near New Madrid was down-dropped during the 1811-1812 earthquakes, and this produced Reelfoot Lake (Fuller, 1912).

UPLIFT

Uplift is known to be occurring in two regions in the eastern United States, the Adirondack Mountains and the Great Smoky Mountains-Blue Ridge region (Figs. 3 and 6). First-order level-line studies indicate that the Adirondack Mountains are rising at present (Isachsen, 1975; Barnett and Isachsen, 1980). Geomorphic and geologic evidence, also, indicates that the domal structure of these mountains is a relatively recent phenomenon. The position of the maximum measured uplift corresponds to the axis of the dome, the drainage system is in a youthful stage of evolution, and the sedimentary rocks surrounding the dome show no evidence of a nearby highland source during the Paleozoic (Isachsen, 1975). The axial part of the central and southern Appalachian Mountains from eastern Pennsylvania southwestward, centered on the Great Smoky Mountains-Blue Ridge region, is believed to be rising according to studies of deformed Tertiary erosional surfaces, mineral alteration (fission tracks), and level lines (Campbell, 1933; Fenneman, 1938; Meade, 1971; Zimmerman, 1980). Both these regions experience moderate levels of generally mild seismic activity (Figs. 3 and 6). The Adirondack Mountain area is the only upland area in the northeastern United States experiencing any significant activity at the present time, although some earthquakes have also occurred in the southern edge of the White Mountains of New Hampshire. Seismic activity also occurs across the international border in an upland area in central New Brunswick (Fig. 3). The present seismograph network may miss very small earthquakes in other upland areas, but no significant activity should be missed.

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PART IV: RELATION OF SEISMICITY WITH GEOLOGIC STRUCTURE

To understand the causes that are responsible for seismic activity and vertical movements, it is necessary to look for geologic features that might both cause the movements and match the distribution of activity.

Several different causes have been proposed for earthquakes in New England and elsewhere on the East Coast. Earthquakes have been ascribed to glacial rebound (Leet and Linehan, 1942), granitic plutons (Collins, 1927), basic plutons (Kane, 1977; Simmons, 1977), reactivation of major Precambrian and early Paleozoic structures (Bollinger, 1983), movement on ancient deep thrust faults (Seeber, 1983), reactivation of Mesozoic rifts (Ratcliff, 1983), post-Cretaceous reverse faulting (Wentworth and Mergner-Keefer, 1980), intersections of major structural zones (Hobbs, 1907) and a special Boston-Ottowa seismic zone (Hobbs, 1907). All of these fail the test of adequately matching the distribution of seismicity, although the latter two do match to some degree.

The earthquakes can safely be assumed to be tectonic in origin and related to fault movement, which may cause vertical change in the earth's surface. However, no fault breakage at the surface during an earthquake has been definitely proven in the eastern United States, although it may have occurred during the New Madrid earthquakes of 1811-1812 (Fuller, 1912). Even though more and more post-Cretaceous faults are being recognized in the east, positive identification of active fault zones has been difficult to achieve due to a number of reasons. Also, many of the younger faults recognized may be due to gravitational rather than direct tectonic movement. Relatively few faults have been mapped until recently in the Appalachian Mountains, but where they have, they are

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found to be very abundant and, for the most part, very old. In the north, glacial activity has smoothed possible old fault scarps and produced numerous erosional and depositional scarps. To further complicate things, the glacial deposits are commonly broken by faults due to slumping. In the south a thick weathered zone, saprolite, obscures the faults. In both areas generally insufficient use has been made of geomorphic, LANDSAT, aerial photographic and geophysical data to look for young faults. In the mid continent many faults have been identified by stratigraphic offsets, even where covered, but a general lack of post-Paleozoic rock makes it difficult to determine recent movement. However, Holocene faults are now being recognized and in Texas and the Gulf Coast many Quaternary faults are known. In New England and adjacent Canada small offsets of glacially smothered surfaces, that occur at many locations, have been attributed to Holocene movement (recently summarized by Adams (1981)). However, rapid freezing, that produces fractures and local shaking with up to intensity V affects in the region, could have caused most of these offsets (Barosh and Smith, 1980).

Thus few young faults are known with which to determine movement. But even if many were known, another approach to determining movement would probably be needed as the fault movements are only local manifestations of broader surface movements. Some type of datum planes, which may have recorded these movements, are needed. Fortunately, a series of them exists along the Gulf and Atlantic Coastal Plains: the lower boundary of the Upper Cretaceous rock, the Upper Cretaceous-Lower Tertiary contact, the Upper Tertiary-Quaternary contact, and the present shoreline (Fig. 9). In a simplified way these contacts form a series of

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still pictures of the changing oceanic shoreline of the eastern United States over the past 100 million years and recorded the surface movements that have occurred. There is no geologic evidence to suggest any recent significant change in this regime, and the earthquakes are related to the continuation of these movements.

These Cretaceous to Quaternary deposits are part of the results of a profound change in tectonic activity that affected the eastern United States in the early Mesozoic as the Atlantic and Gulf of Mexico basins began to develop. Earlier tectonic features influence this development, and locally there was significant reactivation on properly oriented structures, but in general the active tectonic features of the late Precambrian and Paleozoic ceased to play a dominant role. Grabens formed in the Late Triassic and Early Jurassic parallel to the future Atlantic coast as widening over the subsequent position of the North Atlantic basin caused a stretching and thinning of the crust (Fig. 10). By mid-Jurassic, restricted marine basins were forming near the coast and filling with salt deposits and the clastic sediments in the Late Jurassic as the precursor of the great clastic wedge that built out over the downwarping new edge of the continent during the Cretaceous and Tertiary (Balley, 1981; Grow, 1981; Dillon and others, 1983) (Fig. 11), and these movements have continued in a more modest way to the present (Watts, 1981).

The present configuration of the Cretaceous and Tertiary deposits is due to a combination of depositional, erosional, and tectonic processes whose separate effects can generally be divided and studied in order to understand the movements that have taken place. The erosional changes are not so serious as to prevent using the contacts as general

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planes of reference. The depositional history is sufficiently well known to add considerable detail to how the depositional basins, reflecting movement, have shifted with time. The sediments were not deposited uniformly in the wedge, but formed greater and lesser thicknesses reflecting a series of transverse basins and arches spaced along the Coastal Plain. The transverse sags producing the basins caused the inner contact of the deposits to extend further inland, embaying the continental interior and thus are referred to as embayments, and they extend shoreward across the arches to produce an irregular sinusoidal contact (Fig. 12). If movement occurs over the embayments and arches while sediments are being deposited, the contacts will be closer together in the embayments and further apart on the arches (Fig. 12), instead of being roughly parallel, and the units thicken in the embayments. This relation is common along the Coastal Plain. Thus the sedimentary and structural history of these deposits provide a record of movements beginning in the Late Cretaceous.

The earlier Upper Jurassic and Lower Cretaceous deposits appear too irregular to provide good reference planes. Their distribution seems to be strongly affected by the earlier grabens in many places, and they serve as a transition to the more broadly warped Late Cretaceous.

Areas of suspected or known Late Tertiary to recent vertical movements appear to coincide with the position and direction of the Late Cretaceous-Early Tertiary movements. Nearly all of the areas of major seismicity in the eastern United States lie in or adjacent to the heads of Late Cretaceous-Early Tertiary embayments. Where deposits of this age are lacking in seismically active coastal parts of New England and adjacent Canada other structural characteristics that are associated

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with the embayments are found, and the seismicity occurs near bays and lower river valleys (Fig. 9). The arches generally appear to be passive structures, that lagged behind the embayments in subsidence and have little seismicity associated with them. The exception is the Alabama arch which forms the southwest end of the uplift forming the Appalachian highlands. The highlands are apparently still rising and producing earthquakes.

The embayments themselves probably have no direct bearing on the cause of seismicity; they merely indicate places around the edge of the continent where weak zones, with a history of subsidence, are located. It is these weak zones that underlie and extend from the heads of the embayments that appear to be the major source zones of earthquakes.

These source zones may have formed at different times, have complex structures and histories, but have been reactivated or formed under the strain of the formation of the North Atlantic and Gulf of Mexico basins. Those around the Gulf appear to be reactivations of late Precambrian rift systems. Those along the Atlantic are due to Mesozoic rifts and oceanic fracture zones. A single simple zone of movement may reflect the interplay of a great variety of overlapping spatial and temporal features. It is the integration and relations within this mosaic of structure that are important in understanding the cause of seismicity. Looking for relations of seismicity with single features such as reverse faults or grabens is too simplistic and misses important overall relations.

The seismicity within the continent away from the embayments on the Coastal Plain also appears related to the continuation of these same Late Cretaceous-Early Tertiary movements. Upland earthquakes in this

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area occur in areas that rose to provide the sediments for the Cretaceous and Tertiary deposits. Other interior seismicity appears associated with fracture zones, that are mainly of northwest and north trends, probably extensional and related to the extensional movements of the Atlantic opening.

All of these movements are related to the larger concept of tectonic movements of crustal plates which in turn may be due to variation in the earth's rotation.

The broad regional tilts that have affected North America since the Cretaceous, such as the Pleistocene warping due to glacial loading and unloading, appear to have no discernible effect in the present day seismicity.

PART V: SEISMIC SOURCE ZONES

The identification and delineation of seismic source zones provide
the necessary basis on which seismic zoning can be done with confidence.
Without this step, zonation involves too many assumptions to be
satisfactory. Without knowing why a large earthquake occurs in a
particular region it can be argued that it may occur any place (Devine,
1982) and obviate the need for zonation, merely assigning a value from
the largest earthquake known for an entire region. This is the problem
in making quantitative treatment of earthquake data alone - dealing with
epicentral densities or energy release - although it may actually have
practical application - as it does reflect source zones, it does not,
however, provide boundaries between them. Attempts to circumvent the
problem of understanding source zones by probabilistic studies involve
too many stated and unstated assumptions about source zones and
therefore lack a sufficient scientific basis for reliability (Knopoff
and Kagan, 1977; Evernden, 1975, 1982). The presentation of results of
these studies in percent probabilities and tenths of percent gravity
gives them an aura of undeserved precision.

Frequency-magnitude studies to estimate maximum potential
earthquakes also have limited value, especially if the data are
inadvertently taken from two or more source zones.

Seismic source zones must be identified to provide reliable
zoning. It is a geologic problem, aided by geophysical studies, to
determine the controls on earthquake locations. Focal plane solutions
are of very limited help at present as they only address a small part
of the problem, vary greatly between determinees, are inconsistent in

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orientation, are often inconsistent with the known movement on active faults for small earthquakes in the Western United States and are untested in the eastern part.

Earthquakes are the result of earth movements, and source zones reflect present day earth movements and need to be studied by investigations of these and the most recent movements indicated by geologic evidence. Geologic investigations in the Eastern United States have understandably concentrated on the tectonic events affecting the widespread Paleozoic rock, with relatively little effort put on the present day events. These older events and resultant structures only affect the modern day seismicity to the degree they have been reactivated. This lack of good correlation of present seismicity with ancient structure has led to seeking some novel causes of seismicity.

The vast amount of new geophysical data and local areas of new detailed geologic mapping onshore combined with the explosion of stratigraphic and geophysical data along and offshore provide a greatly improved basis for revealing structure and examining movements in the recent geologic past.

The following sections will review the geologic and tectonic history of the areas of concentrated seismic activity to seek to reveal these features most closely related to the spatial distribution of the earthquakes and apparently their cause.

The earthquake source zones could be described in various ways, direction of trends, age, or kind of geologic structure, for example. In the following sections, however, they are grouped by the relation to the Late Cretaceous shoreline and movements in terms of coastal, upland, and interior. These are somewhat loose groupings without clear-cut divisions

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in places, but are felt to best reflect source zones with similar characteristics (Fig. 3). In the following discussion of source zones the largest earthquake known in each zone is given in parentheses, along with its intensity (I) and magnitude (Mag.) if known, beneath the zone heading.

COASTAL ZONES

The coastal zone is in reference to the Late Cretaceous-Early Tertiary coastal region which in places is far interior to the present Gulf of Mexico and Atlantic shorelines. The areas of seismic activity associated with the coastal zone are designated ʰy the embayment they are associated with, and north of the Coastal Plain, generally by the associated bay (Fig. 13). The coastal zone, as marked by the inner boundary of deposits of Late Cretaceous-Early Tertiary age, extends offshore near New York, but continues along the coast to north of Boston. Farther north off the Maine coast the inner edge has apparently been removed by glacial scour. The similarities of geologic environment and subsidence of the seismic zones associated with bays and lower river valleys along the submerged inner boundary and Maine coast to the north with the embayments have led to grouping them together. The causative structures underlying the embayments may extend well interior of the head of the embayments but are nevertheless included with them as a single genetic grouping.

The embayments and bays are described below in order of their location from the Mexican border northeastward to Canada. The seismicity associated with them changes from a typical "western" to

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"eastern" in the southwest and the embayments demonstrate that the crustal mcvement related to both types is fundamentally the same.

Rio Grande Embayment

(Valentine TX, 1931.   I = VIII, M = 6.4)

Summary

The Rio Grande embayment is a southeast plunging Late Cretaceous - Early Tertiary basin that lies within the Gulf Coastal Plain along the Texas-Mexican border. A northwest-trending structural zone, the Texas Lineament, which may have begun as a Precambrian rift and has a long and complex pre-Cretaceous history, extends into and beneath the embayment. Subsequent movement along this lineament produced folds and faults in the Early Tertiary. Rifting began in mid-Tertiary time and is associated with the development of the southern end of the Rio Grande Rift; abundant volcanism and extensional faulting continued during the late Tertiary. The structural zone along the lineament is seismically active and the Valentine earthquake of 1931 (I = VIII) occurred on one of the northwest-trending faults (Dumas, 1980).

Geology

A southeast-plunging Late Cretaceous-Early Tertiary basin lies within the Gulf Coastal Plain along the Texas-Mexican border. It is referred to as the Rio Grande Embayment after the river that flows through it (Figs. 7 and 5). The Rio Grande embayment was a subsiding basin at the end of the Mesozoic and contains the maximum thickness of

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Upper Cretaceous rock found in the Gulf Coast (Stephenson, 1928; Murray, 1957). The subsidence is estimated at between 0.2 to 0.4 inches (0.5 to 1.0 cm) per year in mid-Cretaceous (Renfro, 1973). The northern flank of the embayment was the focus of fault and volcanic activity during the Late Cretaceous (Luttrell, 1977). Upper Cretaceous strata thin from the embayment to the northeast over the San Marcos arch, which separates the Rio Grande from the East Texas Embayment (Sellards and others, 1933)(Fig. 14) The San Marcos arch was also subsiding, but at a much lower rate (Loucks, 1977).

A northwest-trending zone of deformation, referred to as the Texas Lineament, lies just eastward of the Texas-Mexican border in the Big Bend area (Albritton and Smith, 1957; Muehlberger, 1980)(Fig. 14). Structural features at the southeast end of the lineament, in the vicinity of the Devils River uplift, are believed to continue farther southeastward and to have influenced sedimentation and controlled the shape of the Rio Grande Embayment since the Late Jurassic (Luttrell, 1977). The Texas Lineament, where best seen in West Texas, is a fundamental crustal discontinuity, as shown by recurrent deformation, metamorphism and intrusion, which separates more stable crust on the northeast from less stable crust on the southwest (Muehlberger, 1980). The lineament has been extended northwestward into central Arizona along the Mogollon Rim (Elston and Bornhorst, 1979, Fig. 6) following various structural and volcanic features. Gravity and magnetic maps (Keller and others, 1982) show northwest-trending discontinuities along the zone and northwest-trending LANDSAT lineaments appear to bound it in West Texas (Muehlberger, 1980).

Deformational periods affecting the Texas Lineament have recently

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been summarized by Muehlberger (1980) and Dickerson (1980). Volcanism and thrust faulting occurred during the late Precambrian and deformation took place during every period of the Paleozoic; the deformation in the Permian was accompanied by right-lateral strike-slip faulting.

Several mid-Mesozoic events, related to the opening of the Gulf of Mexico, can be identified in the Texas Lineament zone (Muehlberger, 1980). Repeated subsidence occurred in the area as basins developed and changed from terrestrial to restricted marine to more normal coastal marine environments as they filled, merged, and subsided to form the Rio Grande Embayment. Several northwest-trending raised blocks and basins were formed, probably from block faulting of Triassic to Middle Jurassic time and clastic red beds were deposited (Humphrey, 1956). The Sabinas Basin was a grabenlike trough or intermountain basin that probably began filling in the Late Triassic or Early Jurassic with clastic red beds; it then changed during the Middle or Late Jurassic to a restricted embayment that received thick evaporite deposits (Conklin and Moore, 1977) as did the Chihuahua Trough to the southwest of it (Fig. 14). Very thick Lower Cretaceous strata were then deposited in both these basins as well as some in the Maverick Basin (Scott and Kidson, 1977; Loucks, 1977). Most of the Mesozoic history of the southeastern end of the Lineament zone involves the gradual filling of the Sabinas Basin and the gradual submergence of the Tamaulipas and Coahuila Peninsulas (Conklin and Moore, 1977) resulting in the broader basin of the Cretaceous Rio Grande Embayment.

At the end of the Late Cretaceous, Rio Grande Embayment, as well as the area to the northwest, was affected by Laramide faulting and folding (Muehlberger, 1980; Wall and others, 1961; Conklin and Moore, 1977) and

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left-lateral movement occurred along the Texas Lineament (Muehlberger, 1980). The Paleozoic to Cretaceous structure appears to have controlled both the style and trend of Laramide folding (Wilson and Pialli, 1977).

During the early Tertiary the dominant deformation was extensional accompanied by widespread volcanism, intrusive activity, and probably evaporite diapirism. In Miocene time the Rio Grande Rift at the northwest end of the lineament began forming in response to extension accompanied by right-lateral strike-slip movement (Muehlberger, 1980; Dickerson, 1980) and developed into a major north-trending structure through central New Mexico into central Colorado (Fig. 15). Just south of El Paso, Texas, the rift turns southeastward into the Texas Lineament (Figs. 15 and 16). Chapin (1979) states that "Rifting began between 32 and 27 M.Y. ago when regional extension reactivated the southern Rocky Mountains, a major north-trending zone of weakness that had developed during late Paleozoic and Late Cretaceous-Early Tertiary orogenies. By 26 M.Y. ago, the crust along the developing rift had sagged sufficiently to form broad shallow basins in which mafic flows and volcanic ash beds were interculated with alluvial fill." Further opening produced basin-and-range block faulting accompanied by extensive volcanism.

Late Cenozoic faults are abundant in West Texas including ones that cut Quaternary deposits in the Salt, Hueco and Presidio and Marfa grabens (Muehlberger and others 1978; Muehlberger, 1980; Goetz, 1980; P.W. Dickerson, written commun, 1983) (Figs. 16 and 17). The contemporary Marfa graben is superimposed upon a Late Pennsylvanian - Early Permian graben that received thousands of meters of sediment (P.W. Dickerson, written, commun. 1983). The remnants of Late Cretaceous strata associated with the head of the embayment at that time extend

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northwestward along the lineament to 31° latitude and are extensively cut by the Cenozoic faults (Renfro, 1973, section F-F').

Seismicity

The Rio Grande Rift and the Texas Lineament appear to be important seismogenic zones. Most of the seismicity and all of the larger earthquakes of West Texas and New Mexico (Coffman and others, 1982) are associated with the Rio Grande Rift and a large amount of the activity in southwestern New Mexico, and Arizona (Sanford and others, 1979; Dubois and others 1982) occurs along the proposed extension of the Texas Lineament (Fig. 2). There are many known late Cenozoic faults and the earthquakes are the result of further movement, except for one area near Socorro in central New Mexico, where a large concentration of earthquakes have been recorded and considered due to a body of magma rising within the rift (Sanford and others, 1979).

The Valentine earthquake of 1931 is interpreted to have been the result of right-lateral movement along the northwest-trending Valentine fault (Dumas, 1980) (Figs. 17 and 18). Geodetic measurements have also indicated relative subsidence near Valentine that may be related to the earthquake (Ni and others, 1981). Several moderate earthquakes have occurred to the northwest within the rift near Socorro, New Mexico where one in 1906 reached an intensity of VII-VIII; four others of intensity VI and eight of V have occurred. The Santa Fe, New Mexico area experienced a VII to possibly VIII intensity earthquake in 1918 (Coffman and others 1982) (Fig. 2).

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The Texas Lineament is much more active than the historic record indicates as shown by a recently operating seismic array that recorded a few hundred earthquakes from 1976 to 1980 (Dumas, 1980) (Fig. 19). Almost all of the recorded earthquakes, occur within the zone of late Cenozoic faults, suggesting that the outer boundary of these observed faults can be used to delineate the source zone. Several of the earthquakes that occur outside this zone in Texas (Reagor and others, 1982) are apparently due to oil field activities and dam filling (Sansom and Shurbet, 1983).

The earthquake activity along the Texas - Mexican border has the attributes of both "western" and "eastern" seismicity and may help clarify the differences between the two.

The seismicity of the southern end of the Rio Grande Rift in the Texas Lineament and its relations with geologic features is typical of that of most of the western mountain regions of the U.S. The relation of the seismicity to an extensional structural zone near the head of a Late Cretaceous embayment, which it apparently controls, is also typical of most areas of eastern seismicity.

The earthquakes of the rift are associated with faults that are part of extensional movements forming physiographically well expressed basins with northwestward and northward trends being dominant. Many late Cenozoic faults and fault scarps are present, formed by normal and,

along northwest-trending segments, some right-lateral movement. Volcanoes and abundant volcanic rock are present (Fig. 15) and have hot springs and high heat flow associated with them (Swanberg, 1979). The volcanoes generally came up extensional faults at the border of the basins and probably overlie igneous bodies emplaced along the faults. The Precambrian surface is depressed beneath the basins in the rift, and the rift area is generally marked by a gravity high due to stretching and consequent thinning of the crust that has allowed the dense lower crustal and mantle material to rise (Fig. 20). The rift has been well studied by many excellent geologists and the general cause of seismicity is generally understood and agreed upon, although the potential hazard of any particular fault and the maximum possible earthquake that might be generated may not be.

The earthquakes in West Texas are also in a major structural zone that experienced a great amount of earlier deformation and underwent episodes of subsidence and possible extension during the Mesozoic that resulted in the broad embayment. The rifting, transcurrent movement along the Texas Lineament, and subsidence in the area of the embayment are all related (Fig. 21). Embayments associated with older rifts are present at the many other major active areas in the eastern U.S. The manifestations of subsidence produced by the rifting in West Texas and that now known or suspected in other areas of eastern seismicity may reflect a difference in rate rather than nature. Similarities in fault trends also occur between the Rio Grande Rift and other active areas in the east.

East Texas Embayment

(El Reno, OK, 1952, I = VII, M_b = 5.5)

Summary

The East Texas Embayment, like the Rio Grande Embayment, appears to have developed as a sag over a major northwest-trending rift that began in the late Precambrian and was a zone of repeated deformation usually characterized by subsidence. Folds in the Lower Cretaceous rock are controlled by earlier structures and appear to be a continuation of it. Post-Cretaceous faults appear to be present, but hard to document. Most of the earthquake activity in the region occurs along the flanks of the old rift and along an extensional fault system, the Humbolt fault zone, that trends northward from the rift; a pattern similar to that associated with the Rio Grande Embayment. The greatest concentration of earthquake activity occurs where the rift intersects the northerly-trending extensional fault system and earthquakes farther north along the system may be controlled by northwest-trending cross faults.

Geology

The East Texas Embayment is an Early Cretaceous-Early Tertiary basin that is situated in northeast Texas and overlaps into southern Oklahoma (Figs. 9 and 14). It is separated from the Rio Grande Embayment to the southwest by the San Marcos Arch. A salient of

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pre-Cretaceous rock, that may represent a gentle arch across southern Arkansas separates it from the northeast-trending Mississippi Embayment to the east. The axis of the East Texas Embayment trends northwestward near the Texas-Oklahoma border, but this trend is interrupted farther southeast by the Sabine uplift (Fig. 14). Deposition began in a more irregular fashion in the embayment during the Early Cretaceous (Fig. 22) and became more regular in the Late Cretaceous forming rock that now forms a broad uniform downwarp.

The head of the East Texas Embayment overlies and is aligned with a major structural trough system through southern Oklahoma, consisting of the Anadarko, Marietta and Ardmore Basins that forms a great sag in the basement of the mid-continent (Figs. 23 and 24). This structural zone, along with the Wichita Mountains and Amarillo Uplift to the southwest, first formed as a major rift zone that was active from Late Precambrian to Middle Cambrian and filled with volcanic rock (Ham and others, 1964; Feinstein, 1981) (Figs. 25 and 26). The rift zone, referred to as the Southern Oklahoma geosyncline or aulacogen, is thought to have formed as an inland extension of more general rifting that tore the continental crust to the east and southeast completely away from the North American craton (Hoffman and others, 1974; Burke and Dewey, 1973). The rift, continued to subside, in part or whole, during the rest of the Paleozoic (Ham and others, 1964; Feinstein, 1981) and the Paleozoic sediments thin to the northeast from it (Fig. 24).

The history of the structural development of the rift and its influence on sedimentation from the Late Mississippian to the end of the Permian (Rascoe and Adler, 1983) shows it was one of the most important

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structural zones in the mid-continent during this period. The rift began to break up and its southwestern edge started to rise in Late Mississippian or Early Pennsylvanian to form the Wichita Uplift in Oklahoma and its more subdued extension, the Amarillo Uplift in north Texas (Gilbert, 1982) (Figs. 23 and 26). The Anadarko Basin on the northeast subsided, especially during the Permian, to form the deepest basin in the North American craton (Ham and Wilson, 1967) and the Marietta and Ardmore Basins formed (Fig. 24). The Arbucle Mountain Uplift formed during the Pennsylvanian along the northeast side of the old rift (Ham and others, 1964). The border faults of the Wichita Uplift are reverse at the surface. Recent seismic work has shown these faults to dip gently towards the uplift (Brewer, 1982). Such inward dipping faults are a common occurrence at the borders of uplifted blocks in the Rocky Mountains and in experimental studies (Logan and others, 1978; Cook 1978) and also occur along strike-slip fault zones (Sylvester and Smith, 1976). Some evidence indicates left-lateral strike-slip movement did occur along the principal faults in the rift in the post-early Paleozoic and pre-Permian interval (Pruatt, 1975; Donovan, 1982). The deformation in the rift since the Permian is less apparent, although some faults do cut Pennsylvania and Permian rock in Oaklamoma (Hart, 1974; Havens, 1977) and Triassic rock along the Amarillo Uplift in Texas (J. Peck, oral commun., 1983).

The axis of the East Texas Embayment is aligned with the ancient rift zone and also roughly mimics the curve of the Pennsylvanian Ouachita structural belt (Figs. 14 and 23, 24, and 25). This suggests the embayment is controlled by both earlier structural zones or, more

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likely, that both the embayment and bend in the Quachita belt were influenced by the rift. Lower Cretaceous rock near the head of the embayment is deformed by several northwest-trending folds. One of these, the Marietta Syncline, is coincident with an axis of synclinal subsidence that began in the Late Mississippian and this shows the Cretaceous features are a result of continued structural adjustment (Frederickson and Redman, 1965). The folds parallel and are the same order of magnitude as the older fault bounded structural blocks in the adjacent Arbuckle Mountains, Criner Hills and Wichita Mountains (Hart, 1974; Havens, 1977). This suggests movement on similar blocks beneath the Creaceous may have been reflected as folds in the softer overlying sediments. This deformation of the Cretaceous may be Laramide in age as it is in the Rio Grande Embayment and be similar to other Laramide folds drapped over underlying offset blocks in the Rocky Mountains (Matthews, 1978)

Other post-Cretaceous deformation near the head of the embayment is difficult to demonstrate. The Cretaceous in the Marietta Basin appears faulted (Westheimer, 1965) and it may be elsewhere. Some of the many northwest-trending faults cutting the pre-Pennsylvanian rock the Arbuckle region (Hart, 1974) may have been reactivated. The stream trends across the Pennsylvanian, Permian and Cretaceous rock in the surrounding region is parallel to and, in several cases, extends from these older fault zones indicating some structural control of the drainage, perhaps by some reactivation of these zones to fracture the younger rock.

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Some post-Cretaceous deformation may have occurred northwest of the embayment along the old rift zone. Cretaceous rock extends east of the general distribution of Cretaceous in the southern Great Plains as scattered outcrops along the Anadarko Basin (Miser and others, 1954). This suggest the basin was either a low area that received a greater thickness of Cretaceous sediment or there has been some post-Cretaceous subsidence to preserve the rock. Relief along the Meers fault, that bounds part of the northeastern side of the Wichita Mountains, suggests Quaternary movement along this northwest-trending fault (Donovan and others, 1983).

Another major structural zone of the mid-continent, the Nemaha Uplift, extends northward from the edge of the ancient rift through central Oklahoma, and Kansas to the southeast corner of Nebraska (Figs. 23, 24 and 27). The uplift is flanked on the east by a zone of large normal faults, the Humbolt fault zone, and on the west by a zone of much smaller normal faults. The Nemaha Uplift was mainly active in Pennsyvanian time when the Wichita Uplift was also rising, however, it existed as a positive structural element since Early Cambrian time or possible even earlier (Berendsen and others, 1981). The Humbolt zone appears offset, in places, by northwest-trending faults (Wilson, 1979) (Figs. 27 and 28) and prominent northwest-trending aeromagnetic and topographic lineaments (Berendsen and others, 1981; Yarger and others, 1981). The uplift is formed of north-northeast trending horst blocks raised to different heights along northwest-trending tranverse faults; such structured intersections also appear to control small Cretaceous peridotite and kimberlite bodies (Berendsen and others, 1981) (Fig. 28).

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Seismicity

Zones of seismicity appear to follow both the Southern Oklahoma Geosyncline and the Nemaha Uplift (Figs. 2 and 29). A zone of moderate-sized earthquakes lies towards the southwest side of the old rift zone from Marietta Basin, at the head of the East Texas Embayment, to the northwest corner of Texas (Coffman and others, 1982, Reagor and others, 1982; Stover and others, 1981) (Fig. 2). Another follows its northeast flank from El Reno, Oklahoma, on the northeast edge of the Anadarko Basin, southeastward through the Arbucle Mountains (Figs. 24 and 29). The activity in the Marietta Basin and that along the northeast edge of the geosyncline shows up well by the small earthquakes recorded by the seismic network installed in 1976 (Fig. 29). The greatest number of both moderate earthquakes (seven with intensity V to VII) and small ones occurs in a cluster around El Reno, Oklahoma, (Lawson and others, 1979; Coffman and others, 1982; and Luza and Lawson, 1982, 1983) (Figs. 2, 29 and 30). The earthquakes at El Reno and those to the southeast lie approximately along northwest-trending aeromagnetic lineaments (Fig. 30). A few short northwest-trending faults that cut pre-Pennsylvanian rock (Luza and Lawson, 1982, 1983) lie also along the lineaments as well as LANDSAT lineaments (Shoup, 1980) and, for much of its length, the Canadian river. These lineaments may mark pre-Middle Pennsylvanian basement faults along the flank of the geosyncline.

The El Reno earthquakes also lie at the southwest end of a zone of small events that trend north-northeast following the general trend of the Humbolt fault zone along the east side of the Nemaha Uplift (Burchett and others, 1983), although at its south end it obliquely crosses a local bend in the uplift (Lawson and others, 1979; Luza and

Lawson, 1982, 1983). The earthquakes at El Reno may be localized by movement on northwest-trending basement faults along the northeast side of the Anadarko basin where it crosses the Nemaha structure. There is also some suggestion that the earthquakes farther north along the east side of the Nemaha occur where northwest-trending faults may intersect the structure (Berendson and others, 1981)(Figs. 27, 28 and 30).

The geologic relations of other small earthquakes that occur in eastern Oklahoma are less clear, but they are probably related to northeast and northwest-trending pre-Pennsylvanian basement faults (Fig. 30). One apparently long northwest-trending zone of seismicity extends from near the northwest corner of Garfield county through Okfuskee county and into the western edge of Arkansas (Figs. 27 and 30). This zone has a few faults mapped along it and also follows aeromagnetic lineaments that parallel the geosyncline and may represent a basement fault zone related to it.

Some earthquke activity also occurs within the embayment in east Texas, one in 1930 reaching intensity VI (Coffman and others, 1982). These lie near the south side of the Sabine Uplift (Figs. 2 and 14) that rose in the embayment approximately on line with the Wichita Uplift. Little is known about these earthquakes because of lack of information on basement structure and the possibility that the seismicity is associated with oil field operations.

The East Texas Embayment, rift zone and Nemaha extensional faults have the same geometrical arrangment as the Rio Grande, Texas lineament and Rio Grande Rift (Fig. 31). This suggest similar present—day movement in both systems, albeit weaker in the East Texas Embayment, although the structural pattern may be an inherited Early Pennsylvanian or much earlier one.

Mississippi Embayment

(New Madrid, MO, 1811-1812, I = XII)

Summary

The Late Cretaceous-Early Tertiary Mississippi Embayment overlies, and is apparently controlled by the late Precambrian Reelfoot Rift that also influenced Early Paleozoic and Pennsylvanian sedimentation. The area has a long history of relative subsidence that may still be continuing, although locally both uplift and subsidence are indicated within the embayment. Numerous pre-Cretaceous faults and some post-Cretaceous faults are present including Holocene ones. These generally parallel the underlying late Precambrian structure. The present–day seismic activity is concentrated on northeast-trending probable fault zones along the southeast side of the rift under the head of the embayment and on northwest-trending faults along the northeast side of the Pascola Arch, that crosses the rift. The greatest activity occurs where the structures cross near New Madrid. Other seismicity around the head of the embayment may be associated with normal faults, especially ones of northeast and northwest trends.

Geology

The Mississippi Embayment is a very large northeast trending synclinal basin of Upper Cretaceous and Lower Tertiary rock (Fig. 9). It is separated from the East Texas Embayment to the southwest by the Ouachita

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Mountains of southeast Oklahoma, that may be slightly arched and from the Southwest Georgia Embayment to the southeast by the very broad arch across the southwest end of the Appalachian Mountains, referred to here as the Alabama Arch (Fig. 9). The axis of the syncline lies toward the northwest side of the embayment, approximately along the northeast-trending reach of the Mississippi River. Where the river enters the embayment it travels southwestward and flows along the axis to the southwest end of the embayment where the river turns again and flows southward to the Gulf. The embayment developed during the Late Cretaceous, and there is no indication of it in the Lower Cretaceous thickness pattern (Rainwater, 1971; McFarlan, 1977) (Fig. 22). The strong trough developed in the Late Cretaceous and continued to subside slowly during the Early Tertiary (Cohee, 1961; King, 1969; King and Beikman, 1974), with the result that the Tertiary sediments nearly overlap the Creataceous (Figs. 9 and 32).

The embayment directly overlies and is aligned with a late Precambrian Rift system, the Reelfoot Rift (McGinnis, 1975; Ervin and McGinnis, 1975), shown as a large graben for lack of details (Fig. 33). The graben was found by magnetic and gravity surveys of the region, and it is being better defined as the geophysical data improve, (Braile and others, 1981; Johnson and others, 1980; Keller and others, 1980). A series of buried mafic intrusives are located along and outside the borders of the rift (Russ, 1981; Braile and others, 1981) and were probably emplaced during the Early to Middle Mesozoic (O'Leary and Hildenbrand, 1981) (Fig. 34). These are apparently similar to the volcanic centers along the borders of the Rio Grande and other large rifts, where their position is controlled by the border faults. Two

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other basement structures extend from the rift: the Rough Creek Graben in western Kentucky and the Ozark Dome (Fig. 33). Northwest trending zones of intrusive, and linear basement features also delinate a rift-like structure along the northeast side of the Ozark Dome (Fig. 34).

Post Late Precambrian subsidence over the Reelfoot Rift is shown by a coincident trough in the Precambrian surface (Buschbach, 1980; Fig. A-1). Subsidence occurred during the Cambrian and Ordovician resulting in the thickest deposits being laid down over both the Reelfoot Rift and adjacent Rough Creek Graben (Schwalb 1980, Fig. E-1). The Ozark Dome rose northwest of the rift during the mid-Paleozic and a southeast-trending ridge, the Pascola Arch, extended across the rift (Fig. 23) bowing up the basement surface. The rift apparently subsided again in the Early Pennsylvanian and then the entire region rose during the Permian (Rascoe and Adler, 1983). The Ordovician and Pennsylvanian depositional axes approximately coincide with the axis of the Late Cretaceous and Early Tertiany basin. This suggests that reactivation of the rift again formed a depression where Cretaceous and younger strata accumulated and then draped over the underlying rock as it deformed (Erwin and McGinnis, 1975). The faults and radar lineaments trend predominantly northwest and northeast, except in western Kentucky where easterly trends are common (Figs. 34 and 35). Most of the numerous pre-Cretaceous extensional faults that cut the Paleozoic rock around the north end of the embayment parallel the underlying late Precambrian structures and may have formed in response to subsidence along them. The zone of normal faults along the northwest side of the embayment step down to the southeast and are apparently related to relative downward movement over the rift (O'Leary and Hildenbrand, 1981).

Essentially all the displacement on the faults in the Paleozoic basement underlying the head of the embayment in southern Illinois occurred prior to the Cretaceous, although later deformation is suggested at several sites (Kolata and others, 1981). Sundeen and Baker (1980) suggest that the occurrence of volcanic rock in the Cretaceous strata in the embayment indicates renewed extensional faulting during deposition. Post-Cretaceous faulting is demonstrated in the New Madrid region where some faults offset Holocene deposits (Fisk, 1944; Krinitzsky, 1950; Satterfield and Ward, 1979; Stearns, 1979 and 1980; Russ, 1979; Zoback, 1979; Hamilton and Zoback, 1979) (Fig. 36). These generally trend northeastward parallel to the rift and embayment axes, as does the 11 mile (18 km) long Reelfoot Lake, formed during the 1811 and 1812 earthquakes (Fuller, 1912). The trend of the probable fault movement, accompanying the 1812 earthquake, that caused waterfalls on the Mississippi River above and below New Madrid is not well known (Fuller, 1912).

Various kinds of vertical movements have affected the region since the Cretaceous. The New Madrid area has been uplifted over 300 feet (100 m) since the early Tertiary as part of a broad regional movement across the mid-continent. There are, however, suggestions that the embayment itself could still be relatively subsiding at present. By latest Tertiary time the embayment must have attained a normal coastal plain aspect, however by mid-Pleistocene time the west side of the embayment had foundered and the Mississippi River marks the axis of present major down-drop; the structural form of the embayment north of Helena, Arkansas might now be considered a half-graben (O'Leary and Hildenbrand, 1981). The surface water of about a quarter of the United

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States drains through the embayment with many rivers converging at its head. The Mississippi River turns and flows southeast down the embayment, appoximately along its axis, where it enters it, and turns again to the south where it leaves it. Stretches of the Mississippi River are anomalous in the upper embayment. These stretches, which appear to lie adjacent to the positions where faults project across the river, are out of grade, the slope of the bottom being opposite the southward flow of the river (R.B. Whitney, written commun., and S.A. Schumm, oral commun., 1981). Evidence for both Holocene uplift and subsidence within the embayment near New Madrid is summarized by Stearns (1979). The Lake County Uplift, that consists of a north-northwest trending series of ridges and domes near New Madrid, appears to have recently rose, whereas, the Reelfoot Lake basin to the northeast has subsided (Fig. 37). Monoclinal structure and shallow faults have been located along the scarp between the uplift and basin (Stearns, 1980) (Figs. 36 and 37); a northward projection of this scarp crosses the Mississippi River near where one of the waterfalls formed in 1812 (Johnston, 1982).

Seismicity

A large number of moderate to major earthquake epicenters cluster in and around the head of the embayment. (Coffman and others, 1982) (Fig. 2). The largest earthquakes known to have occurred in the United States took place within the embayment near New Madrid, Missouri, in 1811 and 1812 reaching an intensity of XII (Fuller, 1912; Nuttli, 1973).

A very definite and consistent pattern of small earthquakes has been recorded in the region over the past nine years by an expanded

seismic network (Stauder and others, 1983) (Fig. 34 and 38). The pattern is one of a northeast-trending zone, just northwest of the Mississippi River, that shifts where it crosses a northwest-trending zone along the northwest Tennessee-Missouri border; seismic zones A and B, respectively of Johnston (1981). Some other events also occur scattered adjacent to this zone, especially to the north.

The seismicity is considered related to a Quaternary block fault subsidence pattern and the clear linear pattern of seismicity indicates the source zones can be interpreted as fault zones (O'Leary and Hildenbrand, 1981; Johnston, 1982). These epicentral zones show a correlation with the contoured surface of the Precambrian within the rift (Fig. 39). The northeast-trending zone lies along a trough in the Precambrian and the northwest-trending zone follows the northeast side of the Pascola Arch that crosses the rift (Fig. 23). The seismic zones may show the approximate trace of active faults within the rift and the direction of relative movement across the faults may be the same as the deflection of the Precambrian surface (Barosh, 1981) and present surface (O'Leary and Hildenbrand, 1981); the northeast side of the northwest-trending zone is relatively down.

The main sources of earthquake activity in the New Madrid region appear to be faults along the southeast side of the Reelfoot Rift and northwest-trending ones along the northeast side of the Pascola Arch with the greatest activity at their intersection near New Madrid. Much of the activity on the Pascola trend near New Madrid underlies the Lake County Uplift (Stearns, 1979; Johnston, 1982) (Figs. 37 and 38). The relative movement between this uplift and the Reelfoot Basin to the

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northeast is the same as the deflection on the underlying Precambrian surface (Fig. 39). A gravity map of this area (Stearns, 1979), shows a northwest-trending lineament, passing between New Madrid and Reelfoot Lake, that may reflect part of this fault zone. Northwest-trending normal faults along the Pascola Arch are exposed to the northwest in the core of the Ozark Dome when they are associated with some seismicity (Heinrich, 1937) (Fig. 34). Most of the activity on the Pascola zone dies out a short distance to the southeast in Tennessee (Fig. 52).

A recent swarm of small earthquakes northeast of Little Rock, Arkansas, lies just northwest of the border of the embayment (Johnston, 1983). These occur a short distance southwest of a recently studied area south of the white River where many northeast and northwest faults (Haley and others, 1976) and radar lineaments (O'Leary and Hildenbrand, 1981) intersect. A similar intersection may occur in the area of the swarm.

Many of the scattered epicenters around the head of the embayment occur in areas of normal faults (Fig. 34). The distribution (relative abundance) of these earthquakes shows a general correlation with the distribution of faults where they are mapped. The moderate to faint trends that can be envisioned in this epicenter distribution also matches the general fault trends (Figs. 34 and 38) A series of north-northeast trending normal faults has been mapped near the west end of the Ohio River and south of the Wabash River in Illinois and Indiana, the Wabash Valley fault system, (Sullivan and Ault, 1980; Kolata and others, 1981) and a zone of epicenters trends northeastward here (Coffman and others, 1982; Docekal, 1970). The epicenter distribution

suggests two or three northwest trending zones in southeastern Missouri to the Illinois border. Many of the faults and lineaments in this area strike to the northwest. The seismicity in northwest Kentucky, however, has, as yet, no clear pattern to demonstrate a correlation with the easterly-trending faults or northwesterly-trending ones there, although a Precambrian Rift underlies them (Figs. 33, 34 and 38).

An earlier suggestion that the earthquakes may be due to some kind of strain buildup on mafic plutons in the area (McKeown, 1975; Kane, 1977) ignored the indications that the plutons occurred on fault zones bordering the rift and that the zones, not the intrusions, were probably the important structure.

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Southwest Georgia Embayment

(less than intensity V)

Summary

The southwest Georgia Embayment began by the Late Jurassic as a sag over a large northeast-trending Late Triassic-Early Jurassic graben. The size of the embayment and subsidence decreased and mainly ended by Late Cretaceous and only a remnant remained near the Gulf shore by mid-Tertiary. No significant earthquakes occur along it, although a Miocene graben developed over its northeast end, and the embayment may have essentially ceased moving unlike the northwest-trending embayments on the East Coast.

Geology

The Southwest Georgia Embayment, sometimes referred to as the Apalachicola Embayment, is a Late Jurassic to Early Tertiary basin that plunges southwest across southwestern Georgia and the panhandle of Florida, west of Tallahassee, into the Gulf of Mexico (Fig. 9). It is bordered on the northwest by the Alabama Arch and on the east by the Peninsula Arch, and its extension, the Central Georgia Uplift (Suwannee Saddle), that forms a low saddle and separates it from the Southeast Georgia Embayment (Fig. 40). The basin probably began in the Late Jurassic (Chowns and Williams, 1983, Fig. 2) and was well developed in the Early Cretaceous when it extended well into southwest Georgia (Rainwater, 1971; Cramer and Arden, 1980)(Fig. 22) and then diminished

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with time. At the end of the Cretaceous it was largely confined to the Florida panhandle (Miller, 1982a) and by Oligocene-Miocene time to only the shore of the Gulf in the region of the Apalachicola River delta (Miller, 1982c). Subsidence in the basin essentially ended in the mid-Tertiary as the Southeast Georgia Embayment, to the northeast, expanded (Miller, 1982b; Cramer and Arden, 1980).

The embayment overlies and is aligned with a large complex buried northeast-trending Late Triassic-Early Jurassic graben, the South Georgia Rift or Basin (Chowns and Williams, 1983), that extends across panhandle Florida and southern Georgia and apparently controlled the position of the embayment (Fig. 41). The graben is set into metamorphic and granitic rock whose northeast-trending structural grain undoubtedly influenced its position and orientation. An apparent reactivation of the Mesozoic graben in Oligocene and Miocene time produced a northeast-trending graben system, the Gulf Trough over the center of the graben and northeast end of the embayment in the southwest co her of Georgia (Cramer and Arden, 1980; Miller, 1982b). The emb ment has apparently stopped developing and no major river follows the embayment, except at its southwest end, that might suggest any subsidence in the area.

Seismicity

No significant seismicity is known to be associated with the Southwest Georgia Embayment (Coffman and others, 1982; Stover and others, 1979) (Figs. 2 and 42). Only two earthquakes of less than magnitude 3.0 have been located in the area (Rinehart, 1983) (Fig. 43)

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and none have been recorded there since the southeast seismic network was improved in 1977 (Bollinger and Mathena, 1982). No significant source areas are indicated and movement in the embayment has probably stopped as suggested from its history of diminishing movement. The embayment lacks the northwest alignment and transverse position to the Cretaceous margin that the other active embayments on the east coast have. This embayment, thus, appears to represent subsidence that continued over a Mesozoic graben, but that has now stopped and may this apparent change indicate the decreasing influence with time of such grabens.

Southeast Georgia Embayment

(Charleston, S.C., 1886, I = X, M_b = 6.6-6.9)

Summary

The Southeast Georgia Embayment is a northwest-striking Late Cretaceous-Tertiary basin that has expanded with time and is indicated to still be subsiding. It appears controlled in part by a large northeast-trending Late Triassic-Early Jurassic graben, and more importantly, by northwest-trending structural features of at least Early Jurassic age. Shoreward projectures of the two large offshore northwest-trending Jurassic fracture zones bound the basin and appear to have controlled it, as they did the earlier Jurassic Blake Plateau Basin offshore, by downdropping of the block between them. Structure near the Fall Line, at the inner edge of the Cretaceous, controls the head of the embayment. Seismic activity occurs around the edge of the embayment and may be controlled by its underlying bounding structures. The greatest amount of activity occurs in a northwest-trending zone along the northeast side of the embayment, where the intensity X 1886 Charleston earthquake occurred. This earthquake most likely occurred from movement along northwest-trending fault along the Ashley River perhaps in combination with a northeast-trending fault.

Geology

The Southeast Georgia Embayment is a Late Cretaceous to Late Tertiary basin over eastern Georgia and southern South Carolina

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(Figs. 9, 40 and 44). The northern end of Peninsula Arch separates it slightly from the head of the southwest Georgia Embayment to the southwest and the broad Cape Fear Arch separates it from the Chesapeake-Delaware (Salisbury) Embayment to the northeast. The axis of the Southeast Georgia Embayment plunges southeastward into the curve of the coastline north of Florida approximately parallel to the major drainage of the region.

The embayment mainly developed towards the end of the Late Cretaceous (Brown and others, 1979; Cramer and Arden, 1980; Gohn and others 1980). It continued to subside during the Tertiary as shown by the much greater northwestward overlap of the Lower Tertiary deposits over the Cretaceous and the Upper Tertiary over the Lower along its axis (King and Beikman, 1974) (Fig. 44). In contrast, the Upper Cretaceous-Lower Tertiary contact swings southeastward across the Cape Fear Arch and the Upper Tertiary deposits are missing onshore. The Paleocene to Miocene sediments thicken greatly across its axis near Brunswick, Georgia on the coast (Miller, 1982a).

The embayment appears to have increased in size during the Tertiary. Its southwest side shifted from northeast to central peninsula Florida from about the Late Cretaceous (Miller, 1982b) to the Oligocene (Miller, 1981c) as the Peninsula Arch shifted from a position down the center of the Florida peninsula to its west coast (referred to as the Ocala Arch in this position) (Miller, 1982c). This westward shifting of the edge of the embayment and arch is at the expense of the Southwest Georgia Embayment. The embayment was very well developed in the Miocene (Riggs, 1984). The northeast-trending narrow sag forming the Suwannee Trough across the north end of the Florida peninsula also

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disappears in the mid-Tertiary (compare Miller 1982a, b and c; Cramer and Arden, 1980). It appears, thus, that during the early to mid Tertiary, northwest-trending sags developed as northeast-trending ones diminished.

The position of the Pliocene deposits near the axis of the Southeast Georgia Embayment and the greater inland extent of the Quarternary deposits across it (King and Beikman, 1974) as well as relative downwarping of Pleistocene shoreline features across its axis (Winkler and Howard, 1977) (Fig. 8) suggest the embayment has continued to subside to the present, relative to the Cape Fear and Peninsula Arches, although the entire region may be rising slightly.

The embayment overlies Precambrian granites and metamorphic rock of mainly the Piedmont province of the Appalachian orogen (Fig. 44). This province underwent a considerable amount of generally compressive deformation, especially in the early Paleozoic. The structures predominantly trend northeastward crossing the Mesozoic and Tertiary embayments and arches of the region nearly normal to their axes (King, axes (King, 1969; Popenoe and Zeitz, 1977) (Fig. 44). These early structures appear to have had little influence in controlling the transverse sag across the Paleozoic rock beneath the Southeast Georgia Embayment.

Two opposing trends of pre-Late Cretaceous Mesozoic extensional structures, a graben and dike system, seemingly influenced the development of the embayment. The large Late Triassic-Early Jurassic graben, the South Georgia Basin or Rift (Chowns and Williams, 1983) (Figs. 41, and 45), that underlies southwest Georgia extends northeastward under the embayment. A narrow, Oligocene and Miocene

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graben system is superimposed over the earlier one in Georgia (Cramer and Arden, 1980; Miller, 1982a and c) and may be a partial reactivation of the earlier one. The presence at this large graben in the basement may have had some controlling influence on the location and subsidence of the embayment as it apparently has on the Southwest Georgia Embayment, although the trends of the embayment are different. The graben swings eastward into the mid South Carolina coast (Chowns and Williams, 1983) and does not continue northeastward beyond the embayment. Two northwest-trending offsets of the graben border in southern Georgia (Chowns and Williams, 1983) lie near the southwest side at the embayment (Fig. 45). A northwest-trending dike system crosses the region, parallel to the embayment axis, and a concentration of them lie along the northeast flank of the embayment (Lester and Allen, 1950; May, 1971; Popenoe and Zeitz, 1977; Daniels and others, 1983, Fig. 8). They are probably Early Jurassic, about 180 M.Y. and not older than 195 M.Y. (Dooley and Wampler, 1983). This straight dike swarm indicates a controlling fracture system. The dike concentration is aligned with the offshore Blake Spur fracture zone and is probably related to it (Fig. 44). Northeast of this fracture concentration the dikes trend northerly beneath the Cape Fear Arch (Daniels and others, 1983) as does a probable volcanic chain, recognized by geophysical studies, just offshore (H. Krivoy, written commun., 1978). This change in dike trend may indicate a change in the orientation of local strain across a zone of movement.

The Blake Spur is one of a series of northwest-trending oceanic fracture zones along the Atlantic Coast, that began developing during the Jurassic as the Atlantic Basin opened (Klitgord and Behrendt, 1979;

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Klitgard and others, 1983) (Fig. 46). They offset or bound large magnetic anomalies and sedimentary basins along the eastern edge of the continent. The two larger fracture zones in this region, the Blake Spur and Jacksonville (Abaco), bound the Blake Plateau Basin and the block between them sank more than 7 km during the Jurassic (Dillon and others, 1983) (Figs. 43 and 46). The onshore projecture of these two fracture zones bound the Southeast Georgia Embayment (Fig. 44).

It appears that although both northeast and northwest-trending Mesozoic structures apparently influenced the embayment. The principal effect of the northeast-trending ones was apparently earlier and control by northwest- trending features, especially fracture zones, was dominant by at least mid-Tertiary and are probably the important ones at present.

Many Cenozoic normal and a few reverse fault zones now have been found in the region and some may have Holocene movement. These occur as scattered faults near the inner edge of the Coastal Plain, a graben system within the central Coastal Plain and a series of faults along and offshore of the Atlantic Coast. The onshore pattern of faults in Georgia and South Carolina consists of both northeast and northwest trends similar to the underlying early Mesozoic structural trends. A northeast-trending graben system, the Gulf Trough, active in the late Oligocene and Miocene, crosses the embayment through the middle of the Coastal Plain (Cramer and Arden, 1980; Miller, 1982 a and c). It overlies and may be a partial reactivation of the central portion of the early Mesozoic South Georgia Graben. A few northeasterly trending, mainly reverse, faults have been found near the Fall Line at the inner edge of the Cretaceous (Wentworth and Mengner-Keefer, 1983). One of

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these, the Belair fault, cuts mid-Tertiary rock at the northern edge of the embayment on the Alabama-South Carolina border (Prowell and O'Conner, 1978). The southwest projection of this fault follows a slight change in thickness at Lower Tertiary rock (Miller, 1982a) that suggests it continues to the Gulf Trough. Some probable mid-Tertiary faulting has also occured in the Piedmont (White, 1965).

High-angle northwest-trending faults cut the coastal plain of Georgia, including significant ones, that cross the middle of the state approximately where the South Georgia Graben is offset (Cramer and Arden, 1980) (Fig. 45). Several short northwest-trending Cenozoic faults have also been found offshore of Charleston (Behrendt and others, 1983, Fig. 1) and a probable northwest-trending fault follows the Ashley River in the epicentral area northwest of Charleston (Talwani, 1982). Northwest-trending faults may be common as both the gravity and aeromagnetic data (Long and Champion, 1977; Popenoe and Zietz, 1977; Daniels and others, 1983 and Klitgord and others, 1983) show regional northwest-trending lineaments apparently not caused by dikes, to extend inland from offshore in the general vicinity of both the Blake Spur and Jacksonville fracture zones . These lineaments may represent faults or fault zones. Such lineaments pass through the epicentral area of the Charleston earthquake.

Many faults lie near and parallel to the coast. A series of near vertical post-Late Eocene and at least one post-Oligocene faults lie along the east coast of Florida, parallel to the shore; with, north-northwest trends in the south and northerly trends in northern Florida and southern Georgia (Miller, 1982c). Other small northeast-trending near vertical faults parallel the coast and cut Upper

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Cretaceous and Lower Tertiary rock in the epicentral area of the Charleston earthquake (Hamilton and others, 1983). The faults near Charleston are mainly down to the southeast and both normal and reverse ones are present (Hamilton and others, 1983; Schilt and others, 1983). Some are also found just offshore of South Carolina and Georgia, including the long northeast-trending Helena Banks fault that lies offshore of Charleston and is as young as Miocene or Pleistocene (Behrendt and others, 1983), and a major fault that displaces Lower Cretaceous strata off of Georgia (Cramer and Arden, 1980).

The faults appear to be tensional features (Cramer, 1969). The northeast-trending faults of the Gulf Trough and reverse ones at the inner edge of the Coastal Plain are probably both due to differential vertical movements of the Piedmont rising relative to the central Coastal Plain. Faults bordering uplifted blocks, such as the Piedmont, are commonly reverse near the surface, although they may be normal at depth (Matthews, 1978). The northwest-trending faults could also be due to minor vertical adjustments, although the offshore fracture zones, to which they may be somehow related, have some lateral movement. The faults paralleling the coast, on and off shore, appear similar to those found along the Gulf Coastal Plain (Renfro, 1973; Berryhill and Trippet, 1981) where they are generally interpreted to be growth faults caused by a slight seaward shift due to the sediment load and not necessarily first-order tectonic features.

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Seismicity

The patterns of historical seismicity (Hadley and Devine, 1974; Bollinger and Visvanathan, 1977; Tarr and Rhea, 1983; Stover and others, 1979; Reagor and others, 1980 a and b) and recent instrumentally recorded earthquakes (Tarr, 1977; Tarr and Rhea, 1983; Bollinger and Mathena, 1982) in the southeastern U.S. are very similar and the persistent zone of recent activity to the northwest of Charleston may be associated with the source of the 1886 earthquake of Intensity X, M_b 6.6 - 6.9, (Nuttli and others, 1978; Bollinger, 1977; Tarr, 1977; and Tarr and Rhea 1983) (Figs. 2, 3, 58, 59 and 70). No significant change in pattern is discernible in the source zones with time. The present activity northwest of Charleston is thought to represent aftershocks of the 1886 earthquake by Tarr and Thea (1983), but further studies indicate that the level of seismicity decreased rapidly after the 1886 event and in three years reached what appears to be a longer-term background level (Armbruster and Seeber, 1983). Earthquakes occurred in the Charleston area prior to 1886 (Bollinger, 1983) and the overall pattern would not be changed by eleminating the Charleston earthquake and associated aftershocks.

The regional pattern in the southeast United States around the embayment is one of clusters of activity in South Carolina and central Virginia, a diffuse northeast-trending cluster along the axis of the Appalachian Mountains from western Alabama to northern Virginia and the absence of significant activity in eastern North Carolina and southern Georgia to Florida (Figs. 2, 43 and 47). The cluster of activity in South Carolina has a general northwest alignment that passes through Charleston from both historical and recent activity. This was first noted by Hobbs in 1907 and by many others since then (Wollard, 1969;

Bollinger, 1973) (Fig. 47). It has been referred to as the South Carolina - Georgia or South Carolina seismic zone. The effects of several earthquakes, as shown by intensity distribution also show this northwest trend (Bollinger and Visvanathan, 1977; Tarr, 1977; Bagwell, 1981; Bollinger, 1983) and a few focal mechanisms of earthquakes northwest of Charleston also suggest movement along northwest-striking near-vertical planes (Tarr, 1977; Tarr and Rhea, 1983). The earthquakes in the South Carolina seismic zone do not, however, form a clearly defined zone as in the Mississippi Embayment, but occur in clusters with gaps between them. The zone also widens considerably where it crosses the Fall Line, at the edge of the Coastal Plain, and enters the Piedmont (Bollinger and Mathena 1982; Tarr and Rhea 1983; Rinehart, 1983) (Figs. 42, 43, 47 and 48). Here scattered activity extends along near the Fall Line from central Georgia northeastward to the North Carolina border. Several events have also occurred outside this zone in southern Georgia and northeast Florida.

This activity in South Carolina, Alabama and northeast Florida is located near the borders of the Southeast Georgia Embayment with concentrations of events along its northeast flank and near head in the Piedmont (Barosh, 1981). There is little activity in the areas of the Cape Fear and Peninsula arches, but activity is beyond the Cape Fear arch in the Chesapeake-Delaware (Salisbury) Embayment of central Virginia (Figs. 2, 42, 43, 44 and 47). The pattern of seismicity in the Coastal Plain and Piedmont changes at the Brevard fault zone, which appears to form the border of the northeast-trending cluster of events of the southern Appalachian Mountains (Barosh, 1981) (Figs. 42 and 47). The Brevard zone also forms the boundary of the northwest-trending Early

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Jurassic dike system cutting the Piedmont (Lester and Allen, 1950; Popenoe and others, 1977).

The Southeast Georgia Embayment is probably still subsiding and this may be largely accomplished by movements in the pre-Cretaceous rock along its borders: the extensions of the Blake Spur and Jacksonville fracture zones and faults near the inner edge of the Cretaceous. Much more activity occurs along the northwest flank adjoining the Cape Fear Arch, just as the activity in the Chesapeake-Delaware Embayment to the northeast is also concentrated towards the arch. The local control of the 1886 Charleston earthquake may well be due to movement on a northwest-trending normal fault along the Ashley River (Bollinger, 1983) as is suggested by the northwest elongation of recent earthquake effects along the river, P-wave motion data and geophysical lineaments. Bollinger (1983) considers the activity in the epicentral area to represent dip slip movement on this northwest-trending fault, but that the intensity distribution of the 1886 earthquake favors a northeast-striking feature and that both trends may be important. Thus, the activity near Charleston may well be localized by a structural intersection.

Chesapeake-Delaware Embayment

(Arvonia, VA, 1875, I = VIII)

Summary

The Chesapeake-Delaware Embayment is a depositional basin that began by the start of the Cretaceous, continued into the Late Tertiary and is probably still subsiding. It has had a complex history, but its latest, Late Miocene, depositional axis trends northwest from near Norfolk through Richmond, Virginia, where it crosses the north trending Central Fall Line. If projected farther, this axis would continue along the James River northwest of Richmond. The embayment appears controlled by both a large buried northeast-trending graben system and the shoreward projection of the northwest-trending Norfolk fracture zone. This projection coincides with the Late Miocene depositional axis. The central Virginia seismic zone, which lies over the James River from Richmond northwestward, is approximately coincident with the projection of the fracture zone, and depositional axis and is thus apparently controlled by them. Perhaps subsidence, between the central Fall Line and Brevard zone to the northwest over this structural zone is causing the earthquakes. A minor amount of activity also occurs in southern Delaware near the onshore projection of the Delaware fracture zone.

Geology

The Chesapeake-Delaware (Salisbury) Embayment (Murray, 1961) is a Cretaceous-Tertiary basin centered over the Chesapeake Bay region northeast of the Broad Cape Fear Arch (Clark and others, 1912) and

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southeast of the much smaller Normandy Arch in southern New Jersey (Brown and others, 1972) and a subtle trough over Delaware Bay (Figs. 9 and 44). The arches trend northwest, but the axial trend of the embayment has varied with time. The axis shows up strongly in the structural surface of the pre-Late Jurassic with approximately a west-trend, but has had north and northeast-trends as well as northwest-trends at different intervals in its development. Murray (1961) describes its general features and Brown and others (1972) present a detailed and complex history from the Late Jurassic to the end of the Tertiary. Brown and others (1972) ascribe the complexities as due to alternations of compressional features and grabens forming during sedimentation under a wrench fault regime. They think the present Chesapeake Bay may be subsiding as a north oriented graben. Despite the variation in trend and slight local shifts in depositional centers with time within the embayment, its position on the northeast flank of the Cape Fear Arch has not changed and it occupies a position comparable with the Southeast Georgia Embayment, on the southwest flank (Fig. 44). The Chesapeake-Delaware Embayment was well developed in the Miocene, the latest period for which there is good control, with a northwest trending axis (Brown and others, 1972) as was the Southeast Georgia Embayment. This axis lies across the southern side of Chesapeake Bay approximately on line with the offshore Norfolk fracture zone (Fig. 49). The flooding of much of the embayment by the waters of the Chesapeake Bay suggests it may still be subsiding. Brown and others (1972) considered the bay to be a subsiding north-trending graben. Precise releveling indicates a broad contemporary downwarping of the Chesapeake-Delaware Embayment (Brown 1978) with the maximum subsidence

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indicated about 25 miles west-southwest from Norfolk, Virginia, (Balazs, 1974) near the axis of the Miocene downwarp. A second subsiding center within the embayment is at Delaware Bay (Holdahl and Morrison, 1974) which is also the site of a slight depositinal trough (Brown and others, 1972) (Figs. 49 and 50). The mouth of Chesapeake Bay is relatively higher than the northern part of the bay (Harrison and others, 1965), but the entire Bay is subsiding (Holdahl and Morrison, 1974).

The embayment overlies a large buried Late Triassic-Early Jurassic graben system, part of which is exposed as the Richmond Basin near the head of the embayment (Fig. 49). This graben system undoubtedly controlled in part the development of the embayment and episodic movement in the graben system may be responsible for some of the shifting depositional trends. Other grabens are exposed in the Piedmont around the head of the embayment (Fig. 49). The graben system was almost certainly controlled by underlying structures in the granitic and metamorphic rock terrane of the Piedmont province, but these structures probably had no direct influence on the embayment.

A major offshore fracture zone, the Norfolk, trends northwestward into the southwest side of the embayment (Figs. 46 and 49). The change in dispositional pattern across it at sea is mimicked by those across its onshore projection. The offshore depositional troughs, the Carolina and Baltimore Canyon, are offset left-laterally across the fracture zones, similar to the shift of the major onshore depositional wedges of Cretaceous and Tertiary deposits (Brown and others, 1972) from the outer Coastal Plain of North Carolina to Virginia. Even the change in trend of the East Coast magnetic anomaly from northeast to north across the fracture zone is mimicked by the change in trend of the Fall Line at

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the inner edge of the Coastal Plain deposits. The Miocene axis of the embayment parallels the projection of the fracture zone and the Late Miocene depositional trough (Brown and others, 1972) overlies it. The Norfolk fracture zone, thus, appears to have had significant influence on the Chesapeake-Delaware Embayment in determining its position and structure. Other slight shifts in offshore boundary of the Baltimore Canyon Trough are also over projections of smaller fracture zones and correlate with changes onshore. One aligns with the northeast side of Chesapeake Bay, another with the slight sedimentary trough in Delaware Bay.

Several post-Cretaceous faults are now known in the region and many others have been inferred, some cut the Coastal Plain, others lie along the Northern Fall Line and others occur offshore. Most strike northeastward. In addition, some northwest-trending joint zones cut the Coastal Plain (Thompson, 1983) and many northwest-trending faults cut the Late Triassic - Early Jurassic grabens in the adjacent Piedmont (Fig. 51), also several north-trending LANDSAT lineaments cross Delaware and the eastern shore of Maryland (C. Withington, oral commun., 1976). The onshore data have been compiled by Barosh and Pease (1979, 1981b), Thompson (1983) and Wentworth and Mergner-Keefer (1983). Most of the faults are near vertical. Some along the Fall Line southwest of Washington, D.C. are reverse (Fig. 49). Those offshore are normal and appear to be growth faults, due to gravity, (Grow, 1981). Some normal faults have been mapped (Spoljaric, 1973) and many other normal faults are postulated to have been active at times in the Coastal Plain during the Cretaceous and Tertiary from analysis of the depositional patterns; these are thought to be due to left-lateral movement across the area (Brown and others, 1972).

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The Northern Fall Line from Delaware northeastward is now known to be an Early Paleozoic fault zone with right-lateral movement that was reactivated in the early to mid Mesozoic with probable left-lateral movement and again in post-Cretaceous time, as shown by analysis of the geophysical and LANDSAT data (Barosh and Pease 1976, 1981) and follow up field studies (Thompson, 1983; Tillman, 1983). The northwest-trending fault and joint fractures appear to have the more recent movement (Thompson, 1983).

Seismicity

Almost all of the earthquake activity from Delaware to southern North Carolina is concentrated in central Virginia at the southeast side of the Chesapeake-Delaware Embayment and along the Blue Ridge of western Virginia and North Carolina (Figs. 2 and 43). This forms the Central Virginia seismic zone and part of the northeast-trending Southern Appalachian seismic zone respectively (Fig. 47). This definite pattern is the same on compilations of historical earthquakes (Coffman and others, 1982; Tarr and Rhea, 1983) and recent monitoring on the improved seismic network (Bollinger and Mathena, 1982) (Figs. 42 and 43). Eight events of intensity V or more have occurred in the region near the James River, from the Richmond area west to the Brevard zone, and intensity V occurred a little outside this cluster at Milford, north of Richmond and at Charlotte Court House to the south (Coffman and others, 1982) (Figs. 42 and 43). The most active area is in the western part near Arvonia which has experienced intensity VI and VII effects (Coffman and others, 1982).

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Elsewhere in the embayment away from the Brevard-Northern Fall Line zone, there is very little activity, earthquakes of intensity V or greater are recorded for northeastern North Carolina and only two of IV-V in central Delaware (Coffman and others, 1982).

The Central Virginia seismic area appears related to the latest subsidence in the region. The depositional trough of the Upper Miocene sediment strikes northwestward from south of Norfolk, along the southwest side of the lower James River to Richmond and its projection northwest would continue along the James River. This coincides with the landward projection of the Norfolk fracture zone. The activity is not distributed every where along this zone but only at its head, northwest of the central Fall Line (Figs. 42 and 43). Small earthquakes occur near the Fall Line and probably indicate some related movement along this structure near the intersection.

The only other significant activity in the embayment appears to be in eastern Delaware (Stover and others, 1981) adjacent to the slight sedimentary trough near Delaware Bay and the onshore projection of the Delaware fracture zone.

The activity along the northwest side of the embayment follows the Brevard-Northern Fall Line zone, which will be discussed under the Appalachian Highlands.

Raritan Embayment

(Jamaica Bay, N.Y., 1884, I = VI)

Summary

The Raritan Embayment is a northwest-trending Cretaceous-Tertiary basin across the bend in the inner Cretaceous boundary around eastern Long Island. It appears controlled by both a buried northeast-trending graben, controlled in turn by earlier structures, and an extension of the northwest-trending offshore fracture zone that crosses the north end of the Baltimore Canyon Trough. The embayment is still subsiding and the major seismicity is concentrated in its center, apparently due to this movement. Earthquakes may be controlled by both northeast and northwest-trending structures in the embayment.

Geology

The Raritan Embayment is a southeast plunging Cretaceous-Tertiary basin heading in Raritan Bay between the north end of the Coastal Plain and the southwest end of Long Island (Figs. 9 and 52). The inner Cretaceous contact makes sweeping bend around its head, changing from a northeast-trend along the Northern Fall Line in northern New Jersey to a nearly east trend along its submerged position in Long Island Sound. The embayment is reflected on the basement surface (Brown and others, 1972; King, 1967; Watts, 1981), but its history is not as well known as that of the Chesapeake-Delaware Embayment due to lack of stratigraphic

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data. It was well established by the end of the Early Cretaceous northeast of the conspicuous thining over the axis of the Normandy Arch, that separates it from the Chesapeake-Delaware Embayment to the southwest. Its depositional axis has varied similar to that of the Chesapeake-Delaware Embayment. A slight northwest-trending sedimentary trough is indicated during the Early Cretaceous, but at the beginning of the Late Cretaceous a slight northeast-trending sedimentary trough crossed the outer edge of the bay (Brown and others, 1972).

A Late Triassic-Early Jurassic graben underlies the embayment as is the case with the Chesapeake-Delaware Embayment (Fig. 52). The embayment is also approximately aligned with a northwest-trending offshore New Jersey fracture zone that extends across the north end of the offshore Baltimore Canyon Trough, (Fig. 45) just as the Chesapeake-Delaware Embayment is aligned with fractures at its south end (Figs. 45 and 52). The graben and fracture zone have both controlled the embayment at times as is shown by sedimentary troughs of both trends. The embayment conducts the Hudson River to the sea and is indicated to be subsiding at present by tidal gage measurements near the mouth of the bay at Sandy Hook, New Jersey (Walcott, 1972).

A great number of Paleozoic and early Mesozoic faults are in the region (Manspizer, 1981; Thompson, 1983; Tillman, 1983) and most of the Mesozoic ones were controlled by the older faults. The pattern of normal fault movements in the Late Triassic - Early Jurassic Newark Basin (Figs. 52 and 53) suggests they were formed under a left-lateral shear oriented northeast (Manspizer, 1981). Such normal faulting is to have continued during Cretaceous and Tertiary sedimentation on the Coastal Plain (Brown and others, 1972). The Northern Fall Line at the

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head of the embayment is now shown to be a fault zone with post-Cretaceous movement, as indicated by both field and geophysical data (Barosh and Pease, 1979, 1981; Tillman, 1983; Thompson, 1983). A north-northeast-trending vertical fault, with east side down cuts the Cretaceous a short distance offshore of the mouth of Raritan Bay (J. Grow, oral commun., 1981) and many other normal, growth, faults cut it further out to sea (Grow, 1981). The youngest faults, found cutting the Northern Fall Line zone in southwest Connecticut, trend northwest (Tillman, 1983) as due joint and fracture zones cutting the Cretaceous in New Jersey (Thompson, 1983). Northwest-trending magnetic lineaments, that may represent faults, are present offshore of northern New Jersey and Long Island.

Seismicity

The Raritan Bay area is presently the most seismically active area in the northern New Jersey-southeastern New York region, despite the numerous small earthquakes recorded by the greater number of seismographs near the north end of the Newark Basin (A Kafka, oral commun., 1982) (Figs. 2, 54 and 55). The Ramapo fault bordering the northwest side of the basin does not appear to be anymore associated with seismicity than the myriad of other faults in the region (Fischer and McWhorten, 1978; Fischer, 1981) (Fig. 52). The Raritan Bay area has been the center of the most widly felt earthquakes in the region (Nottis, 1983). The large historic earthquakes recently have been restudied and relocated by Nottis and Mitronovas (1983) and found to be concentrated more towards the northeast side of the Bay (Fig. 55).

The seismicity appears due to continued subsidence in the embayment perhaps the result of movement on an underlying northwest-trending fracture system. Such movement, however, might locally reactivate segments of northeast-trending faults intersected, the Fall Line in particular. This might be similar to the local activity along the Fall Line where crossed by the northwest-trends in the Southeast Georgia and Chesapeake-Delaware embayments. The epicenters of the several larger earthquakes in the Raritan bay area do appear to have a northeast alignment parallel to the underlying graben, older structures and gravity features.

The Hudson River extends northward from Raritan Bay and is apparently controlled by a system of fractures (Y.W. Isachsen, written commun., 1981). This geometric pattern is similar to that made by the Rio Grande Rift and Nemaha Uplift extending northward from their associated northwest-trending fracture zones.

The zone of earthquake activity across northern New Jersey (Stover, Barnhard and others, 1980) and into southwestern Connecticut is somewhat symmetrical about the head of the embayment and may be caused by movements periferal to those in the embayment. This area, however, lies at the north end of the Southern and Central Appalachian Upland zone of seismicity, under which it will be discussed.

Moodus

(Moodus, CT, 1791, I = VII)

Summary

The earthquakes at Moodus, Connecticut, occur in an area of unusual structural complexity, where several major fault and fracture zones converge, that has a long history of tectonic activity beginning in the late Precambrian. Two zones may extend through the crust: the very long northeast-trending Early Jurassic Higganum diabase dike system, and a northwest-trending fault and geophysical lineament zone that lies along the Connecticut River and may extend a considerable distance seaward. The well—located earthquakes occur east of the junction of these two zones in an area where apparently small northwest-trending faults cross a zone of small northeast-trending faults and joints. Movement on these faults could be caused by distortion of the rock across the major northwest-trending zone, perhaps in conjunction with deep fractures near the dike zone.

Geology

The Moodus area lies in a complex region composed of six distinct structural provinces: the Hartford Graben, Killingworth Dome, Glastonbury Dome, Merrimack province, Southeast Connecticut Fold Belt portion of the Southeast New England Platform and a thin remnant of the Nashoba Thrust Belt (Fig. 56). Major fault zones separate these provinces (Fig. 56). The Honey Hill fault zone is a north-dipping

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thrust fault that, along with a sliver of the Nashoba Thrust Belt, separates the Merrimack province from the Southeast Connecticut fold belt and has local relative north over south movement. It is part of the largest fault system known in New England and probably represents an early Paleozoic plate boundary that may have begun movement as early as late Precambrian (Barosh, 1982). Most of the length of the fault bounding the north side of the Nashoba thrust sliver along the Honey Hill fault has been intruded by the Devonian Canterbury Gneiss. The Nashoba thrust belt is much wider along the eastern boundary of Connecticut (Barosh, 1982; Pease, 1982)

The Bonemill Brook fault zone is a northerly-trending, steeply dipping zone dividing the Merrimack province from the provinces to the west (Pease, 1982). The Honey Hill fault zone is cut off on the west by a northerly-trending fault sliver of steeply dipping deformed rock about 800 m wide that lies on the east side of the Bonemill Brook Fault. The Bonemill Brook fault zone has not been mapped south of the Falls River fault zone (Fig. 56); it may continue southward, but the structure is extremely complex and is as yet unresolved. What happens west of this zone to the Honey Hill and the underlying rocks of the southeast Connecticut fold belt is as yet unknown. However, the kinds of structures at the southern end of the Killingworth Dome area are very similar to the kinds of structures south of the Honey Hill and a continuation of the fault may lie in this area.

A broad northwest-trending zone of structural dislocation divides the Glastonbury and Killingworth Domes. This zone extends northwestward from the Moodus area mainly along the northeast side of the Connecticut River. Major structural features on either side of this zone terminate

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against it. The left-lateral Middle Haddam fault zone lies along the northeast side of the structural zone and a series of older gently north-dipping thrust faults lies mainly south of the river.

A normal, extensional, border fault along the east side of the Hartford Graben separates the graben from the domes to the east. This fault appears to be a composite fault zone; northeast-trending segments of the fault appear to extend into the Killingworth Dome. A strong linear gravity high follows the border fault (Kick, 1982) (Fig. 57). It is highest where the thick lava flows in the graben abut the border and may represent a basaltic intrusion that came up the fault and fed the flows.

The provinces exhibit different structural styles and contain separate stratigraphic sequences and, except for continuation of the Monson Gneiss across the boundary between the Killingworth and Glastanbury Domes, correlation between provinces is mainly conjectural.

The Moodus seismic area lies mainly near the western border of the Merrimack province, where a number of diverse structural features of different ages and deformational styles converge (Barosh, 1981a). The deformational styles vary with both age and structural province. The ductile and mixed ductile and brittle features appear to be all pre-Mesozoic, and the brittle features Mesozoic or younger; this appears to reflect a greater depth of formation for the former. These characteristics are of great help in separating structures by probable age. Considerable information is available on the relative ages of pre-Mesozoic structures but little is definitely known about their absolute ages.

The broad northwest-trending and gently plunging axis of the Moodus anticline deforms rock of the Hebron formation and passes east of the

town center (Fig. 58). Much of its northeastern limb is covered by very gently dipping rock of the Brimfield Group that apparently has been thrust slightly over it. Its southwest limb gradually steepens to moderate to vertical dips to the west where it meets and is cut off by the Cremation Hill fault zone (Fig. 59). The Hebron formation is highly deformed adjacent to the fault and is intruded by many irregular pegmatite bodies. The pegmatites formed during deformation, probably not as an anatectic melt, but under near melt conditions (London, 1982). Elsewhere, the bedding within the Hebron formation is generally undeformed, except for local zones that are contorted and cut by small reverse faults (Fig. 59). The faults are mostly axial planar shears that strike north and display east over west transport. These small zones of deformation lie in and appear to control the location of many of the small valleys and gullies in the area.

The Cremation Hill fault zone has been feldspathized under conditions approaching a melt and is expressed at the surface as a zone of migmatites. Sheared and rotated blocks occur within this foliated migmatitic zone. Structures along and within the fault zone indicate right-lateral movement with the east side up. However, a few local exceptions are present. The Cremation Hill fault zone may be part of the same general zone of movement as the Bonemill Brook fault zone.

The Bonemill Brook fault zone lies to the west of the Cremation Hill fault zone and is separated from it by slivers of gneiss and schist that appear to belong to the Brimfield Group (Pease, 1982). The Bonemill Brook fault zone forms the eastern boundary of the Monson Gneiss that occurs in both the Glastonbury and Killingworth Domes.

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The north-striking Injun Hollow fault zone lies west of the Bonemill Brook fault zone in the Moodus area (Figs. 58 and 59). It is a steeply dipping reverse fault with indications of left-lateral movement. Movement along it has continued over a long period and both early ductile and late brittle features are found. The movement along it and some of the other faults have sheared the rock into interleaved fault-bounded lenses with little coherent order.

The style and orientation of folds and foliation are different within the Cremation Hill fault zone from those to the east and west (Fig. 60). The foliation trends across the Injun Hollow fault zone are also different (Fig. 60).

South of Moodus the Bonemill Brook fault zone is cut by several small thrust faults that dip about 45° north and have north over south relative movement (Fig. 58). They commonly curve from a westerly to a northwest trend as they are followed westward. The larger ones make curving valleys crossing the northerly-trending ridges and valleys that reflect differential erosion of the northerly striking lithologic units. These thrust faults and fractures related to them are commonly invaded by post-tectonic Permian pegmatite. Roadcuts in the area expose great numbers of these north-dipping pegmatites. The largest of these faults is the Falls River fault that passes through the towns of Centerbrook and Ivoryton and terminates several units on the north side (Figs. 56 and 58) (Wintsch, written commun., 1981); this termination was previously considered the nose of the Ivoryton synform (Lundgren, 1964).

The north end of the Killingworth Dome to the west is a fairly well defined, large, north-plunging anticline (Eaton and Rosenfeld, 1972;

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Lundgren, 1979); in contrast, its southern end is very poorly defined and it may not be a dome. A very gently north-dipping thrust fault trends northeastward across the anticline toward the Moodus area, but appears to end against or merge with a northwest-trending thrust fault west of the river. The very low angle of this fault, 10°-25° N, and associated syntectonic pegmatite suggest it is older than the westerly-trending thrust faults mentioned above. The amount and sense of movement on this thrust are not known as its observed trace is entirely within one stratigraphic unit.

A parallel and probably similar northwest-dipping fault lies to the northwest in the Haddam quadrangle and yet another farther northwest in the Durham quadrangle. The trace of the later one is closely followed locally by the Mesozoic border fault of the Hartford Graben (Fig. 58).

A steep 3-milligal gravity gradient forms a prominent northwest-trending linear feature along the northeast side of the Connecticut River between the Glastonbury and Killingworth Domes (Kick, 1982) (Fig. 57). Its northwest end is the Middle Haddam fault zone (Fig. 56). The anticlinal nose of the Killingworth Dome and the Great Hill syncline and adjacent west flank of the Glastonbury Dome end against it. The Maromas Gneiss, south of the fault, is a strongly foliated left-laterally offset portion of the Middle Ordovician Glastonbury Gneiss to the north (M.H. Pease, Jr., oral commun., 1981). Strong northwest-trending joint sets are present near the fault (J.S. Sawyer, written commun., 1979) and prominent northwest-trending topographic and LANDSAT lineaments lie in the zone.

The Middle Haddam fault zone does not appear to offset the Bonemill Brook fault at the surface. The northern boundary appears to merge with the Bonemill Brook fault where it makes a bend to the southeast. This junction is near the seismically active area. The gravity and topographic features apparently associated with the Middle Haddam fault zone continue southeastward down the Connecticut River and may represent faulting not exposed at the surface. Both gravity and aeromagnetic features are offset left-laterally across this zone down the Connecticut River. The gravity anomalies appear to be offset 5 to 6 km left-laterally across this zone (Kick, 1982) which is approximately the offset of the Maromus across the fault. The topographic features could have resulted from reactivation of a basement structure in the Mesozoic. The Selden Neck fault appears to be a small, late, left-lateral feature down the lower Connecticut River and a small fault may lie in the river in the Haddam quadrangle (Fig. 58). These, however, appear to be much too small to explain the geophysical features. Another feature suggesting a deep structure is a northwest-trending belt of small ultramafic bodies that lies near the Connecticut River (Sawyer, 1979) (Fig. 58). This belt follows approximately the Middle Haddam and the southern part of the Bonemill Brook fault zones. These bodies may indicate the former presence of deep fractures along these zones. Aeromagnetic lineaments associated with the southern part of this zone and the southern, northwest-trending, part of the Bonemill Brook extend southeastward across Long Island Sound (Barosh and others, 1977). These may continue seaward and join northwest-trending magnetic lineaments present south of Long Island.

The northeast-trending, west-dipping to vertical Early Jurassic Higganum diabase dike zone is the most prominent known Mesozoic feature passing through the Moodus area. This zone of dikes extends from Long Island Sound to near Portland, Maine, and apparently represents a deep fracture zone in the crust, although it does not necessarily follow a fault zone at the surface. The dike locally follows the surface trace of an early thrust fault west of Moodus (Fig. 58). The local northwest dip of the normally vertical dike may be due to some deflection of the dike by the thrust zone. The dike is cut by small faults that have northerly and northwesterly trends. Fractures in the dike near Higganum, Connecticut, may have been produced by a subhorizontal axis of compression oriented approximately north-northwest--south-southeast (Sawyer and Carroll, 1982).

Abundant small northeast-northwest-and north-trending faults that may be Mesozoic in age cut rocks of the Moodus area (Fig. 58). The northwest-trending faults appear to be the youngest. These faults are exposed at places in roadcuts and other fresh excavations in bedrock and have steeply dipping fractures and shear zones with slickensides, breccia and gouge; these zones are easily eroded and are rarely exposed in natural outcrop. Many of the topographic and LANDSAT lineaments in the region have the same trends as these small faults and may be controlled by them.

The northwest-trending Connecticut River forms the most prominent lineament in the area. The Selden Neck fault, which has an apparent 0.7 km left-lateral offset and is well expressed topographically, may extend northwest up the river to near the mouth of the Salmon River (Fig. 58). Projection of mapped contacts from opposite sides of the Connecticut

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River where it flows east in the Middle Haddam quadrangle indicates that a fault with right-lateral offset of about 200 m probably lies beneath the river.

The northeast-trending Salmon River forms another very prominent lineament which appears to be controlled by small faults and joints. It crosses the nose of the Moodus anticline and appears to be locally controlled by small faults in the northeast (R. J. Fahey, written commun., 1978) and southeast corners of the Moodus quadrangle. Joints parallel to the river occur other places along it. A gravity low forms a lineament along the river (Kick, 1982) that does not appear to be explained by lithologic differences at the surface (Fig. 57). The river also has several north-and northwest-trending jogs along its course where northwest-trending topographic lineaments extend across it.

No recent offsets have been identified in the Moodus area, but mapping is still incomplete. Recent displacement of bedrock, however, is known to have taken place along Connecticut Route 11, east of Moodus, where drill hole scars are offset across small Paleozoic thrust faults in roadcuts (Block and others, 1979). The movement is up, north side over the south. The greatest offset occurs in the exposures that are isolated by cuts on both sides between the roadways; little is seen in the outer cuts. "Noises" were heard from these highway cuts soon after they were opened that probably resulted from these small shifts in the rock (J. deBoer, written commun., 1982). This movement is similar to that reported from quarries and other highway construction in the region and is apparently from a release of residual strain due to changes in pressure caused by removal of the rock and indicated the rock is highly strained. It is not possible to determine when this strain was imposed

on these rocks because the direction of recent movement in the road cuts is approximately the same as the apparent direction of earlier movements dating probably as far back as the Precambrian.

Vertical movement has been found, from precise releveling studies, to be occurring across the position of the eastern border fault of the Hartford Graben and Higganum Dike east of New Haven, Connecticut (Brown, 1978) (southwest corner of Fig. 57). The northwest side is relatively down. These structures project into the seismically active area at Moodus and, if the movement is associated with them, it may also be occurring near Moodus.

Seismicity

The area around Moodus, Connecticut (Figs. 54 and 61) is one of the most continuously seismically active places in the northeastern United States. Indian legends from pre-colonial times noted the area for its earthquakes. The name Moodus is derived from the Indian name Morehemoodus meaning "place of noises" (Chapman, 1840). The earthquakes are very shallow and are accompanied by "noises": rumbling and booming sounds (Ebel and others, 1982) created when the high frequency vibrations of the ground couple with the atmosphere. Earthquakes of less than magnitude 1 have been felt and ones as small as magnitude 0 have been heard (Ebel and others, 1982). The area, thus, has a class of earthquakes that are heard but not felt. The largest known earthquake in the Moodus area occurred on May 16, 1791, and was felt across all of southern New England. This earthquake had an intensity of about VII, causing slight damage in the Moodus area (Linehan, 1964).

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Activity had been low in recent times, up to about 3 years ago. Since then earthquake swarms, numbering a few hundred events each, have occurred in September, 1980, September-October, 1981 and June, 1982 (Fig. 62). The historic activity is clustered near Moodus and earthquakes of the much more accurately located recent swarms are mainly located north of the Moodus center. The most recent earthquakes and the better located and larger ones of the swarms occur just south of the bend in the Salmon River (between seismograph stations MD1 and MD2 on Fig. 62) with a few farther southeast of these (G. Lablanc, written commun., 1983).

The earthquakes since 1980 occur in an area apparently crossed by several northwest-trending faults, as deduced by mapped faults projecting into the area from the northwest and topographic, radar and LANDSAT linements within the disturbed area. The bend in the Salmon River is probably controlled by one of these faults. Perhaps northeast-trending fracture zones near the river provide structural intersections that help localize movement. Earthquakes prior to 1980 are poorly located, but tradition places most of them west of Moodus center, just east of the Salmon River.

The active area lies just east of the intersection of two fracture zones that probably extend through the crust: the northeast-trending Higganum dike zone and the northwest-trending fault and geophysical lineament zone astride the Connecticut River and continuing seaward to the southeast. Perhaps movement in the area of the earthquake swarms is due to adjustments, on local brittle faults, from distortion of the rock between these zones or along the larger northwest-trending zone alone. It might not require much north-northwest--south-southeast aligned

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compression, such as apparently affected the Higganum dike or vertical movement, as recorded near New Haven, to produce shallow earthquakes in these highly strained rocks.

Narragansett Bay

(Wareham - Taunton, MA, 1800, I = V)

Summary

The Narragansett Bay region consists of a largely fault bounded Pennsylvanian basin within a late Precambrian batholithic complex that is overlapped just offshore by the submerged northern extension of the Atlantic Coastal Plain. Faults with Mesozoic movement trend northeast, north and northwest, with the latter two sets the youngest. Movement on the northerly trending normal faults that control the southern end of the bay and possible similar nearby ones might cause the low level seismicity in the area. This movement could be in response to lateral movement along a probable fault zone, represented by a northwest-trending magnetic lineament zone, just offshore of the bay entrance.

Geology

The region around Narragansett Bay consists of a late Precambrian batholithic complex, the Southeast New England Platform, into which the largely fault bounded Pennsylvanian Narragansett Basin has been dropped (Barosh and Hermes, 1981) ((Fig. 63). Western Rhode Island is formed of a complex domal structure cut by northeast-trending faults. The batholithic complex east of the basin is mainly covered by glacial material, but aeromagnetic lineaments suggest northeast and

north-trending structures (Barosh and others, 1977). Permian granite has intruded the shore area southwest of the bay.

During the Mesozoic, Late Triassic - Early Jurassic grabens formed offshore to the east (Ballard and Uchupi, 1975) and the area was overlapped by the submerged northward extension of the Atlantic Coastal Plain in the Late Creataceous (Fig. 63). Prior to the Pleistocene glaciation the Coastal Plain extended farther inland (Kaye, 1983). Mesozoic faults trend northeast, northwest and north, with the later two sets being the youngest. These are all high-angle faults and many can be shown to have normal offsets. Northeast trends include the border faults of the grabens and the Watch Hill en echelon fault zone that extends from southwest Rhode Island across Narragansett Bay into the southeast border of the Narragansett Basin (Fig. 64). This fault zone is, at least in part, a reactivation of a late Precambrian one. South of this zone are a series of north to north-northeast-trending faults and geophysical lineaments that cut the bay (McMaster and others, 1980) (Fig. 65) and the areas on either side. In southwestern Rhode Island the Watch Hill fault is offset by north and northwest-trending faults, that may be related to the New Shorham fault offshore to the south (Figs 64 and 65). The New Shorham, which cuts Upper Cretaceous rock, has both these trends (McMaster, 1971). A zone of northwest-trending magnetic lineaments lies offshore of the bay entrance (Fig. 65). These may represent Mesozoic or younger faults as detailed ground magnetic surveys (Frohlich, 1981) show they continue on shore at Point Judith and offset other magnetic trends.

Remnants of Coastal Plain Tertiary sediment are present on Marthas Vinyard to the southeast and the mainland cost to the northeast

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(*Fig. 63*). These deposits have been disturbed by the movement of glacial ice, but some of the high-angle faults that cut them may possibly be tectonic (C. A. Kaye, oral commun., 1983).

The Narragansett Bay is structurally controlled, and has been flooded due to a combination of the post-Pleistocene regional southward tilt of New England and rise in sea level, but whether or not local Holocene subsidence has also occurred is not yet known.

Seismicity

A diffuse grouping of epicenters is centered over the Narragansett Bay and adjacent area to the east (Fig. 54) and several intensity IV's and V's are located in the lower part of the bay. The largest earthquake reported occurred near Wareham and Taunton in 1800 with an intensity VI (Smith, 1962; Nottis, 1983) in the southeast side of the Narragansett Basin; a location near both northeast and north-trending faults at the border. A possible earthquake cause would be normal fault movement along northerly-trending faults, perhaps related to lateral movement along the northwest-trending lineament zone offshore.

Cape Ann

(Cape Ann, MA, 1755, I = VIII)

Summary

Numerous earthquakes have occurred in northeastern Massachusetts and adjacent coastal New Hampshire centered around Cape Ann with the largest ones located offshore north of the Cape. The area lies along a broad embayment of the New England coast and remnants of Cretaceous and Tertiany strata offshore indicate an embayment of the submerged Coastal Plain may have existed here. Subsidence is indicated to have occurred in the early Holocene and at the present time. It is a highly faulted area with a great many northeast-trending Paleozoic faults. A zone of very large faults, bounded by the Clinton-Newbury and Bloody Bluff fault systems passes through the center of the seismic area, as does a major northwest-trending fracture zone, that crosses New Hampshire. Movement across the fracture zone, perhaps localized by the intersection with the zone of faults may be related to the subsidence and have controlled the sites of the large earthquakes. The smaller peripheral earthquakes may be due to local adjustments to the subsidence on various faults.

Geology

Northwestern Massachusetts and coastal New Hampshire straddle the largest structural zone known in New England, the Nashoba Thrust Belt, that marks the collision boundary between the Paleo-African plate, on the southeast, and the North American plate, on the northwest (Fig. 63).

The belt underwent intermittent movement during the Paleozoic, movement that formed local grabens along it at times (Barosh, 1982). It trends east-northeast offshore just north of Cape Ann. The southeast New England Platform, southeast of the thrust belt is a late Precambrian batholithic complex into which the east-trending graben-like Boston Basin was formed in the latest Precambrian and early Paleozoic (Kaye and Zartman, 1980). The lower Paleozoic granite that intrudes the platform and forms most of Cape Ann is cut by numerous northeast-trending faults (Dennen, 1981) (Fig. 66). Metamorphic rock northwest of the thrust belt is cut by north-northeast-trending faults.

A broad northwest-trending zone of some faults and topographic, LANDSAT and geophysical lineaments crosses New Hampshire through Lake Winnipesaukee and enters the sea trending towards the area north of Cape Ann (Barosh, 1976, 1982) (Fig. 67). It must represent a major fracture zone although no great offset may occur along it. The zone crosses the older structure and appears to be mid-Paleozoic or younger in age. This zone is parallel to the oceanic fracture zones at the edge of the continent to the southeast (Fig. 64), but the wide gap across the Gulf of Maine makes it difficult to correlate with a particular fracture zone.

Much of the Mesozoic record is covered by the sea. Northeast-trending Late Triassic - Early Jurassic grabens were formed offshore (Ballard and Uchupi, 1975) and at least one small graben along the Nashoba Thrust Belt (Kaye, 1983) (Fig. 68). Early Jurassic diabase dikes also cut the area and trend both northeast and northwest in coastal New Hampshire (F.X. Bellini, written commun., 1981). Some Jurassic and Cretaceous plutons occur in New Hampshire and adjacent

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Maine and apparently represent the cores of ancient volcanos. A circular magnetic anomaly at sea north of Cape Ann may represent one of these (Boston Edison Co., 1976), but it may also represent older rock. Some northwest-trending post-Cretaceous faults are known to cut these plutons in southern New Hampshire (Freedman, 1950) and may represent the age or reactivation of the northwest-trending fracture zone across the state.

Cape Ann juts out into a broad embayment in the New England coastline, and thin remnants of Cretaceous and Tertiary strata are found offshore (Weed and others, 1974) (Fig. 63). These remnants may have formed an embayment in the Coastal Plain, similar to those farther south, before they were largely removed by glacial scour. The area is indicated to be subsiding in the Holocene after post-glacial rebound elevated the late Pleistocene shoreline. Submerged tree trunks along the New Hampshire coast are thought to indicate tectonic subsidence 2,400 to 3,250 years ago (Harrison and Lyon, 1963). Releveling studies show present day subsidence increasing southwest of Portland, Maine (Taylor and Ladd, 1981) (Fig. 7), and the sea level rise over rock ledges in Salem Harbor on the south side of Cape Ann is on the order of 2.5 feet in 90 years (8.4 mm/year) (D.C. Smith, written commun., 1982). The extensive salt marshes and shoreline erosion north of the Cape would be compatible with submergence.

Seismicity

Most of the seismic activity in eastern Massachusetts is concentrated in a great arc around Cape Ann, a peninsular that juts out into

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the great embayment in the southeast New England coast (Fig. 2, 54, and 69). The active area extends from the south side of Boston to the Maine border. It extends inland to the south edge of the active zone along the Merrimack River in New Hampshire (Figs. 54 and 69), but appears distinct from it. The largest earthquakes occurred offshore of Cape Ann and the earthquakes tend to decrease in size away from these towards the western edge of the zone (Fig. 69). The few determined depths are shallow.

The two principal events are the 1755 earthquake of probable epicentral intensity VIII and the 1727 events of intensity VII (Fig. 69). Intensity VI earthquakes were felt in 1627, 1744, 1817 and two in 1963. Twenty one intensity V's have been recorded and numerous III's and IV's. Many other small earthquakes have probably occurred offshore and have not been recorded. The region was move active prior to 1850 than since (Shakol and Toksoz, 1977), but there has not been any noticeable areal shift in activity.

The earthquake of November 18, 1755, that occurred the same year as the Great Lisbon earthquake, was felt from Halifax, Nova Scotia, to at least Anapolis, Maryland, over approximately 1,000,000 square km (385,000 square miles) Boston Edison Co., 1976). It caused damage across eastern Massachusetts and thoroughly frightened the inhabitants. Both this and the 1727 event have been determined to have offshore epicenters north of Cape Ann and a reevaluation of the effects for the 1755 earthquake has resulted in lowering the estimate of the epicentral intensity to VIII (Boston Edison Co., 1976). The isodeismal data was used to estimate a magnetude, M_b, of 5.8 for the 1755 event (Street and Lacroix, 1978). but this is probably a minimal value.

The center of the seismic activity lies in the vicinity of the intersection of the two largest structural zones in the region and in an area of subsidence; all of which are probably related. The Nashoba Thrust Belt is not in itself an active zone; no seismic zone follows it and it is known to be cut by a younger north-trending fault. The only other activity that occurs near it is at Moodus, Connecticut, the only other area where a major northwest-trending fracture zone cuts it (Fig. 58). Seismic activity occurs along the northwest-trending fracture zone in central New Hampshire (see section on Merrimack River Valley), and it may be that the intersection with the thrust belt has localized another active area. The area of subsidence is too ill defined to know if its center corresponds to the center of activity, as at Passamaquoddy Bay, Maine and New Brunswick, but they could be related. The subsidence may represent a sag, by small movements, across the broad fracture zone or the fracture zone could be locally reactivated by the subsidence from a deeper cause. The smaller peripheral earthquakes may be due to local adjustments on various faults to the subsidence.

Casco Bay - Lower Androscoggin River

(Bridgeton-Norway, ME, 1918, I = VI)

Summary

Western Maine is traversed by a broad northwest-trending structurally disturbed zone, across which the Paleozoic strata is shifted left-laterally. Within it are many Devonian intrusions and the regional northeast strike of the strata is generally twisted in other directions, except near the coast where northeast-trending formations and faults are dominate. Most of the earthquakes appear associated with northwest-trending structural lineaments along the northeast side of the disturbed zone. Others lie near the northeast-trending Lewiston Lineament and perhaps others parallel fault zones in the southern part of the area. Some of these locations lie near where the northeast-trending faults are intersected by northwest-trending topographic lineaments.

Geology

The generally northeast striking metamorphosed Paleozoic strata of Maine is disturbed in across a northwest-trending zone in Western Maine from near Casco Bay to the northwest corner of the state (Fig. 67). The Devonian plutons are well represented in the zone, especially along its western side. The regional strike of the strata is twisted into various strikes at the northeast side of the zone and the formations appear to make a general left-lateral shift of about 20 km across this boundary.

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Many twists and jogs in the geologic contacts near the boundary are aligned and coincident with topographic lineaments. These form northwest-trending structural lineaments that probably represent fault zones.

A major northeast-trending structural zone, the Lewiston Lineament, marked by some mapped faults and topographic, LANDSAT and geophysical lineaments, crosses the area through Lewiston, Maine (Barosh, 1976, 1982; D Carter, written commun., 1977; Lee and others, 1977) (Figs. 67 and 70). Southeast of it the structure has a predominant northeast-trend and includes many high-angle thrust faults similar to those in southeastern New Hampshire (Fig. 71). The disturbed zone is only marked here, near the coast, by some Devonian plutons. A number of the northeast-trending faults can be shown to have post-metamorphic movements which probably is post-Devonian in this area.

Mesozoic dikes and plutons cut the area. A wide variety of basic dikes are exposed along the coast, especially southwest of Portland but few inland. Several Jurassic to Creataceous granite plutons apparently mark an ancient northerly trending volcanic belt near the western edge of the state.

The only post-Cretaceous faults known are a few northeast and northwest-trending ones cutting the late Mesozoic plutons, but some of the numerous northwest-trending LANDSAT lineaments may represent others. Similar lineaments coincide with mapped relatively young faults to the northwest in northwesternmost Maine (Westerman, 1981) (Fig. 72)

Leveling and other studies near the coast indicate some present-day subsidence that increases in amount towards the southwest from Portland (Taylor and Ladd, 1981) (Fig. 7). This may be more related to center of

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subsidence farther southwest and have little to do with area to the north.

Seismicity

Scattered epicenters occur in southern Maine where a slight clustering of the larger earthquakes near the lower Androscoggin River shows a northwest elongation (Figs. 54 and 73). These epicenters are separated from the activity in central New Hampshire to the west and that in the Penobscot region to the northeast by narrow areas of very low seismic activity.

Some general spatial correlation of the epicenters with structure has emerged. Many epicenters lie near the Lewiston Lineament between the Kennebec River and Sebago Lake (Figs. 70 and 73). Nearer the coast there is a small cluster near Portland and one northeast of Casco Bay near the mouth of the Androscoggin River. One of the better located earthquakes and after-shock sequences occurred in the latter group in 1979 near the apparent intersection of a northeast-trending fault and a north-trending one (Ebel, in press) (Fig. 74). However, a prominent northwest-trending topographic lineament zone also crosses the area and the epicenters also are aligned along its northeast side.

The epicenters north of the Lewiston Lineament show no clear pattern in themselves, but most of them lie along the northwest-trending structural lineaments along the northeast side of the disturbed zone. Several earthquakes are clustered near Lewiston where one of these lineaments intersects the Lewiston Lineament (Fig. 70) and a few others occur on it farther northwest. A recent well located earthquake near

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Dixfield (Ebel, in press) (Fig. 75) lies on another parallel lineament to the northwest. Another lineament passes near the 1949 intensity VI earthquake farther north, and if projected to the northwest, goes through the epicentral area of the widely felt 1973 Maine-Quebec earthquake. Few earthquakes occur in the area of extensive Devonian granite along the western side of the state.

The earthquakes, thus, appear mainly associated with northwest-trending structural lineaments that lie along the northeast side of a major structurally disturbed zone. Some also may be associated with northeast-trending faults crossing this zone, but these too may be at intersections with northwest-trending fractures.

Penobscot Bay

(Milo, ME, 1928, I = VI)

Summary

Earthquake epicenters form a slightly northwest elongated cluster north of Penobscot Bay across a flexure in the regional geology. The earthquakes within the cluster appear spatially associated with lineament zones, known, at least in part, to represent fault and joint zones. These zones are the north-trending Penobscot Lineament, a fracture zone extending northwest from Belfast and a section of the northeast-trending Lewiston Lineament and a more easterly trending branch from it.

Geology

The area north of Penobscot Bay on the central Main coast consists mainly of northeast striking metamorphosed Paleozoic formations similar to those in the southern Casco Bay-Lower Androscoggin River area (Hussey and others, 1967). The principle structure also parallels the strike of the formations and includes the major Norumbega fault in the south and the Lewiston Lineament near the north edge of the area (Fig. 76). A zone of faults with a more easterly trend branches off the north side of the Lewiston Lineament near Milo and passes west-southwest through Dover-Foxcroft. The eastern side of the area along the Penobscot River is bounded by the Penobscot Lineament zone, that consists of a zone of generally northerly trending lineaments, that crosses most of Maine

(Barosh, 1981). Within the area of the Bay the lineament zone consists of northeast-trending fault zones (Stewart, 1974; Barosh, 1981; Rogers, 1983) (Fig. 77). The area east of the lineament is extensively invaded by Devonian plutons (Hussey and others, 1967) and the strike trends are not as uniform or prominent.

A series of northwest-trending lineaments extends northwest from Belfast and passes through the position where the regional strike changes from north 40° east on the southwest to north 65° east on the northeast (Fig. 76). The structure resembles a fractured hinge zone across a flexure. Many of those northwest-trending lineaments and others in the area have been shown to be small fault and joint zones (Westerman, 1981; Rogers, written commun., 1980) (Fig. 78).

Most of the faults are apparently Paleozoic in age, but Mesozoic dikes along the Lewiston Lineament and fault slivers of late Paleozoic strata along the Norumbega fault zone suggest possible Mesozoic Movement. The brittle features along other faults also suggest shallow and therefore, probable late movement. The northeast-trending faults along and northeast of Penobscot Bay have brittle features (Barosh, 1981; Rogers, personal commun., 1980); at Sears Island Pleistocene material is offset slightly across one of these (Rand and Gerber, 1976) (Fig. 79). Arguments continue as to whether or not this represents Holocene tectonic movement of some syn- or post-glacial adjustments to changing ice load. Only very small present day crustal movements have been found in the area (Taylor and Ladd, 1981), but a great deal of movement has occurred in the past and the rock is highly strained (Lee and others, 1977, 1979).
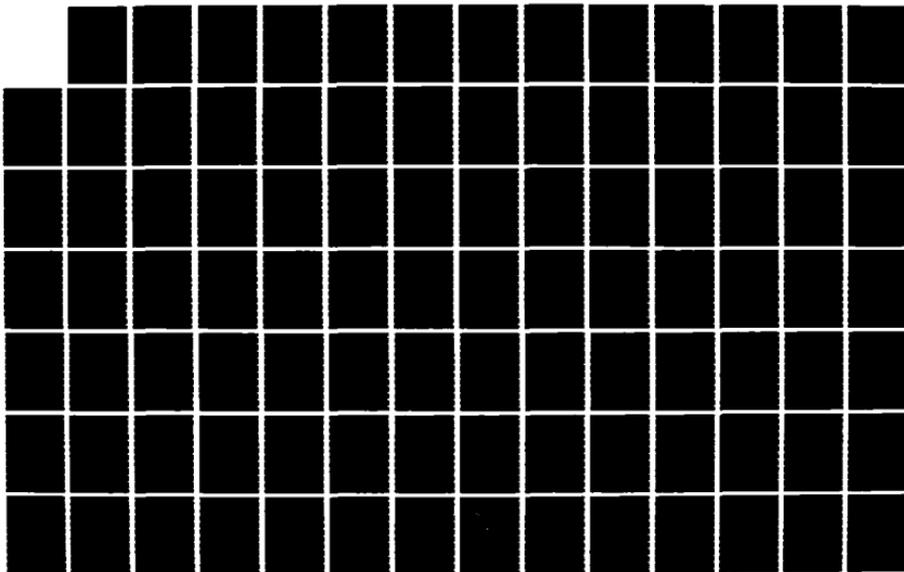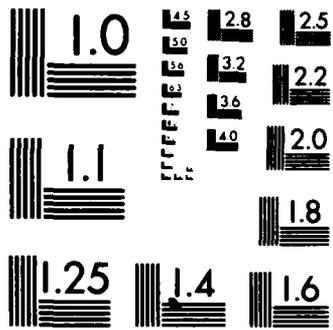
MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS-1963-A

Seismicity

The earthquake epicenters north of Penobscot Bay form a slightly northwest elongated cluster (Figs. 54 and 76). Most of the small to moderate sized earthquakes occur in the triangular shaped area of a flexure bounded by the Penobscot Lineament on the east, the northwest-trending fracture zone on the southwest and the Lewiston Lineament and its branch on the north (Fig. 76). The latest earthquake catalog (Nottis, 1983) places most of the earthquakes near the bordering structural zones. A few other earthquakes have taken places further north and have a slight northwest-trend in their distribution. These lie near Chesuncook and Caribou lakes where a few northwest-trending faults have been mapped and LANDSAT lineaments of this trend abound. The largest earthquake known in this area took place in the vicinity of Milo near the junction of the Lewiston Lineament and its easterly trending branch. The most active place in the area over the past several years has been near Dover-Foxcroft, which lies farther west on this branch.

Passamaquoddy Bay

(Eastport, ME, 1904, I = VII)

Summary

The earthquakes near Passamaquoddy Bay, Maine and New Brunswick occur in an actively subsiding area that was the site of significant Paleozoic subsidence and graben development. The structure in the bay forms a complex northwest-trending zone across the regional northeast-trending strike, and many of the earthquakes lie near the main northwest-trending fault zone, the Oak Bay fault. A few other earthquakes occur near probable adjacent parallel structures.

The activity appears caused by adjustments on structures due to subsidence by extensional movement along a northwest-trending structural zone extending inland from the Bay of Fundy.

Geology

Passamaquoddy Bay, straddling the Maine and New Brunswick border, lies in an area that consists mainly of lower to mid Paleozoic formations and intrusive rock. It was an area of subsidence during the Silurian (Gates, 1981) and the location of Late Devonian grabens (Schluger, 1973) (Fig. 80). The regional northeast structural trend swings to the northwest in the bay area forming a transverse zone of very complex structure (Bastin and Williams, 1914; Gates, 1975; Gates and Ludman, 1983). This zone shows up strongly as lineaments on the magnetic and gravity data (Hildreth, 1979; Hodge and others, 1980;

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Weng, 1981). Many northwest-trending faults amd LANDSAT Lineaments are present including the large high-angle Oak Bay fault near the southwest side of the Bay (Gates, 1981) (Fig. 81). A large gravity high projects into the bay along this fault zone from offshore (Fig. 82).

The large northeast-trending Late Triassic - Early Jurassic Bay of Fundy Graben lies offshore and the Fundy fault zone along its northwest border lies just off the coast (Bell, written commun., 1972) (Fig. 80).

The area around the bay is shown to be subsiding at a very rapid rate, about 9 mm/year relative to Bangor, Maine, by a series of independent studies using different data bases (Taylor and Ladd, 1981; Borns, 1981) (Fig.7). The center of subsidence lies in the bay near the Oak Bay fault zone and Devonian grabens.

Seismicity

The seismicity in the Passamaquoddy Bay area is concentrated along the southwest side of the bay and extends from the eastern end of Grand Manan Island to the head of the bay (Figs. 54, 80 and 83). Much of the activity lies close to the Oak Bay fault zone and near the center of subsidence (Figs. 80 and 83). A couple of historic earthquakes and a few recent ones (J. Ebel, personal commun., 1984) occur a little to the west near a lineament extending northwest from Machias (Fig. 81). A few others lie near the coast close to the Fundy fault zone (Figs. 80 and 83). This peripheral activity also lies within the area of subsidence.

These spatial relations of the epicenters suggest the earthquakes are due to the subsidence which has triggered local movement along the

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Oak Bay end perhaps associated faults (Barosh, 1981). The gravity high, that extends into the area from the Bay of Fundy, further suggests that perhaps the crust is thinner here, as it apparently is beneath the offshore Bay of Fundy Graben, and the subsidence is due to extensional movement. The concentration of activity along the northwest-trending structure versus the minor amount along the northeast-trending graben seems to indicate the present stress regime favors reactivation of the former trend.

UPLAND ZONES

Seismic activity occurs in places in highlands and mountains, which, where evidence exists, are shown to be rising at present (Fig. 14). These are the southern and central Appalachian Mountains, the Adirondack Mountains and the highlands of Central New Brunswick that, although outside the United States, extend to the Maine border. The southern and central Appalachian Mountains were actively rising and eroding during the Late Cretaceous and Early Tertiary and the Adirondack Mountains may have been. The record for the New Brunswick area is not clear.

The relative lack of seismicity in the Berkshire, Green and White Mountains is not clear. Perhaps it is due to a combination of their north-trends and being away from the zones of movement along the Atlantic Coast and St. Lawrence River.

Southern and Central Appalachian Highlands

(Giles County, VA., 1897, I = VIII, m_b = 5.8)

Summary

The southern and central Appalachian highland, which is controlled by northeast-trending Precambrian and Paleozoic structures, coincides with a major zone of seismicity. The highlands were rising during the Cretaceous and Tertiary and are indicated to be doing so at present. The seismicity appears to be due to this general vertical movement. The activity is not uniform, but has local concentrations, some of which display a spatial relation with cross structures. This suggests that the location of much of the activity within the zone is controlled by intersections with cross structures, especially, northwest trending ones. The larger earthquakes appear to occur at spaced structural intersections and not everywhere within the zone.

Geology

The highlands forming the Blue Ridge and Valley and Ridge regions of the Appalachian Mountains continue northeastward across northern New Jersey and the Hudson Highlands of southeastern New York into southwest Connecticut where they meet the northward trending Berkshire and Green Mountains of western New England (Fig. 84). These physiographic features essentially correspond to Precambrian-Paleozoic geologic provinces that extend southwestward from southwest Connecticut to their

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buried terminus in Alabama (King, 1969) (Fig. 84). The southwest portion of this region is referred to as the Southern Appalachian tectonic province by Waldron (1969). This older structure strikes predominantly northeastward. The Blue Ridge and northern New Jersey-Hudson Highlands is characterized by a narrow rise of Precambrian rock and the Valley and Ridge, to the northwest, by extensive shallowly east-dipping thrust faults, that moved northwestward. Large Late Triassic-Early Jurassic grabens are present at the north end. A good summary of the structure of this region is still provided by King (1951). The southeastern boundary of this region is the Brevard zone, a major Precambrian-Cambrian fault zone (King, 1951) and its northeasternward projection along the Early Mesozoic fault bounding the Culpepper Basin, a graben, and the Northern Fall Line (Barosh and Pease, 1976, 1981) (Figs. 41 and 85). Considerable evidence indicates right-lateral movement along the Fall Line during the Paleozoic (Barosh and Pease, 1976, 1981; Thompson, 1978; Tillman, 1983), left-lateral movement, here and in the adjacent rock during the Early Mesozoic, (Thompson 1983; Tillman, 1983; and Manspizir, 1981) and probably also in the Late Mesozoic (Brown and others, 1972) and followed by post-Mesozoic reactivation (Thompson, 1983). Such movements have probably occurred along the zone to the southwest as well. At least the Northern Fall Line part of the zone also acted as a hinge line, with the rock on the northwest rising relative to that on the southeast.

A great number of transverse structures have been found independently by many workers in the central and southern Appalachian Highlands. These trend predominantly northwest, but north and west structures also occur. They are commonly subtle, showing up as

topographic, radar and LANDSAT lineaments, aligned structures, changes in structural trend, changes in stratigraphic thickness or facies, and gravity and magnetic lineaments (Fig. 86). Field investigations reveal small to moderate fault zones, zones of small faults or fracture or joint zones, that may be quite wide. Many of the cross structures are considered to represent reactivation of larger and older structures at depth, perhaps similar to the surface fracture zones that propagated upward through alluvium 300 to 600 m thick from basement structures by general ground shaking at the Nevada Test Site (Barosh, 1968). Hobbs (1904) may have been the first to recognize the regional set of northwest-trending zones (Fig. 87). Many individual zones have been recognized and studied (a bibliography of such studies was compiled by Wheeler and other, 1981 and many others exist.) Since the discovery of the oceanic fracture zones there has been considerable speculation as to their relations. The most closely related zone is one recognized by several investigations that crosses near the southwest end of the highlands and connects with the Bahama oceanic fracture zone (Gableman, 1979) (Fig. 40). The ages given for the zones range from Precambrian to Holocene. The observed surface features cut the Paleozoic structure and are younger. The northwest-trending faults are known to be younger than the northeast-trending Mesozoic structure in eastern Pennsylvania, northern Delaware, New Jersey and southwest Connecticut (Thompson and Hager, 1979; Thompson, 1983; Tillman, 1983).

The Southern and Central Appalachians Highlands, along with the western position of the Piedmont to the east, rose during the Cretaceous and early Tertiary to form highlands that eroded to provide the sediment for the great deposits of this age in the eastern Gulf and Atlantic

Coastal Plains (Fig. 12). Various geomorphic, geodetic and mineralogic studies suggest they are still rising (Campbell, 1933; Fenneman, 1938; Meade, 1971; Gilluly 1972; and Zimmerman, 1980). The bend of the Coastal Plain boundary around the southwest end of the Appalachian Highlands and the southwestward shift of the Tertiary-Cretaceous contact across the axis of this bend (King and Beikman, 1974) reflects some of this movement.

Seismicity

A prominent northeast striking belt of seismic activity follows the Southern and Central Appalachian Highlands on both compilations of historic data (Hadley and Devine, 1974; Coffman and others, 1982; Tarr and Rhea, 1983; Seay and Hopkins, 1981; Nottis, 1983; Stover and others, 1979 b,c,d; Reagor and others 1980 a,b,c) and in recent seismic network monitoring (Bollinger and Mathena, 1982,) (Figs. 2, 41, 42, 47 and 88). The earthquake pattern along the belt is irregular in detail and slightly less activity appears to occur in the northeastern portion. Small clusters of events appear within the zone; some places have had several events and these are where the larger earthquakes have occurred (Fig. 42). Intensity VII earthquakes have occurred near Birmingham Alabama, Knoxville, Tennesee, Wilmington, Delaware, and possibly western North Carolina and an VIII in Giles County, Virginia (Coffman and others, 1982). These areas are about equal distance apart. The Giles County earthquake of May 31, 1897 (near Parisburg, Virginia) was the largest in Virginia and the second largest in the southeast United States (Bollinger and Wheeler, 1983). It had an epicentral intensity of

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VIII and an estimated magnitude m_b of 5.8 (Nuttli and others, 1979; Bollinger and Hopper, 1971) (Figs. 2 and 88). Many earthquakes have occurred near Knoxville (Fig. 127); fewer are known from the Birmingham area, but the improved seismic network is now recording more from there (Bollinger and Mathena, 1982), and moderate-sized earthquakes have occurred in the vicinity of Wilmington.

There are suggestions of multiple controls of the seismicity in the region, but this may be due to different ways the same kind of movement operates on local structures from place to place and not basic differences. The seismicity in the southern and central Appalachian highlands occurs in an area apparently undergoing uplift and is probably related to this movement (Bollinger, 1973; Barosh, 1981). The northeast trend of the uplift is related to the development of the Atlantic Coast and is undoubtedly strongly influenced by the pre-Mesozoic structural grain. However, tectonic provinces drawn on this earlier structure are inappropriate for matching with the seismicity as they do not reflect the more recent tectonic events (Seay and Hopkins, 1981).

The Brevard zone and the Northern Fall Line coincide with the regional change in pattern of seismicity and appear to form an important boundary of this upland seismic zone (Barosh, 1981) (Fig. 41). During the Cretaceous and Tertiary the Northern Fall Line acted as a hinge line, the northwest side rising, and this combined boundary may still be acting as such. Activity occurs locally along the Fall Line and the Brevard zone, but not everywhere. There is a suggestion that these zones are active where they are crossed by northwest-trending zones of movement.

The northwestern side of the seismic zone from northeastern Alabama to southeastern Virgina is well defined by a series of northeast and

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northwest-trending lineaments, apparently representing structures, that stand out prominantly on both the magnetic and gravity maps of Keller and others (1980) and Johnson and others (1980) (Seay and Hopkins, 1981). These structures are not readily apparent in the geologic maps and are thus in the Precambrian basement, although, small structures and LANDSAT lineaments may be expected at the surface over them. A strong prominant gravity and magnetic lineament, here referred to as the Chattanooga-Knoxville Lineament, extends through these cities and northeastward along the Kentucky-Virginia border (Fig. 86). This lineament separates northeast-trending anomalies reflecting the Appalachian structure on the southeast from irregular north-northeast-trending anomalies of the continental interior on the northwest and apparently represents a major fault zone. It trends slightly oblique to the thrust faults marking the northwest side of the Valley and Ridge province. The zone lies within the province near Chattanooga and extends ⌐tside of it along the southeastern edge of Kentucky. The area between this structure in eastern Tennessee and the Brⁿvard zone to the east is very active. It passes along the northwest side of Birmingham, Alabama, to the southeast and continues beneath the Cretaceous as the buried Appalachian front.

A series of northwest-trending aeromagnetic lineaments, spaced approximately 80 km (50 miles) apart transect eastern Tennessee and Kentucky. The approximate position of three of these have been previously noted by (Seay and Hopkins, 1981) and the LANDSAT lineaments shown by them follow others (Fig. 86). Shifts in anomalies across a few of these lineaments suggests they are faults with left-lateral offsets on the order of 10 km (7 miles). These lineaments show a spatial

relation with the distribution of seismicity and activity occurs along some of them. The concentration of activity at both Knoxville and Chattanooga lie at the intersection of lineaments of this set with the Chattanoga-Knoxville Lineament. A smaller cluster of earthquakes between these two cities lies at the intersection of a third zone. The pattern of seismicity shifts from one northeast-trend to another near the position of some of the northwest trending zones. The activity shifts southeastward northeast of Knoxville to another northeast-trending geophysical structure in northeasternmost Tennessee and western Virginia. There a cluster of minor activity lies near Bristol, Tennessee, at the intersection of another northwest-trending zone (Figs. 86 and 88). The pattern of activity shifts again to the southeast across another northwest-trending lineament that approximately follows the West Virginia-Kentucky border. Northeast of this lineament is the northeast-trending activity of Giles County, Virginia (Bollinger, 1981) (Fig. 89). The pattern of seismicity farther northeast in Virginia suggests other shifts as well. The activity, thus, appears to shift in a step-like pattern along active segments of different northeast-trending structures perhaps activated by different northwest-trending cross faults. This pattern appears too consistant to be fortuadous. The thermal springs of the southern Appalachian Mountains occur in this region (Berry and others, 1980) and their distribution could follow a similar control. No apparent direct structure connects the seismic activity of Giles County, Virginia, with that to the southeast at Knoxville and Chattanooga, Tennessee (Seay and Hopkins, 1981). However, taking into consideration this step-like pattern, the activity from northeast Alabama to southwest Virginia could

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be considered connected, but offset. It is doubtful, however, that an earthquake like the Giles County event (I = VIII) could occur anywhere along this zone, as the northwest-trending cross structures appear important in concentrating strain. These northwest-trending seismic trends and the numerous northwest trending LANDSAT lineaments (Seay and Hopkins, 1981; Titcomb and Hancock, 1981) (Fig. 86) suggest some control by structures of this cross trend, not necessarily everywhere along them but perhaps where they meet susceptible structures.

The activity over the northwestern part of the thrust faults in the Valley and Ridge province probably originates on structures beneath the thrust faults, as they do along the St. Lawrence near La Malbaie, Quebec, (Leblanc and Buchbinder, 1977). They might be expected to show up on LANDSAT images. The continuity of LANDSAT lineament patterns across and on either side of the Southern Appalachian Mountain Zone (Seay and Hopkins, 1981) is therefore suggestive of a continuity of the underlying structures.

The Giles County earthquake occurs on a northwest-trending lineament zone, but has been related to an east-west zone in the geophysical data (Seay and Hopkins, 1981) and a north-northeast and north oriented structure from the distribution of seismicity (Bollinger, 1981). There appears to be some agreement that it is located on some type of structural intersection but the local nature of this intersection is not yet fully understood.

The general northeast elongated grouping of epicenters ends at the Potomac River, where there is a northwest-trending alignment of epicenters. Farther to the northeast the epicenters appear more aligned with northwest-trending zones extending from near Baltimore, Maryland,

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and northern Delaware and near Philadelphia (Thompson, 1980, 1983). Northwest-trending faults lie along the zone from northern Delaware (Fig. 90). The zone near Philadelphia corresponds in part with a mineralized zone (Fig. 91) and appears, on the gravity data, to extend to the Attica, New York area (see section on Attica). Fractures extending from this zone cut the Cretacious to the southeast (Thompson, 1981). The zones extending from northern Maryland and Delaware on the southwest and Philadelphia on the northeast are aligned across bends in the Appalachian structure; the segment between the zones trends more easterly than those on either side (Fig. 53). Farther northeast from northern New Jersey to southwest Connecticut a weak northeast-trend to the seismicity reappears.

A considerable number of earthquakes, including one Intensity VI, have occurred at the northeast end of this seismic zone between the northwest border of the New Jersey and Hudson Highlands and the Northern Fall Line-Long Island Sound area (Nottis, 1983; Aggarwal and Sykes, 1978) (Figs. 54 and 92). The smaller earthquakes are relatively at least slightly over represented due to a concentration of seismographs (Fischer, 1981; A. Kafka, oral commun., 1982). The earthquakes are scattered widely across this region. A great number of faults, of mainly northeast strike and many with proven Late Triassic or younger movement, also cross this area from tunnel, field and geophysical studies (Barosh and others, 1977; Barosh and Pease, 1976, 1980; Geraghty and Isachsen, 1978; Tillman, 1983; Thompson, 1983; Ratcliff, 1971; Fischer, 1981; Manspizer, 1981; Barosh, in press) (Figs. 92 and 93). All earthquakes therefore are located near some fault zone, but there does not appear to be any with a clear concentration along it.

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Earthquakes have been considered to be associated with the Ramapo fault, a post-Late Triassic fault along the northwest side of the Newark Basin, (Page and others, 1968; Ratcliff, 1980), but a study of the recent data did not find this to be the case and that no earthquake greater than intensity V has occurred in the vicinity of the fault (Fischer, 1981). More seismicity, including an intensity VI earthquake, has occurred to the southeast near the Northern Fall Line that has evidence of post-Cretaceous movement. The Hudson River, one of the most prominent features in the area, follows a zigzag course, that is structurally controlled by northeast, north and northwest-trending structures (Isachsen, 1981). The northeast end of the Newark basin, ends at a north-striking stretch of the river that may be fault controlled. The seismicity is, thus, scattered over an area with many northeast-trending faults, and could be associated with some, but perhaps at intersections with the later northwest or north-trending ones.

The seismic activity in the Southern and Central Appalachian Highlands is probably not totally unrelated to that in the embayments to the southeast. The activity appears more concentrated where zones of probable movement in the ocean and Atlantic Coastal Plain project across it. The activity from northern New Jersey to southwest Connecticut is distributed around the head of the Raritan Embayment and the concentrations northwest of Delaware and along the Potomac River are opposite lesser northwest trends (Figs. 42 and 54). There is also some suggestion of this in Virginia where a cross trend extends from Charlottesville northwest across the highlands into northern West Virginia (Fig. 88). A weak northwest trend through the Knoxville, Tennesee region is aligned with one along the northern part of the

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South Carolina-Georgia border and the general higher level of activity in South Carolina is matched by the greater activity to the northwest in eastern Tennessee.

Adirondack Mountains

(Massena, N.Y., 1944, I = VIII, M_1 = 5.1)

Summary

Earthquakes occur within and at the edges of the Adirondack Mountains, which form a structural dome undergoing present day relative uplift. The dome lies on the south side of the location where the graben system that extends southwestward into the continent from the Gulf of St. Lawrence splits into three lowland zones: the west-northwest trending Ottawa River Valley, the southwest-trending upper St. Lawrence lowland and the south-trending Lake Champlain-Hudson River Lowland. The Frontenac Arch rises between the former two and the Adirondack dome between the latter two. These uplifted areas may have resulted from movement along the lowlands.

Northeast-trending faults dominate within the dome, but significant numbers of west-northwest-trending faults and lineaments cross the northern part. Earthquakes occur in weak northwest-trending alignments. The intensity VIII 1944 earthquake near Massena is located at the edge of the mountains within the lowland structural branch along the upper St. Lawrence River where it intersects the Ottawa Valley fault system.

Geology

The Adirondack Mountains are formed of a dome of Precambrian rock surrounded by lower Paleozoic strata (Fig. 9). This constitutes an upland area between the north-trending Lake-Champlain-Hudson River

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lowland on the east and the St. Lawrence Lowland on the northwest (Fig. 94). (The former lowland is discussed separately and the latter is included here). The lowlands are controlled by paralleling graben and extensional fault systems and the upland is cut by a great number of predominantly northeast-trending faults (Isachsen and others, in press and McKendree, 1977; Isachsen and others, in press) (Fig. 95). The area has a long history of deformation (McClellan and Isachsen, 1979) and is cut by many Precambrian faults, including some major ductile thrust faults (Geraghty and Isachsen, 1981). Several of the high-angle faults offset the bordering Paleozoic rock (Fisher and others, 1970) and probably have Mesozoic movement (Cady, 1969). A few northwest-trending faults (Isachsen and McKendree, 1977) (Fig. 95) and several lineaments are present. No direct evidence for Mesozoic or younger faults has been found, but some Mesozoic dikes are present (Isachsen, 1983).

The dome may be a relatively young feature. No evidence of an early upland is indicated in the surrounding lower Paleozoic sediments and the dome is indicated to be relatively rising at a rate of about 4 mm/year at present (Isachsen, 1977).

In a more regional view the Adirondack Mountains are an outlayer of the Canadian Shield; connected to it by the west-northwest-trending Frontenac Axis or Arch (Fig. 96). This arch is nearly separated from the Adirondack Dome by a northeast-trending sag along the upper St. Lawrence Lowland and its continuation through Lakes Ontario and Erie, in the central lowland, to the Mississippi Embayment. Another sag containing Paleozoic strata extends from the St. Lawrence Lowland along the Ottawa Valley parallel to the northeast side of the Frontenac Arch. The structure in the arch, as in the dome, trends predominantly

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northeastward as shown by the pattern of faults, magnetics and gravity (Forsyth, 1981; McGrath and others, 1978; Hildreth, 1979) (Fig. 97). This structure appears to continue northeastward beneath the Ottawa sag and into the shield (Forsyth, 1981). The arch and sag, thus, are transverse structures and relatively young. However, they do have Precambrian precursers as late Precambrian mafic dikes follow these sags (Lumbers and Ayres, 1978). The Paleozoic rock in the sag along the Ottawa Valley is cut by extensional faults and grabens (Fig. 97). This fault zone may be much wider than shown and extend farther northeast into the shield as suggested by the numerous north-northwest-trending topographic lineaments there.

The north side of the Adirondack dome is thus a triple junction of extensional graben-forming fault systems: the Lake Champlain-Hudson River, the St Lawrence Lowland and the Ottawa Lowland (Fig. 97). These graben systems may represent a splitting up (triple point junction) of a major graben system that extends into the Continent along the St. Lawrence Gulf and River (Kumarapeli and Saull, 1966) and probably have Mesozoic movement along them (Cady, 1969). The dome and arch may have risen due to relative movement along them.

The faults related to the Ottawa Valley system extend east-southeast into the northern Adirondack Dome (Fig. 97) and the gravity pattern of the northern half of the dome is disrupted by strong transverse lineaments of this trend. Some lineaments appear to be a continuation of the fracture and lineament zone that passes through Lake Winnipesaukee, New Hampshire, and the more active part of the Merrimack River Valley seismic zone (see section on Merrimack Valley). The southern edge of this zone intersects structure along the upper St. Lawrence near Massena, New York. The Paleozoic strata is thicker where

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this zone crosses western Vermont and it may represent a sag that began in the Precambrian (Cady, 1969).

The upper St. Lawrence Lowland at the western edge of the dome overlies a prominent northeast trending gravity high (Frohlich and others, 1979). However, the river itself crosses this structure slightly obliquely and does not appear controlled here by a major fault zone (Revetta and others, 1975).

Seismicity

A scattering of individual and clusters of epicenters occurs within and around the rising Adirondack Mountains without a clear cut pattern (Figs. 54, 97 and 98). Several places have experienced earthquake swarms in recent years, such as the shallow ones around Blue Mountain Lake (Sbar and others, 1972; Aggarwal and others 1975; Aggarwal, 1978). The largest earthquake known in the area was an intensity VIII event that took place near Massena, New York, in 1944 within the St. Lawrence Lowland. A nearly aseismic area lies to the south of the mountains in the Catskill Mountains where the Paleozoic strata is little deformed (Figs. 2 and 54).

Some variation in the epicentral pattern suggests a series of northwest-trending zones that are more active. One of these extends along the Mohawk River at the southern edge of the area. Another trends northwest from the southern end of Lake George where the Warrenburg earthquake occurred. Yet another lies along the northwest projection of the fracture-lineament zone that passes through Lake Winnipesaukee, New

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Hampshire. The latter extends to the location of the Massena earthquake, which also lies in a northeast-trending structural zone along the upper St. Lawrence (Fig 97). This event may be controlled by a structural intersection.

Investigations of the areas of earthquake swarms did not reveal any obvious correlation of the local epicentral distribution and fault plane solutions to structure (Isachsen and Geraghty, 1979).

The earthquakes are probably due to the rising of the dome with adjustments to the changing stain occurring along fracture zones within the dome and in the bordering lowlands. Northwest-trending structural zones, perhaps in conjunction with northeast-trending ones, may be important in localizing individual earthquakes.

INTERIOR ZONES

The interior zones are those lying inland away from the Coastal and Upland zones. They occur in two broad regions on either side of the Coastal Mississippi Embayment (Figs. 14 and 9). Similar patterns of structure that coincide with seismicity have been revealed by geophysical and remote sensing studies at several places across these regions. The pattern represents basement structures, that are buried beneath the Paleozoic cover of the mid continent, but emerge in New England. The pattern is one of prominent northwest-trending features associated with north-trending ones such as in northeastern Oaklahoma and southeastern Kansas (discussed under East Texas Embayment), and much of Missouri, central and eastern Tennessee and Kentucky, Ohio, western New York and New England. North of central Kansas and Missouri, the northwest-trends appear to be associated with northeast ones. As the geophysical data over the region has improved, these patterns have gradually become clearer and more prominent.

Northern Alabama-West-Central Kentucky

(Alabama-Tennessee, 1959, I = VI)

Summary

A difuse northerly trending seismic zone extends from Birmingham, Alabama, in the Southern Appalachian Highlands to west-central Kentucky, on the far edge of the activity associated with the New Madrid area. A slight extension of this zone projects like a spur southeastward from northeast of Nashville, Tennessee. This zone trends across the western Nashville Dome, a Paleozoic uplift, and the Rough Creek Graben, an east-trending late Precambrian graben. It does not appear to correspond to structures in the mapped geology, although northwest-trending faults are present that parallel the spur of activity. The earthquakes in northern Alabama and Tennessee lie along north and northwest-trending magnetic and gravity lineaments that, where data are available, also correspond to LANDSAT Lineaments. These probably represent high-angle fault zones. The activity in Kentucky cannot as yet be correlated with any particular structure. The greatest activity is at both ends of the zone where it intersects other more active areas.

Geology

Western Tennessee and Kentucky, east of the Mississippi Valley, is underlain by the western side of the Nashville Dome and the Rough Creek Graben (Fig. 33). The Nashville Dome is a northeast trending Paleozoic

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uplift extending from northern Alabama to south-central Kentucky between the Appalachian Mountains and the Mississippi Embayment. It is covered by lower Paleozoic rock and surrounded by upper Paleozoic rock. The Rough Creek Graben of Cambrian and possibly late Precambrian age trends eastward north of it in west-central Kentucky and joins the northeast end of the Reelfoot Rift in western Kentucky. Numerous east-northeast to east-trending normal faults cut the Upper Paleozoic rock over the Rough Creek Graben (Figs. 33 and 34). No faults younger than the Paleozoic are known and in northern Alabama and adjacent Tennessee, none have thought to have moved since early Cretaceous (Tennessee Valley Authority, 1977). A few northwest-trending faults cut the south end of the dome in Alabama (Kidd, 1980) and adjacent Tennessee (Buschbuch, 1980). There are also a couple of north-northwest-trending matching LANDSAT and gravity lineaments that are aligned with a fold axis (Kidd, 1980).

The Precambrian surface (Buschbach, 1980) shows the northeast-trending axis of the dome and the Rough Creek graben well; a slight irregularity trends north-northwest across the end of the dome near the Alabama-Tennessee border. The thickness of the pre-Late Cambrian strata only indicates the presence of the dome and graben at that time.

Northwest, north and northeast-trending lineaments show up well in the LANDSAT imagery of the region (Kidd, 1980; Seay and Hopkins, 1981) (Fig. 99). The bouguer gravity data (Keller and others, 1980; Clements and Kidd, 1975) also show these trends. A wide zone of North-trending lineaments cross along longitude 87 degrees. These are cut in places by northwest-trending lineaments and a few east-northeast-trending ones.

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The northwest-trending lineament that passes through Nashville, Tennessee, extends to Chattanooga, Tennessee.

Seismicity

A difuse zone of seismicity extends northerly from east of Birmingham, Alabama, into central Tennessee, and continues across west-central Kentucky with a northwesterly trend to near Evansville Indiana at the edge of the activity centered around New Madrid (Stover and others, 1979 a,b,c,d) (Figs. 42 and 100). A few events form a spur trending southeast from north of Nashville. Four intensity VI and VII earthquakes in Alabama occur along it, as are several small events of intensity V in Tennessee and Kentucky. An intensity VII occurred near Evansville on the Kentucky-Tennessee border. This poorly defined zone stretches between the Southern Appalachian seismic zone in the south and with the Wabash Valley zone, an extension of the New Madrid zone, to the north; the two largest earthquakes are at either end.

The area of this seismic zone is very anomalous on isoseismal maps. Relatively greater intensity values occur along it from earthquakes on the zone in Alabama and lower intensity values from larger earthquakes generated outside of it from either the Charleston or New Madrid areas.

Most of the activity in Northern Alabama and Tennessee lies near a north-trending geophysical lineament that is intersected by northwest-trending ones. The Alabama-Tennessee earthquake of 1959 lies on one of the northwest-trending zones and those earthquakes southeast of Nashville near another. The earthquakes in Kentucky occur in the Rough Creek Graben not far from its junction with the Reelfoot Rift in

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an area of crossing northwest, north and northeast-trending lineaments
and has no clear relation to any particular structure.

Eastern Tennessee and Kentucky

(Sharpsburg, KY, 1980, I = VII, m_b = 5.2)

Summary

Eastern Tennessee and Kentucky are underlain by Paleozoic sedimentary rock around the Nashville Dome and Cincinnati Arch in the central stable region. Most of the few scattered generally small earthquakes of the region appeared to coincide with a series of northwest-trending geophysical lineaments that apparently represent basement faults. The largest earthquake occurred at Sharpsburg, Kentucky, that lies on a north-trending lineament near the intersection with a northwest-trending one and may represent a weaker and slightly offset extension of the north-trending seismic zone through western Ohio.

Geology

Eastern Tennessee and Kentucky lie in the Central Stable region west of the Appalachian Mountains (Fig. 84). The broad northwest to north-northeast-trending Nashville Dome and Cincinnati Arch, formed of Paleozoic rock, cross the area and are bordered on the east by the Rome Trough (Fig. 101). The Precambrian Rough Creek Graben extends eastward across Kentucky beneath the Paleozoic rock to the central part of the state to the west. The general geology is well discussed by King (1950) and reviewed by Hinze and others (1977).

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The aeromagnetic and gravity data of the region (Johnson and others, 1980; Keller and others, 1980) reflect the structure and show a number of lineaments, apparently representing faults in the Precambrian basement. The regional north-northeast-trending structure is shown by the strike of large anomalies. Crossing this is a series of northwest-trending lineaments, spaced about 80 km (50 miles) apart. Three of these were approximately located previously and thought to have some seismotectonic significance (Seay and Hopkins, 1981) (Fig. 86). This is the same series that coincides with offsets in the seismicity pattern along the northwest side of the Southern and Central Appalachian Mountains seismic zone. A few north and northeast lineament trends are also present.

A study of central Kentucky shows the relation of many of the structures and lineaments. New maps showing geologic structure, Bouguer gravity and aeromagnetic intensity in central Kentucky indicate that several major fault systems in the area are controlled by deep-seated ancient faults (Black and others, 1979) (Figs. 101 and 102). These faults divide the area into major blocks, which in turn are subdivided by lineaments into many rectangular northwest-trending minor blocks. Lineaments are expressed at the surface by structural features such as minor faults and fault swarms, alignments of small structural depressions, and flanks of elongate folds (Fig. 103). They are also expressed geomonphically by aligned sinkholes which are probably controlled by joints (Black and others, 1979). The northwest-trending minor blocks also controlled thickness changes in Ordovician strata and indicate relative vertical movement between them.

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Seismicity

Some relatively minor earthquakes have occurred, scattered about
the region (Stover and others, 1979 b,c,) (Fig. 42 and 100). Several
have reached intensity IV and V and a couple of intensity VI lie near
the edge of the region. The Sharpsburg, Kentucky, earthquake of
intensity VII in 1980 is the largest known in the region (Reinbold and
others, 1981) (Fig. 100). Most of the damage from this earthquake
occurred 50 km north of the epicenter at Maysville, probably due to
amplifications over the alluvium there (Reinbold and others, 1981). No
strong trends can be seen in the epicentral data alone, except a couple
of weak northwest-trending ones.

The epicenters show a correlation with the geophysical lineaments
in the region. Most appear to lie along the northwest-trending
lineaments. The Sharpsburg earthquake and a few that occurred to the
north near the Ohio border coincide with a north-trending lineament,
that approximately lies along 83.8°W Longitude, the approximate contact
of Granville age basement rock (Fig. 101). A northwest-trending
lineament intersects this north-trending one at approximately the
position of the Sharpsburg event. This reasonably good match between
seismicity and northwest-trending lineaments suggests structures along
the lineaments, the boundaries of minor structural blocks, are sources of
the seismicity.

The P-wave, surface wave and aftershock studies suggest the
Sharpsburg earthquake was the result of right-lateral strike-slip motion
along a fault oriented N.40 to 45E. and dipping to the southeast,
although the distribution of the highest intensities showed a northwest
trend (Reinbold and others, 1981) coinciding with the lineament.

Sharpsburg is located near the head of a pre-Pleistocene river system that flowed northward to the west side of Anna, Ohio (Wayne, 1956; Teller, 1973) (Fig. 5) and approximately parallels the north-trend of seismic activity in western Ohio (Docekal, 1970; Hadley and Devine, 1974). The activity at Sharpsburg and to the north of the Ohio River is perhaps a related, albeit weak, extension of the western Ohio activity. (Figs. 42 and 100). The earthquake at Sharpsburg could have been caused from lateral movement along the northwest-trending zone producing north-trending extensional faulting.

Attica

(Attica, N.Y., 1929, I - VII, M_{blg} = 5.2)

Summary

The earthquakes near Attica occur along a northwest-trending fracture, topographic and geophysical lineament that apparently represents a fault zone. The greatest activity and largest known earthquake is at the intersection with the northerly-trending Clarendon-Linden fault system. A few other nearby earthquakes also occur along northwest-trending lineaments, most notable a zone along the Ontario-New York border to the west. Regional gravity data suggest the northwest-trending zone through Attica extends northwestward to Georgian Bay, in Ontario, and southeastward to near Trenton, New Jersey. Several earthquakes occur along it in other locations.

Geology

The Attica area is underlain by gently south dipping lower Paleozoic strata cut by the northerly trending Clarendon-Linden fault system and crossed by a series of northwest-trending lineament zones (Figs. 3 and 104). The Clarendon-Linden fault system forms a high-angle horst and graben structure that may have an overall relative displacement of no more than 100 m down to the east (Van Tyne, 1975; Fackundiny and others 1978; Hutchinson and others, 1979) (Fig. 104). Its post-Ordovician history is unknown, but Mesozoic activity did affect this region as shown by kimberlite dikes a short distance to the

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east (Fisher and others, 1970). The fault system lies approximately along a gravity high (Revetta and Diment, 1971; Hildreth, 1979). The high trends northeastward and its southeast side is followed by the fault system across Lake Ontario. South of the lake the high is offset resulting in a more southerly trend. The fault system is also well expressed as a lineament in the magnetic data of Thompson and Kujawski (1975) (Fig. 105).

A series of northwest-trending lineaments are also expressed in the gravity and magnetic data (Fig. 105). This trend is the most prominent trend in the topography, joints, pop-ups and small faults in the area (Figs. 106, 107 and 108) (Fisher and others, 1970; Fackundiny and others, 1978). The pop-ups and other features produced by the high strain that affects the rocks of the area commonly are aligned northwest (Fig. 108). The gravity high adjacent to the Clarendon-Linden is offset left-laterally 9 to 18 km at three places where it is crossed by these zones. They are most probably Precambrian fault zones that have been reactivated later.

Attica lies on a northwest lineament zone of possible great length; herein called the Attica Lineament. The gravity data (Hildreth, 1979) indicate the lineament may extend northwestward across to the northeast side of Georgian Bay. This crosses western Ontario along a buried pre-glacial valley. Some changes also occur across the southeast projection of the lineament to Trenton, New Jersey. Another paralleling lineament zone passes along the eastern shore of Lake Erie and the Niagara River at the New York State border.

Seismicity

Most of the epicenters in this area are tightly clustered near Attica (Figs. 98 and 109), but a few lie farther afield to produce a northwest alignment. Several others form a slight northwest-trending group along the Ontario-New York border to the west. (Fig. 109). The largest earthquake known in the area occurred at Attica in 1929. It had been assigned an intensity VIII, but a recent reevaluation now places it at VII (Nottis, 1983).

The earthquakes near Attica are aligned along the north-west-trending Attica Lineament; the greatest concentration of activity being where it intersects the Clarendon-Linden fault system. Several other earthquakes, outside the area, also lie near the extensions of this lineament zone into Canada and to the southeast. Those at the western edge of New York also coincide with the northwest-trending lineament zone there. The earthquakes, thus, appear related to northwest-trending fracture zones with the principal activity at an intersection with a north-trending fault system.

Lake Champlain - Hudson River

(Warrensburg, N.Y. 1931, I = VII, M_1 = 3.7)

Summary

The earthquakes in the Lake Champlain-Hudson River lowland lie in a northerly-trending belt of grabens and extensional faults of late Paleozoic or Mesozoic age. Many of the faults are expressed as scarps, due to either Cenozoic movement or differential erosion, and some faults leak carbon dioxide rich waters; a feature usually associated with areas of active faults. This lowland belt is subsiding relative to the Adirondack Mountains on the west and earthquakes may be due to continued movement on the extensional faults. However, the pattern of seismicity is not distinctly enough separated from that in the Adirondack Mountains to rule out possible additional controlling structures in broader regional relations.

Geology

The Lake Champlain-Hudson River area is a lowland underlain by lower to middle Paleozoic sedimentary strata between the Green and Berkshire Mountains on the east and the Precambrian cored Adirondack Mountains on the west (Fisher and others, 1970) (Fig. 110). The formations are affected by thrust faults that developed along the western edge of the Paleozoic deformation of the northern Appalachian Mountains and cut by late Paleozoic or Mesozoic Grabens and extensional faults. These later high-angle faults form a northerly trending belt

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with individual faults trending north to northeast (Fisher and others, 1970) (Fig. 111). These are similar to some in the Adirondack Mountains and the western edge of the zone of extensional faults is difficult to define. Some Mesozoic dikes are present (Fig. 111).

Many of the faults associated with the grabens are well expressed topographically and appear young (Geraghty and Isachsen, 1981) (Young and Putnam, 1979) but it is difficult to determine if the scarps are due to Cenozoic movement or differential erosion. The Lake Champlain and Lake George basins both consist of interconnected grabens with Pleistocene material draped over the faults (Hunt and Dowling, 1981; Isachsen, 1979). At several localities within these lakes the late Pleistocene material appears offset, but this could not be definitely proven (Isachsen, 1979; Hunt and Dowling, 1981) (Figs. 112 and 113). A similar situation exists on land to the south near Saratoga Springs where a series of scarps along faults appear young, but could not be proven (Geraghty and Isachsen, 1981; Young and Putnam, 1979; Putnam and other, 1983) (Fig. 114). The carbon dioxide rich springs there, also suggest Holocene offsets, as such springs elsewhere are in areas of active faults (Young and Putnam, 1979).

The Lake Champlain area is subsiding relative to the uplift of the Adirondack Mountains (Isachsen, 1975) and the lake basin appears to be warping slightly (Barnett and Isachsen, 1980).

Seismicity

Earthquake epicenter lie along the Lake Champlain-Hudson River area and are spatially associated with the youthful appearing grabens and

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normal faults; the largest known earthquake in this area, near Warrensburg occurred on the southwest side of the Lake George Graben (Figs. 54 and 98). However, the epicenters do not form a cluster distinct from those over the Adirondack Mountains to the west (Nottis, 1983) and it is therefore possible that they are not due to a separate source zone, but reflect movement over a wider region.

Merrimack River Valley

(Ossipee, N.H., 1940, I = VII)

Summary

The Merrimack River of New Hampshire flows south obliquely across
the strike of the Paleozoic and older metamorphic and plutonic rock, and
is controlled, at least in part, by north to north-northwest-trending
faults with possible Mesozoic movement. A major regional
northwest-trending fracture zone that crosses the entire State passes
through the Lake Winnipesaukee area at the northern end of the
seismically active zone, near the sites of the 1940 Ossipee earthquakes.
Most earthquakes are located near the Merrimack River Valley and are
probably related to movements on the controlling structures. The
greatest activity is near the intersection with the northwest-trending
fracture zone at the north end of the belt of earthquake activity.

Geology

The Merrimack River Valley of southern New Hampshire trends
obliquely across northeast-striking Metamorphic rock formations and
granite plutons (Billings, 1955) (Fig. 115). The formations have been
traced to Pre-Ordovician ones in Massachusetts (Smith and Barosh, 1983;
M.H. Pease, Jr., personal commun., 1980). They are cut by several
northeast-trending, northwest-dipping thrust faults and a few fault
slices of Silurian and Devonian rock occur locally (Fig. 115). A series
of relatively young high-angle north to northwest-trending faults cut

these rocks and, at least in the south, control the course of the Merrimack River (Fig. 116).

A variety of Mesozoic dikes and plutons of different trends cross the region and perhaps indicate a period of shifting tensional stress (McHone, 1980). These include the northeast-trending Early Jurassic Higganum diabase dike system, that crosses southeast New Hampshire, and the Jurassic to Cretaceous White Mountain series that trends northerly, east of and parallel to the Merrimack River Valley. This series marks a line of ancient volcanoes that may have followed a deep seated tensional fracture zone. The parallelism of the valley to this volcanic chain suggests the structures controlling both may be related. The only known post-Cretaceous faults are a few northwest-trending ones cutting the volcanic rock (Freeman, 1950).

A major northwest-trending zone of small faults and topographic, radar, LANDSAT, magnetic and gravity lineaments crosses central New Hampshire from the Atlantic coast through Lake Winnipesaukee and on to Burlington, Vermont. (Barosh, 1976, 1982) (Fig. 67). This zone must represent a major fracture system, although no great offset is seen across it. Several geologic contacts turn and follow it near Lake Winnipesaukee.

Present-day subsidence is indicated at the New Hampshire and southern Maine coast, but it is not known whether or not it extends inland or if the Merrimack River Valley marks a separate area of subsidence.

Seismicity

A group of epicenters is clustered over the Merrimack River Valley of New Hampshire from Massachusetts to the southern edge of the White Mountains, where the two intensity VII earthquakes occurred in 1940 (Figs. 2 and 117). There is also a slight suggestion of a northeast alignment of epicenters extending towards the Maine border from the southern part of the cluster. This cluster of activity appears separate from that to the southeast centered around Cape Ann.

The seismicity in the river valley may be related to movements of the north to north-northwest-trending faults that control the river. The greatest activity is at the north end of the zone where it intersects the northwest-trending fracture zone, that essentially forms the northern boundary of the seismic zone. The distribution of damage resulting from the widely felt 1981 Gaza earthquake that is within the fracture zone has a definite northwest trend. A minor amount of the seismicity in the south may be partially controlled by one of the northeast-trending fault zones. The few fault plane solutions show both north and northeast trends (Grahm and Chiburis, 1980).

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PART IV: SEISMIC ZONING OF THE ATLANTIC SEABOARD
AND APPALACHIAN REGIONS

Earthquakes in the eastern United States occur in particular areas and, as more earthquakes are located and with greater precision, these areas are becoming better defined and samller (Figs. 2, 118 and 119). The seismically active areas have apparent geologic controls that indicate why earthquakes occur in them and not everywhere. These seismological and geological data, thus, provide definable seismic source zones, even though the degree to which they are understood may vary widely.

The delineation of these seismic source zones when combined with estimates of the maximum earthquake potential of the zones provides the basis for preparing a seismic zone map. Such a map can be prepared in different ways using different parameter: to suit its intended use. The near and far field effects can be combined or separated and shown in values of some size parameter: intensity, magnitude acceleration, or some derivative of these. Relations between these parameters have been (Barosh, 1969) and remain approximate. Due to the proliferations of magnitude scales, the lack of measured magnitudes over most of the length of the record and variation of measured magnitude values for intensities, the use of intensity values is considered best at this time to present a more uniform treatment of earthquake effects, despite some skewing of values due to different ground conditions. Only the near field effects will be shown at this time; the far field can be estimated by application of attenuation rates to the near field. The zoning map presented here, therefore, is in terms of epicentral intensity values

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of average firm ground. Return times have not been considered, and this map attempts to describe the hazard, above the source but not the risk.

ASSUMPTIONS USED IN SEISMIC ZONATION

1) Known clusters of epicenters represent all the active source zones in the region, new areas of activity are unlikely and the seismic gaps are permanent. (earthquakes have been and are clustered into particular areas in historic time and these zones have been shrinking in size as earthquake locations have improved).

2) Tight clusters of activity may be indicative of potential for larger earthquakes. (Generally the larger earthquakes occur where greater numbers of smaller ones do).

3) Tectonic environment imposes limits to strain build up and maximum size of earthquakes. (Potential for major earthquakes is not a function of return time alone). (The earthquakes at Charleston, S.C., in 1886 Giles County, Va., in 1897, Massena N.Y., in 1944 and Sharpsburg, Ky., in 1980 are taken as approximate maximums in the region).

4) Most areas in the Eastern United States have not experienced their maximum potential earthquakes during the 250 to 400 years length of historic records. (At most places, therefore, the intensity levels are placed one unit higher than experienced; this may over rate the maximum expected intensity level in places).

5) <u>Recorded epicentral intensities represent average firm ground.</u> (Not valid everywhere as some may represent poor ground conditions and this tends to raise intensity values locally).

6) <u>There are definable geologic reasons for most, and presumably all, seismic zones.</u> (There is a reason for the permanent seismic gaps).

7) <u>The basic geologic cause for earthquakes is similar across the entire region, although manifestations may vary locally, and a uniform treatment in zonation is justified.</u>

8) <u>General seismic zones have local more active places.</u> (The larger earthquakes do not necessarily occur everywhere within a zone, but are localized due to geologic controls, commonly structural intersections).

9) <u>Earthquakes located on a fault may mean that a particular segment, but not necessarily all of it is active</u> (estimating potential earthquake size based on fault length does not apply in the eastern United States as earthquake distribution pattern indicates only local possible reactivation of individual faults).

SEISMIC ZONING VERSUS SEISMOTECTONIC PROVINCES

The seismic zoning of source zones presented here is similar to, but different from delineation of seismotectonic provinces, tectonically definable provinces with a given uniform earthquake potential. The basic approach has been the same, but the boundaries are not shown where the level is the same on either side of a boundary. In addition, it became very apparent in this study the level of potential earthquake size within a single tectonic feature may vary from place to place and necessitate creating sub-provinces or at least recognizing that "hot spots" exist. Where this is apparent, it is these sub-provinces that have been delineated and joined. A third difference is the recognition that active tectonic provinces may overlap. This is a problem in presentation of these provinces, but appears to be of little consequence in the zonation used here.

SEISMIC ZONING VERSUS SOURCE ZONES

Seismic zoning of source zones commonly results in boundaries different from the determined source zone and usually encompasses a larger area. This results from taking into account the locations of epicenters in the historic record. The differences may be due perhaps to (a) inaccuracy of epicentral location, (b) fore and aftershocks from adjustments in the rock off to the side of the source structure, (c) local reactivation of a variety of cross-structures, (d) activity only at intersections of structures, (e) activity in areas of sags, and (f) merging activity from adjacent sources or other similar situations (Fig. 120).

Increase in accuracy of locations tends to narrow the difference, but until this is done, it is better to take the broader zone in case active structures may lie to the side of the main one.

Activity at structural intersections may be common in the region. The parts of these structures not associated with earthquakes are apparently not active and activity across a segment of a structure, by no means, implies all of it to be active.

MAXIMUM EPICENTRAL INTENSITIES EXPECTED FROM SEISMIC ZONES

Estimating maximum epicentral intensities can be done different ways. Frequency-Magnitude studies are often used to indicate maximum intensity for local areas, but the method involves many assumptions (Johnston, 1981) and available data are only marginally acceptable at many places in the east. Intensity-return time studies, which also have many problems (Hand and Hoskins, 1981) are a good method to compare sites, but are difficult to use to establish upper bounds, although an arbitrary time interval might be chosen. It is geologically illogical to extend return time curves too far and expect every region to have a great earthquake with sufficient time. Another approach is more regional, a seismotectonic one in which a number of local areas in the same or similar tectonic environment are considered together. This may lead to a more uniform treatment over a large area. The whole places some constraints on the parts.

Most areas in the eastern United States have probably not experienced their maximum potential earthquake, but some certainly have over the historic record and all potential epicentral intensities should not be automotically raised over the experienced ones. As larger earthquakes generally occur where there are clusters of ones with lesser intensities and non-felt ones, a relative few number of these smaller events near a larger earthquake may indicate the maximum intensity has been felt. Larger intensity earthquakes may also be greater than these experienced in similar nearby environments with similar numbers of experienced epicentral intensities, and indicate they are about maximum

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level or perhaps on poor ground. On this basis a few earthquakes in the region are considered to be at or very near their maximum size (Table 1).

Table 1. Earthquakes Estimated to have reached their approximate maximum epicentral Intensity for Their Tectonic Setting.

Earthquake	Intensity (MM)
Charleston, S.C., 1886	X
Giles County, Va.,1897	VIII
Massena, N.Y., 1944	VIII
Sharpsburg, Ky., 1980	VII

The intensity values for historic earthquakes is, for the most part, assumed to be the average for firm ground and the zonation is done in these terms. The deviation from this intensity in the records is probably more towards poorer ground and higher intensity but how much more this may be above average firm ground is difficult to estimate just as the variation between firm ground and bedrock is.

The assignment of potential maximum epicentral intensity values in the Modified Mercalli scale as shown in the zonation (Figs. 121, 122) may err on the high side and be conservative in the sense of applying them to engineering parameters. Within the larger zones it is doubtful that the value applies everywhere. There may be many local pockets to which a lesser value may be applicable, but until further

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work establishes criteria or more detailed data by which they can be delineated and separated, they are included.

The Southeast Georgia Embayment has three areas of seismicity: northeast Florida, central Georgia and the main activity in North Carolina. The Charleston earthquake of 1886 with an epicentral intensity of X (Bollinger, 1977) is the largest known on the East Coast of the United States and is considered to represent the approximate limit of strain buildup in the coastal region (Nuttli and Hermann, 1978). Armbruster and Seeber (1981) consider vibrational damage in the Charleston-Summerville area is not higher than other areas with intensity VIII. The felt area, however, is larger than other intensity VIII earthquakes in the United States and it might be closer to IX on firm ground. If Armbruster and Seeber's calculations are correct, then the maximum potential earthquake would probably not exceed a IX if not a X assignment might be more appropriate for a maximum.

The highest intensity has been assigned only to the Charleston-Summerville area. The present and past seismicity indicate a northwest-trending belt of activity from Charleston across South Carolina that passes through Columbia. Here, as well as elsewhere in the Eastern United States activity is highly variable along such belts and apparently permanent gaps are present. The Union County earthquake of 1913 is also on this trend. It was originally assigned an intensity VIII (Taber, 1913), but was reevaluated to a VI-VII by MacCarthy (1957). A recent review of all available information concluded an intensity VII is the most reasonable interpretation, although an VIII may be interpreted because of some local susceptibility (Titcomb and Hancock, 1981). It is still the second largest earthquake recorded in the State

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(Reagor and others, 1980) and an area of present day activity (Bollinger and Mathena, 1982). It has a value and geologic position, between the Fall Line and the Brevard zone, very similar to that of the Arvonia earthquake of 1875 in central Virginia and is similarly difficult to evaluate as to its maximum potential for earthquakes. An intensity VIII earthquake was estimated to affect Greenville, South Carolina, to the west and with only an experienced VI, about every 27,000 years and a IX every 250,000 years from the return time curves of Hand and Hoskins (1981). In light of this a maximum earthquake of intensity VIII for the Union County area may not be too high. The other areas in the embayment with scattered earthquakes of intensity V and VI have been rated at intensity VII for a maximum. The intensity assesment in this region is in general agreement with a recent study for a dam on the Savannah River (Titcomb and Hancock, 1981).

The Chesapeake-Delaware Embayment appears to only have significant earthquake potential in central Virginia where a maximum of VII might be expected, except for a slightly higher value, VII-VIII, in the cluster of activity near Arvonia, west of Richmond. The Arvonia earthquake of 1875 was listed as VII or more (Coffman and others, 1982), but reevaluated downward to a VI. The largest earthquakes listed for the Raritan Embayment are the 1737 and 1884 events near New York City and the Asbury Park, New Jersey event of 1927 of intensity VII (Coffman and others, 1982). A recent reevaluation of the earthquakes of the area has indicated that these were more likely intensity VI, VI and V respectively, and that they along with a few others were relocated within or at the edge of Raritan Bay (Nottis, 1983). The number of intensity V and VI earthquake epicenters

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in the region indicate the potential for a VII event, but the general tectonic similarity of the area, with northern South Carolina, Central Virginia, and Cape Ann areas and evidence for subsidence in the bay suggests that a maximum intensity of VIII might be possible in Raritan Bay.

The 1791 East Haddam, Connecticut, earthquake is the largest known in the Moodus area. This was originally evaluated as an intensity VIII, but has been reevaluated and found that the epicentral intensity does not exceed VII (Boston Edison Co., 1976). Four other earthquakes of intensity VII occurred in the 16th century (Nottis, 1983). The concentration of these suggests an VIII might be possible near the active zone along the Salmon River.

The Narragansett Bay area and adjacent southeastern Massachusetts, although having a distinct cluster of epicenters, have experienced a lower level of activity than either the adjacent Moodus or Cape Ann areas. The several intensity V's and two VI's in the region (Nottis, 1983) do not appear to warrant potential intensity as high as VIII and an intensity of VII seems more appropriate. This also seems to apply to the pockets of activity along the Maine coast.

The area offshore of Cape Ann probably experienced an intensity VIII earthquake in 1755 and a VII in 1727 and four VI's have occurred onshore to the west (Boston Edison Co., 1976; Nottis, 1983). This is a little less than the concentration near Charleston, South Carolina, more than at Moodus and the Ossipee, New Hampshire area and much less than at La Malbaie, Quebec. Evaluation of this relative level of seismicity and noting the tectonic similarities with the Charleston region suggests the offshore area around the 1755 event may have a potential for an intensity IX earthquake. The adjacent coastal area to

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the west would be judged to have the potential for a VII, but whether or not it might reach VIII is difficult to evaluate. Given that a few earthquakes offshore may have neglected or assigned lower values than they had and the position to err on the high side, a fringing area of intensity VII-VIII would not be too unreasonable.

The Upland areas do not have quite as high level of intensity as the coastal ones, but do contain a few areas of relatively higher intensities. The Southern and Central Appalachian highlands contain a broad zone with scattered intensity V and VI earthquakes and a few concentrations of earthquakes with intensities of VII or even VIII events. A potential epicentral intensity of VII appears appropriate for most of the zone, although both ends, southwest of Birmingham and in southwest Connecticut, have only scattered V's or less. These ends have been included with the general zone as there is no obvious change in the probable geologic controls. Areas with higher intensities and generally concentrated activity occur at Birmingham, Alabama, Knoxville, Tennessee, Giles County, Virginia, and Wilmington, Delaware, and possibly western North Carolina. Birmingham, Alabama has an intensity VII amongst a group of lesser events. Another reported VII off to the northeast was apparently an explosion (Drahovzal and Keener, 1976). The Knoxville area has had a VII, three VI's and six V's. The largest earthquake in Virginia and the second largest in the southeast United States occurred in Giles County, Virginia on May 31, 1897 (Bollinger and Wheeler, 1983). It had an epicentral intensity of VIII and a total felt area of 895,000 square kilometers (Bollinger and Hopper, 1971). The magnitude of M_b for this earthquake was estimated to be 5.8 using the intensity-attenuative method by Nuttli and others (1979). A few

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intensity VI's have also occurred in the area. Wilmington, Delaware experiencecd an intensity VII in 1871 and in the vicinity there have been two VI's and four V's and numerous small events. In western North Carolina strong earthquakes occurred near Asheville and 55 km (35 miles) to the northeast amidst a cluster of smaller events. These are listed at intensity VI by Coffman and others (1982), but changed to VII by Moneymaker (1957) and Reagor and others (1980). That at Asheville is on the same trend of activity as Charleston-Summerville and Union County, South Carolina. Bollinger (1981) estimated the potential earthquake size for Giles County, Virginia, to have a magnitude, Ms, of 7, and epicentral intensity of IX. However, as Giles County is the only area in the entire upland region to experience an VIII and other similarly tectonically situated areas have only experienced VII's, it may well be that an VIII represents the upward limit of strain accummulation in the region. All of these areas are therefore assigned a maximum epicentral intensity of VIII, except for western North Carolina where a VII-VIII is applied due to uncertainty on the intensities experienced.

Several other areas in the region have had smaller clusters of events with a maximum intensity of VI; such as Chattanooga, Tennessee, southwesternmost Virginia, Charlottesville, Virginia, northern Virginia, and the New Jersey and Hudson Highlands. An analysis by Fischer (1981) of the latter region, which is traversed by the Ramapo fault, that bounds the northwest side of the early Mesozoic Newark Graben, found an "upper limit of credible events that may be associated with the Romapo is not more than intensity VI or VII". This is in agreement with others that have reviewed the geology and seismicity (Thompson, 1983; Tillman, 1983) and is consistent with the general levels of adjoining areas in

New England. VII is also taken as the maximum credible earthquake for the areas of the other lesser clusters of earthquakes in the region.

The Adirondack Mountains have experienced numerous earthquakes, but they are all small with only a couple of intensity VI's. The higher intensity earthquakes of the region occur in the lowland structural zones bordering the mountains to the east and northwest. An intensity VI-VII would therefore appear to be the maximum credible event.

The Central New Brunswick Highland that extends westward into easternmost Maine also has only a few scattered events of intensity VI, similar to the 1982 Miramichi earthquake. An intensity level of VI-VII also appears to be the maximum event here.

The Interior zone has lower intensity values than either the Coastal or Upland zones, except along the St. Lawrence-Arkansas trend. The northwest of the Appalachian Highlands, from northeastern Mississippi to the southwest edge of West Virginia, is characterized by intensity V or less events that appear to be scattered along several northwest-trending and a couple of north-trending zones. A few intensity VI events occur on a couple of these zones near the border of the region and some may have the potential for intensity VI earthquakes. The area between the zones appears aseismic.

The major cross zone of seismic activity in the New Madrid region, zone B of Johnston (1981), continues southeastward as a weakly active zone into northwest Alabama, but appears to only have high intensities associated with it in the New Madrid region. The north-northwest zone of activity extending across Alabama from Birmingham appears to only have significant activity along it northward to the edge of Tennessee. This zone may have potential for intensity VII in Alabama, but is weak

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farther north until the edge of the Wabash Valley fault zone is reached. Drahovzal and Keener (1976) estimated this Alabama part of the zone to have intensity VI-VII earthquakes 0.6-1.2 times/century. The largest event in the eastern Tennessee and Kentucky area is that of an intensity VII earthquake at Sharpsburg. This appeared unusually high perhaps due to thick alluvium (Reinbold and others, 1981) and an intensity VII on firm ground should be the credible earthquake.

From West Virginia to the Mohawk River valley of central New York are aseismic areas separated by zones with a few events of V or less. Some of these zones also appear to be aligned northwest. A general assessment of intensity VI also applies to these more active places. The two VI's and V's forming a northwest zone along the Mohawk River Valley appear to warrant a VII similar to the adjacent Champlain-Hudson region.

The active area in northwest New York near Attica and that in the Massena area to the northeast lie in the more active St. Lawrence-Arkansas trend as does the western Ohio zone to the southeast. Attica had a maximum intensity of VII in 1929 and Massena an VIII in 1944 (Nottis, 1983). The Attica area has had more intensity VI and V events, whereas Massena has many more IV's and III's. The VIII at Massena looks unusual because of the lack of VII's and VI's there and an VIII is perhaps the maximum credible event here as well as the Attica area.

The northern part of the interior zone is characterized by several north and a few northwest-trending zones of activity separated by areas with little or no activity. The north-trending zone of scattered activity along the Connecticut River Valley, which has had a few V events and a few other such zones may have the potential for a VI. The

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areas of lower potential are included with these zones in a general region assigned to intensity VI. A few zones exhibit greater activity. Some Intensity VI events and an intensity VII near Lake George occur along the Lake Champlain-Hudson zone. This area may generally be assigned to a VII for a maximum epicentral intensity, but the intensity VII near Lake George may not be the maximum in this area of possible fault scarps. The area around the southern half of the Lake and to the southwest of it may warrant a VII-VIII rating.

The north-trending zone along the Merrimack River Valley in New Hampshire has experienced several V's a few VI's and, at the north end, two VII's. An intensity VII appears the maximum potential for most of the valley area, but the north end, where northwest-trending geologic, geophysical and LANDSAT features cross the northerly-trending valley area and the higher intensities are located, may have a credible earthquake of VIII.

The activity along the north and northwest-trending zones north of Casco and Penobscot Bays merge with the activity on the coast and have similar scattered events of intensity V and VI. These are all considered areas with a possible maximum epicentral intensity of VII.

Some of the reasoning for particular intensity assignments varies from place to place across the region zoned in the eastern United States, but it is hoped that an overall uniformity of intensity levels has been achieved. When future work suggests a change of level for an area, other similar areas should be examined to judge whether the change should be made and, if so, should the other areas be changed as well.

Explanation of values used in maximum epicentral intensity expected on firm ground for seismic source zones in the Eastern United States (Figs. 121 and 122) is as follows:

V   Areas of no known earthquakes, generalized boundaries
         that avoid large known suspected fracture zones, but
         otherwise not drawn on geologic basis. The level
         chosen at V as rapid freezing may produce this
         intensity locally in the northern half of the area.

VI   Areas that have only experienced scattered Intensity V
         earthquakes. Most of the area is considered to rate at
         intensity V level, but it contains local more active areas
         with potential for I = VI (might separate with more detailed
         zonation).

VII   Areas with an indentifiable clusters of earthquakes
         forming a seismic zone that has only experienced Intensity VI
         and probably has potential for intensity VII in general
         vicinity of intensity VI or areas where an experienced
         intensity VII is the estimated maximum. Boundaries related to
         geologic features and generally drawn on them, but modified
         locally to conform to earthquake distribution.

VIII   Areas with cluster of epicenters within a seismic zone
         with an experienced intensity of or near VII or areas where an
         experienced intensity VIII is the estimated maximum. Where

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near (intensity VII changed to VI by recent reevaluations)
the areas have geologic environments similar to
other areas with an experienced Intensity VIII. Related to
geologic source zone and boundaries generalized around source
zone and related earthquakes.

IX   Areas with cluster of epicenters within a seismic zone with an
experienced intensity of VIII. Probably aproaching limit of
strain buildup in areas of shallow earthquakes; Related to
geologic source zone and boundaries generalized around source
zone and related earthquakes.

X   Areas with cluster of numerous epicenters within a seismic
zone with an experienced intensity of IX or X. Probable limit
of strain buildup on the East Coast of United States.
Related to geologic source zone and boundaries generalized
around source zone and related earthquakes.

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PART VII: SUMMARY AND CONCLUSIONS

Earthquakes in the eastern United States occur in relatively restricted active areas. Nearly all these areas have geologic structures indicating the apparent cause of seismicity. No new areas of significant potential activity are suggested. The cause of the seismicity is apparently related to extensional strain related to the continuing widening of the North Atlantic basin.

Characteristic features found in many of the seismic areas are: (1) present day vertical movement, mainly subsidence; (2) past vertical movement as expressed by presence of mid-Mesozoic or older grabens and sedimentary basins, particularly Late Cretaceous or Tertiary; (3) moderate to major on and off shore northwest-trending fracture zones forming intersections with north and northeast-trending fault zones; (4) extensional faults with only slight Tertiary and Holocene surface movement; (5) lack of major fault zones active along their entire length; and (6) probably by highly strained basement rock.

The active areas form coastal, upland and interior groups. The coastal areas are those near the Early Tertiary Atlantic and Gulf Coast shorelines. These have experienced the largest earthquakes. The areas south of New England are marked by embayments, basins of Late Cretaceous and Tertiary sediments, that indicate areas of subsidence. They are closely associated with northwest-trending fracture systems; major late Precambrian grabens on the Gulf coast and shoreward projections of oceanic fracture zones on the mid and south Atlantic Coast. The exception being the northeast-trending Reelfoot Rift beneath the Mississippi Embayment, which is part of a transcontinental sag from Arkansas to the

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Gulf of St. Lawrence. The upland zones are in areas that are rising inland from the Atlantic Coast and probably have been since the Late Jurassic. Earthquake locations within these areas appear related to northwest and northeast fracture intersections. The interior zone lies inland from these and seismicity there is apparently related to inter-sections of northwest-trending fractures and north-trending extensional faults.

All of this activity appears caused by movements related to the continued widening of the North Atlantic Basin. Movements are indicated to have began in the Late Cretaceous, where the sedimentary record is preserved. The coastal zones bowed down as the inland areas adjacent to the Atlantic Coast rose. Northwest-trending oceanic frac-ture zones apparently connect with onshore ones and differential move-ments between the blocks bounded by them produced variations in thick-ness of the sediments deposited. This created northwest-trending basinal sags and adjacent arches. These basins can be very complex in detail where grabens in the basement are intersected and reactivated.

The recognition and delineation of the seismic source zones from seismologic and geologic data enables a seismic zonation map, in terms of epicentral intensity, to be prepared by determining maximum expected epicentral intensities for the zones. Comparison of similar source zones and their seismic history facilitates determining consistent maximum epicentral intensities, for, whereas most areas may have not experienced the maximum intensity during historic time, some have. The resulting map incorporates far more factual data and avoids many of the questionable assumptions used in probabilistic studies and results in a

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map that presents a far more reasonable and accurate presentation of the earthquake potential of the East Coast.

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EASTERN U. S. HIGH SEISMICITY AREAS
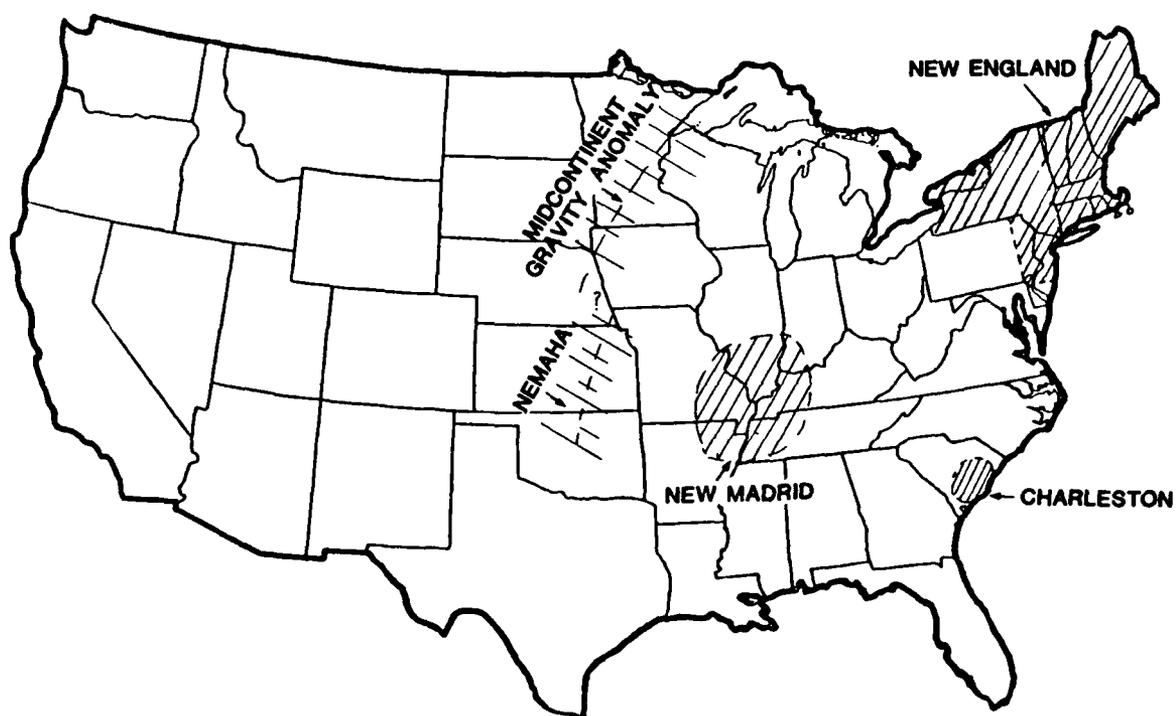
Figure 1. Index map showing seismotectonic studies in the eastern United States sponsored by the US Nuclear Regulatory Commission
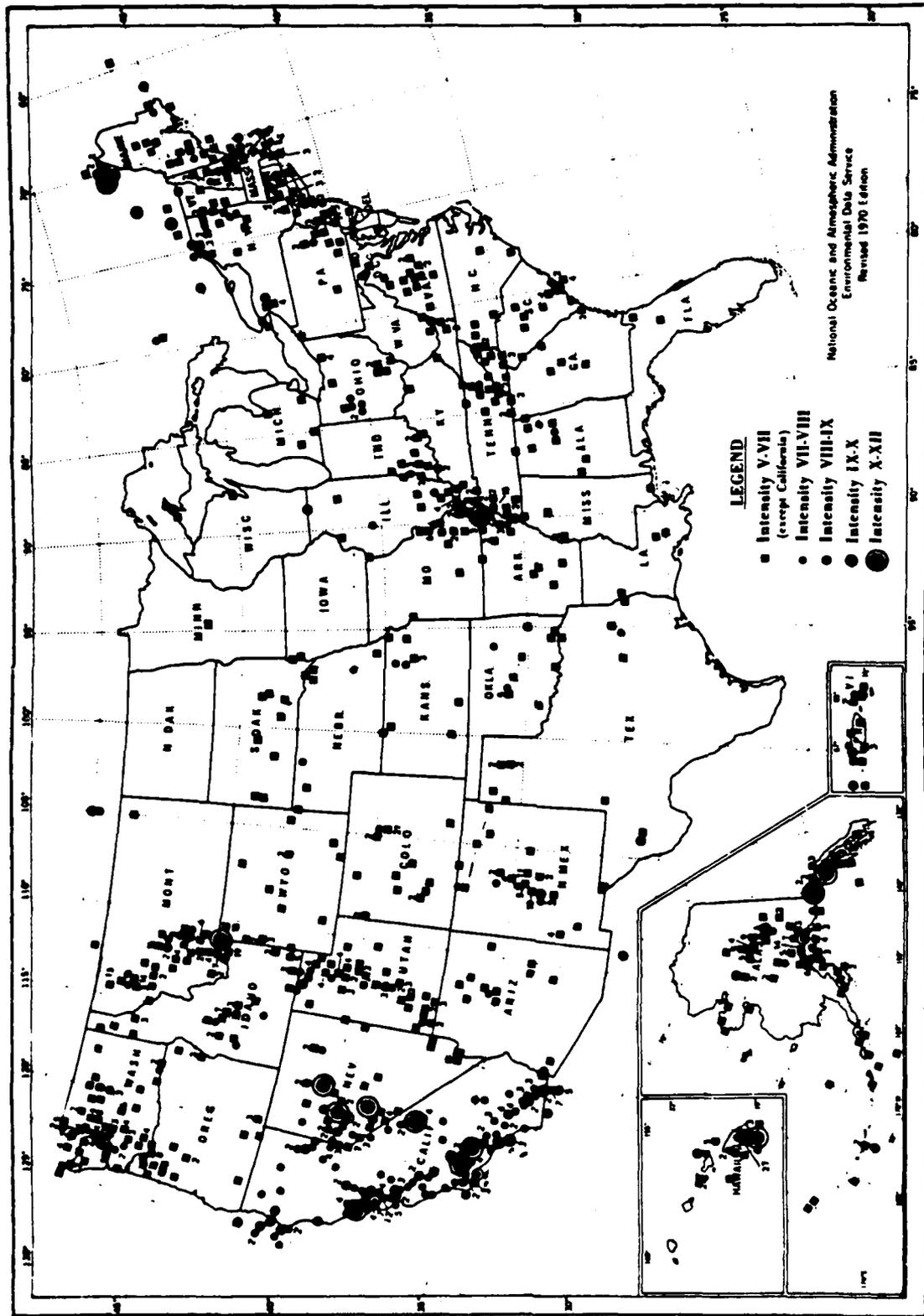
Figure 2.   Epicentral map of the United States showing earthquakes of
intensity V and above through 1970 (Coffman and others, 1982)

Figure 3. Map showing seismic frequency of the eastern United States and adjacent Canada (modified from Hadley and Devine, 1974)
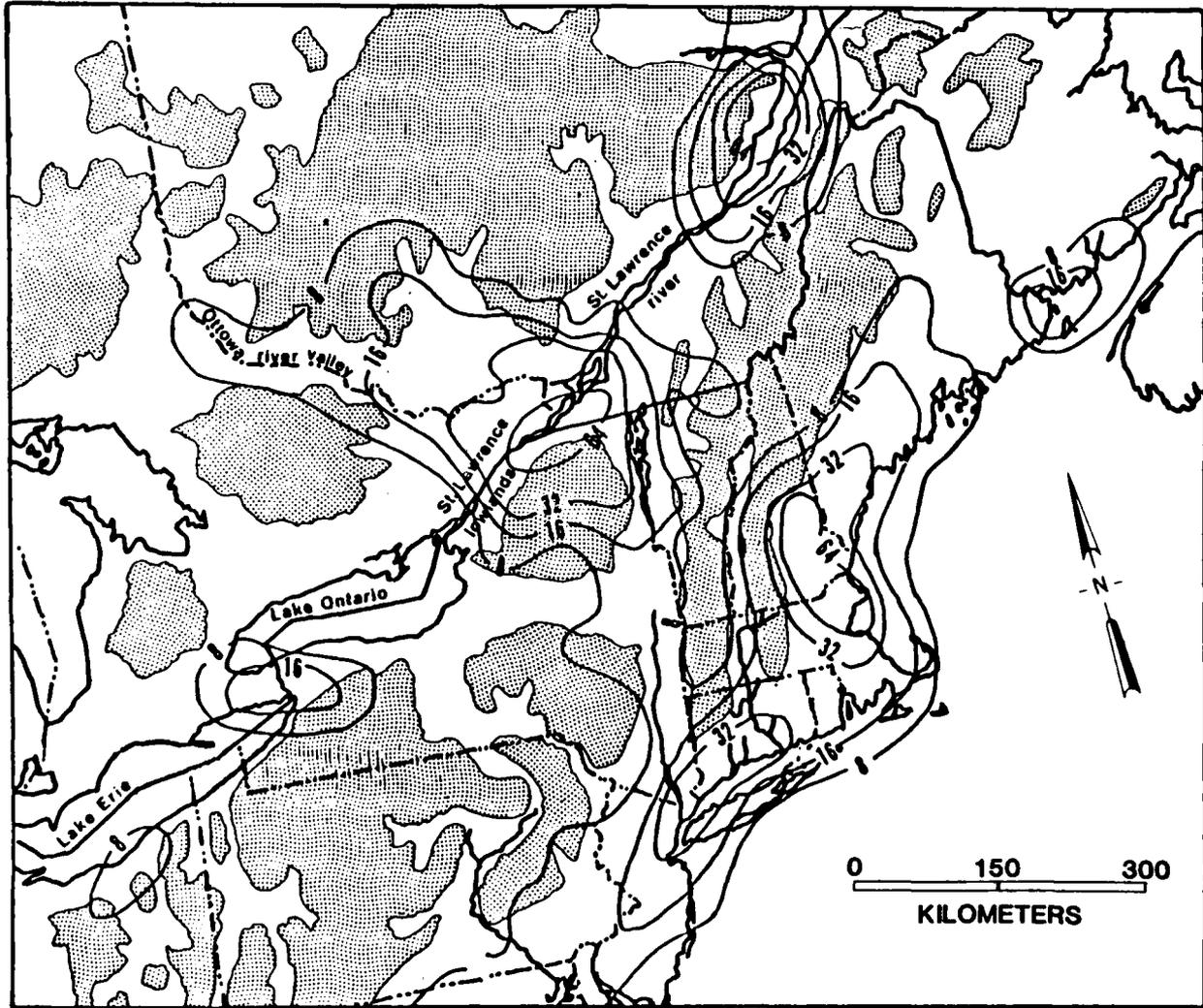
Figure 4. Map of the northeastern United States and adjacent Canada showing areas (strippled) of altitude over 300m and seismic frequency contours (total number of earthquake epicenters per 10^4 km² during the period 1800-1972 (seismic contours from Hadley and Devine, 1974)
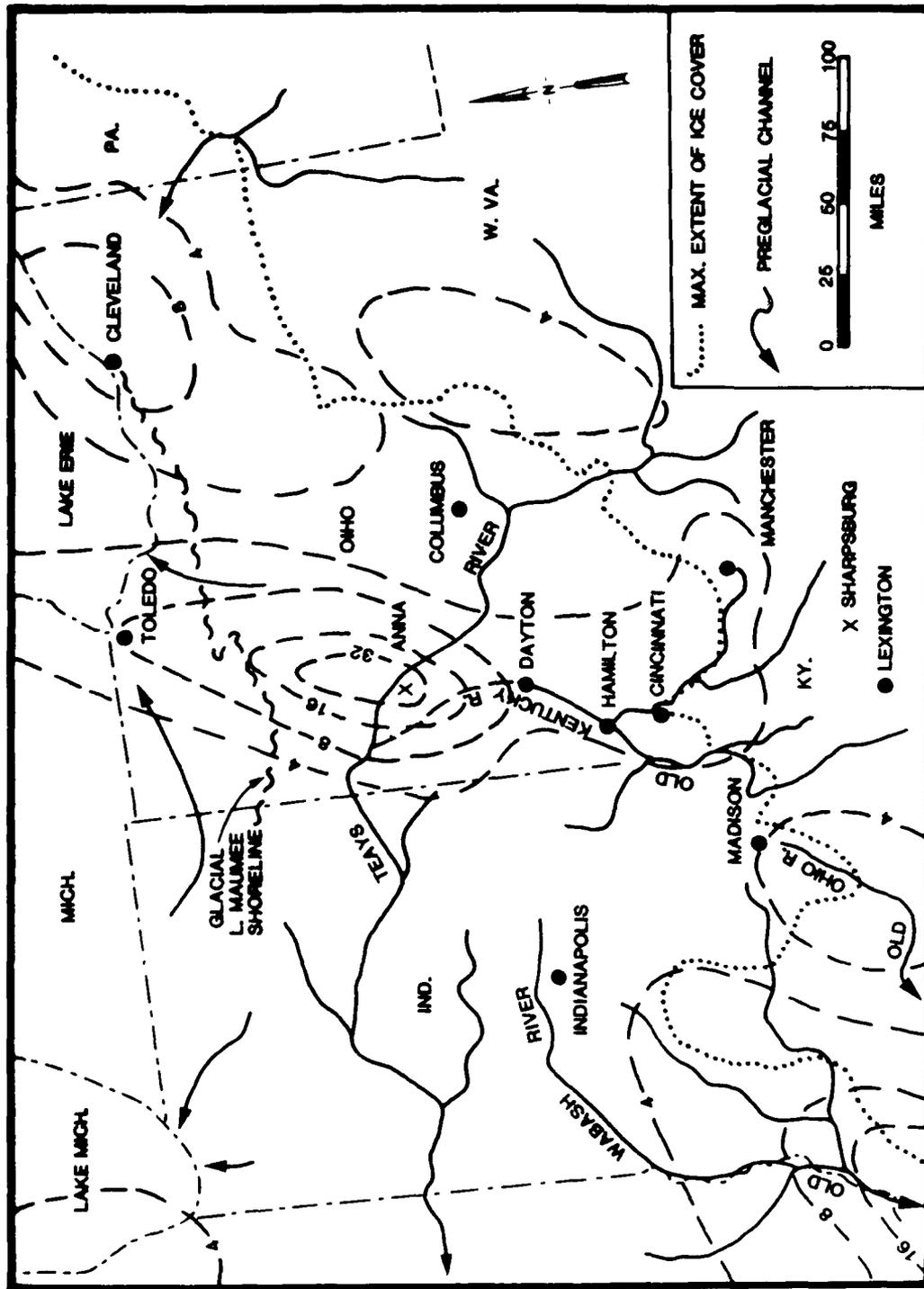
Figure 5. Map showing pre-glacial drainage in Ohio, Indiana, and northern Kentucky (from Wayne, 1956, and Teller, 1973) and seismic frequency contours (from Fig. 3). The approximate locations of the Anna, Ohio, and Sharpsburg, Kentucky, earthquakes are marked by X's
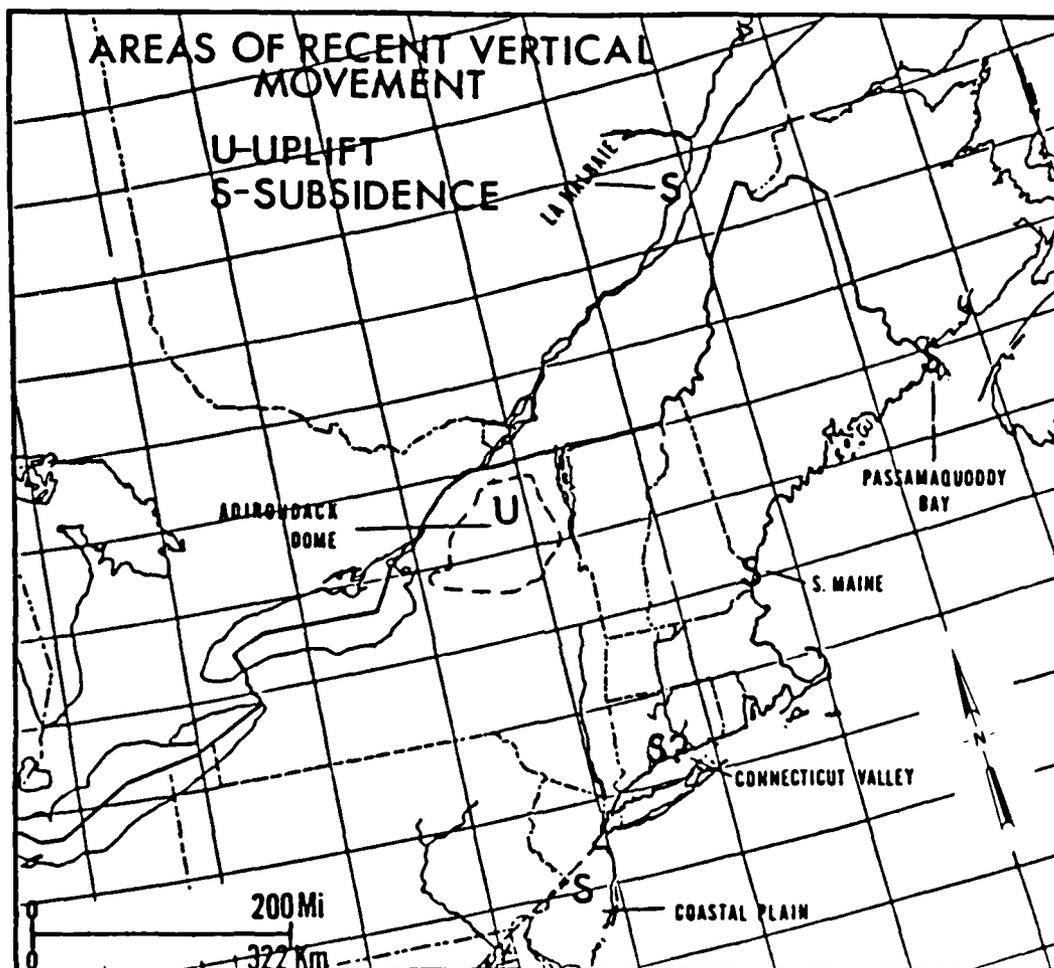
Figure 6. Map showing areas of recent vertical movement in the northeastern United States and adjacent Canada (Barosh, 1982a)
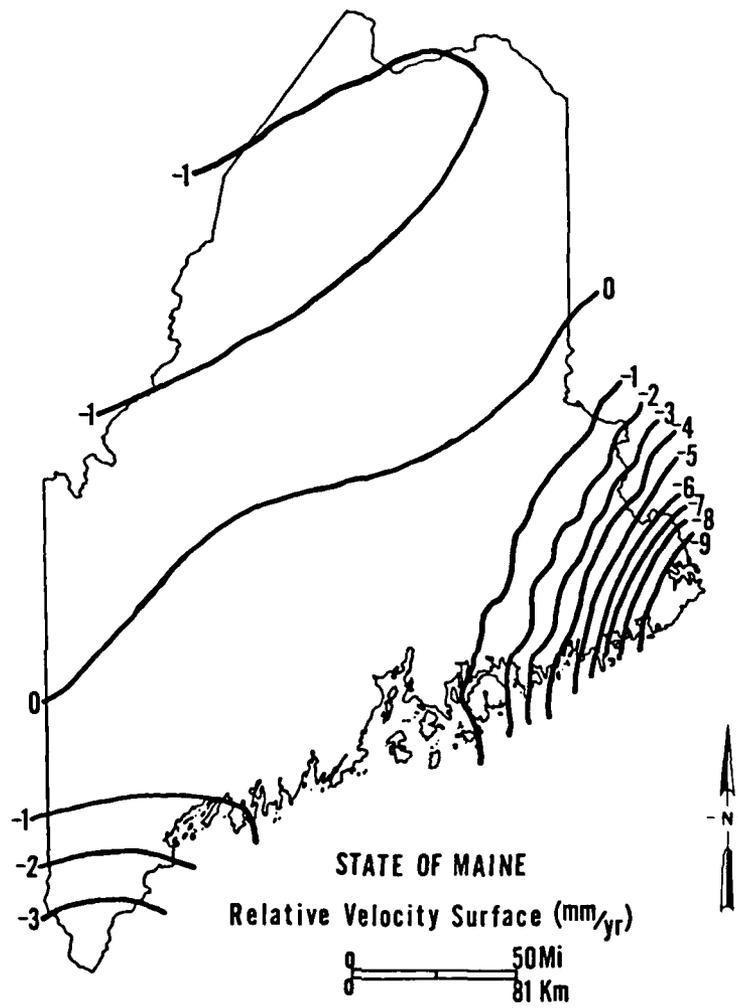
Figure 7. Map of Maine showing the relative rate of present-day vertical movement derived from leveling surveys (Tyler and Ladd, 1981)
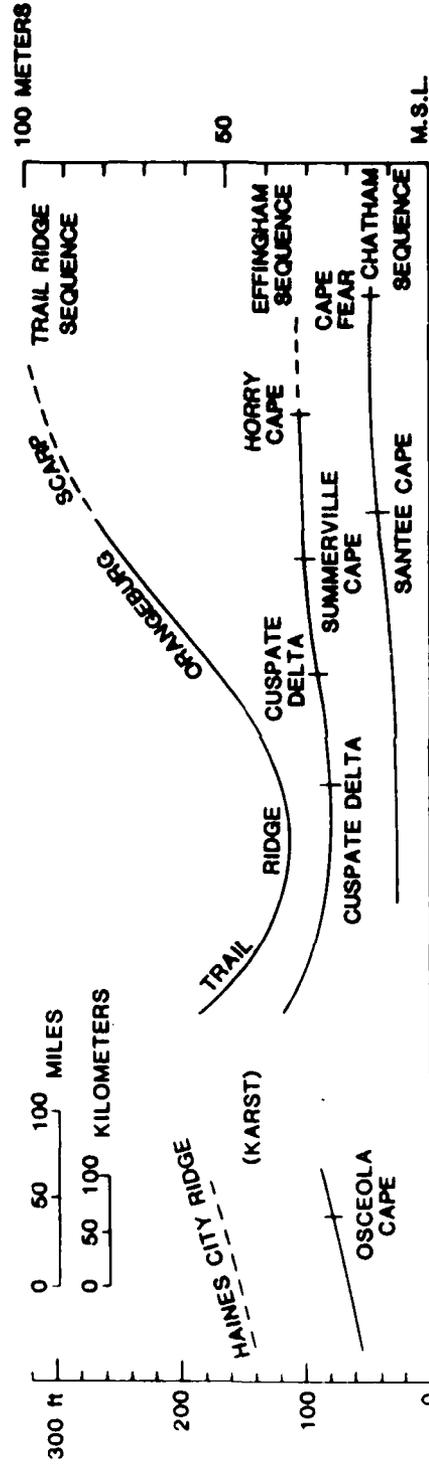
Figure 8. Diagram showing tentative curves of shoreline warping along the south Atlantic Coast (from Winkler and Howard, 1977) and their position relative to other geologic features (compiled from King and Beikamn, 1974; and Cohee, 1961). The curves represent maximum height of transgression inferred for each shoreline sequence. Dashed lines indicate considerable uncertainty about sea levels. Vertical scale is altitude above present mean sea-level
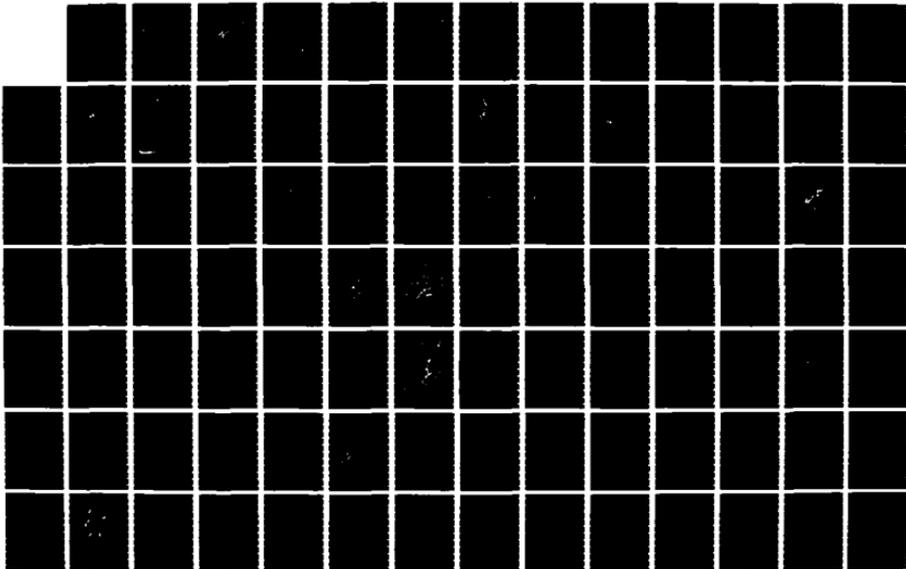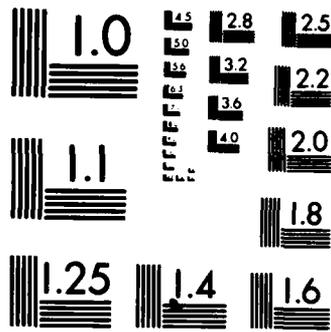
MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS-1963-A
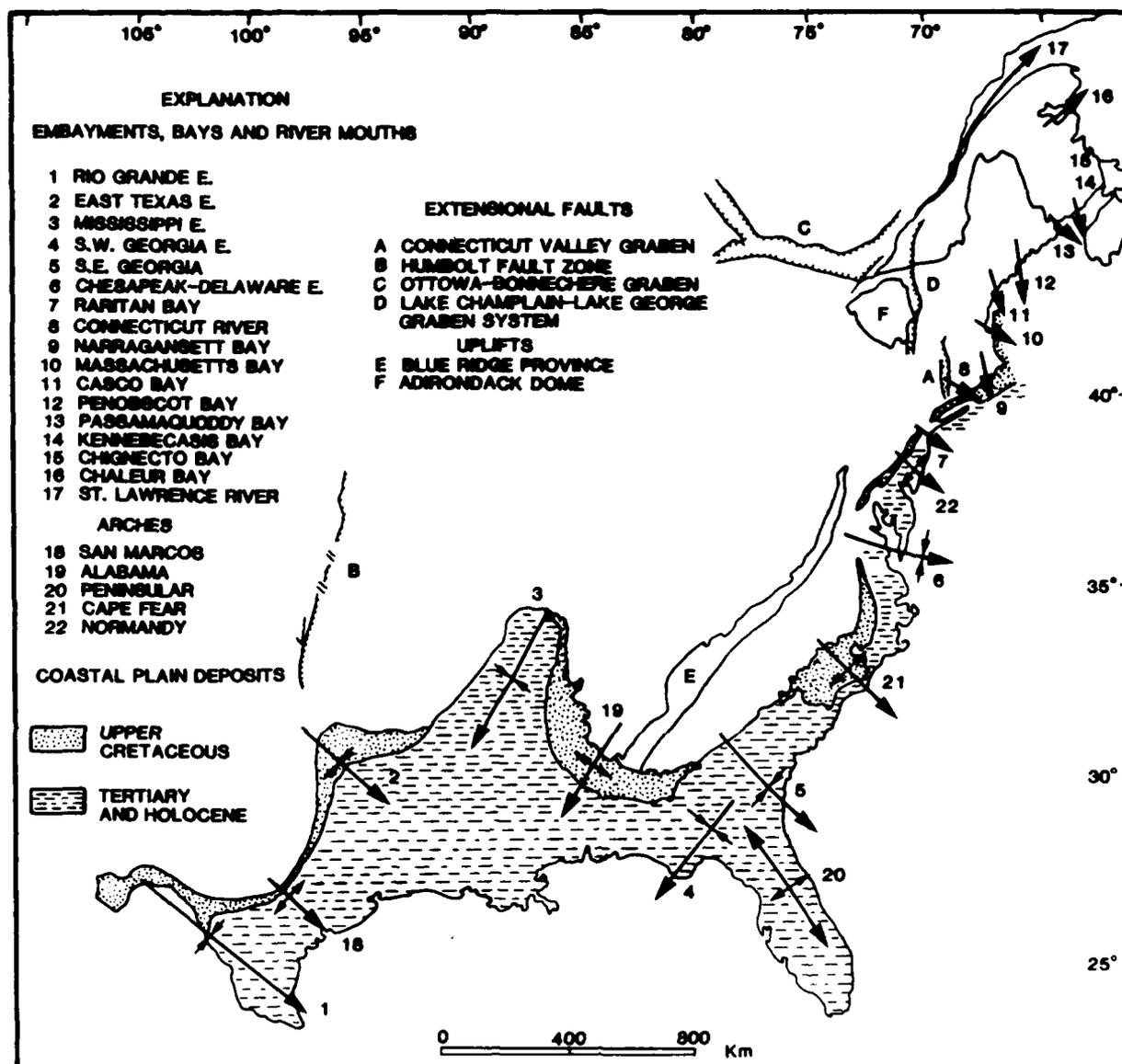
Figure 9. Map of the eastern United States and adjacent Canada showing embayments along the Late Cretaceous continental margin, bays in the northeast, areas of uplifts and selected extensional faults and grabens
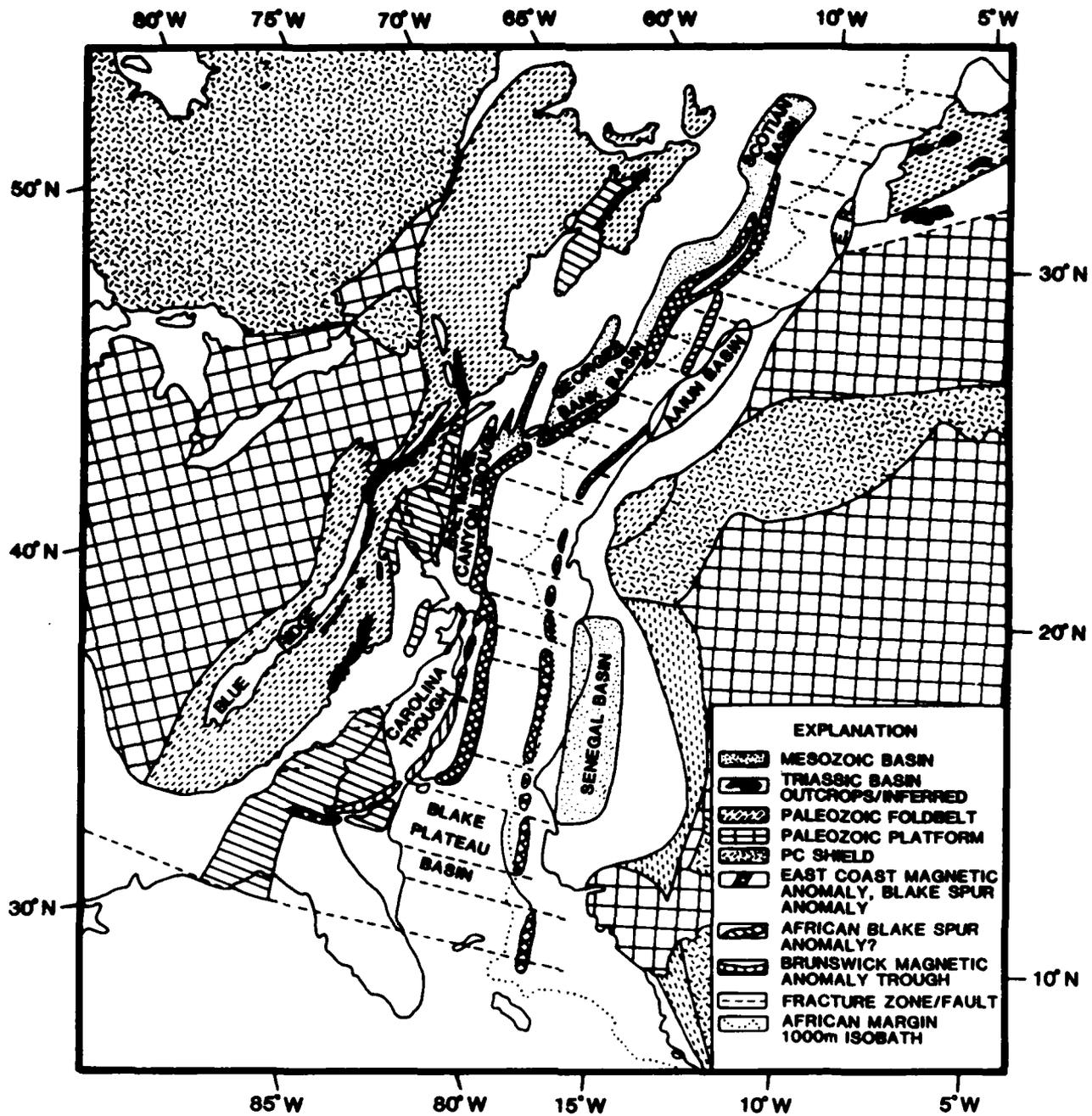
Figure 10. Map showing reconstruction of the North Atlantic Basin and adjacent area about 175 m.y.B.P (Klitgord and Behrendt, 1979, Fig. 15)

Middle of
Triassic

175 M.Y.
Ago

End of
Jurassic

End of
Cretaceous

Present

100 M.Y.

0   100 Km.

sedimentary rock

volcanic rock

continental basement

oceanic basement
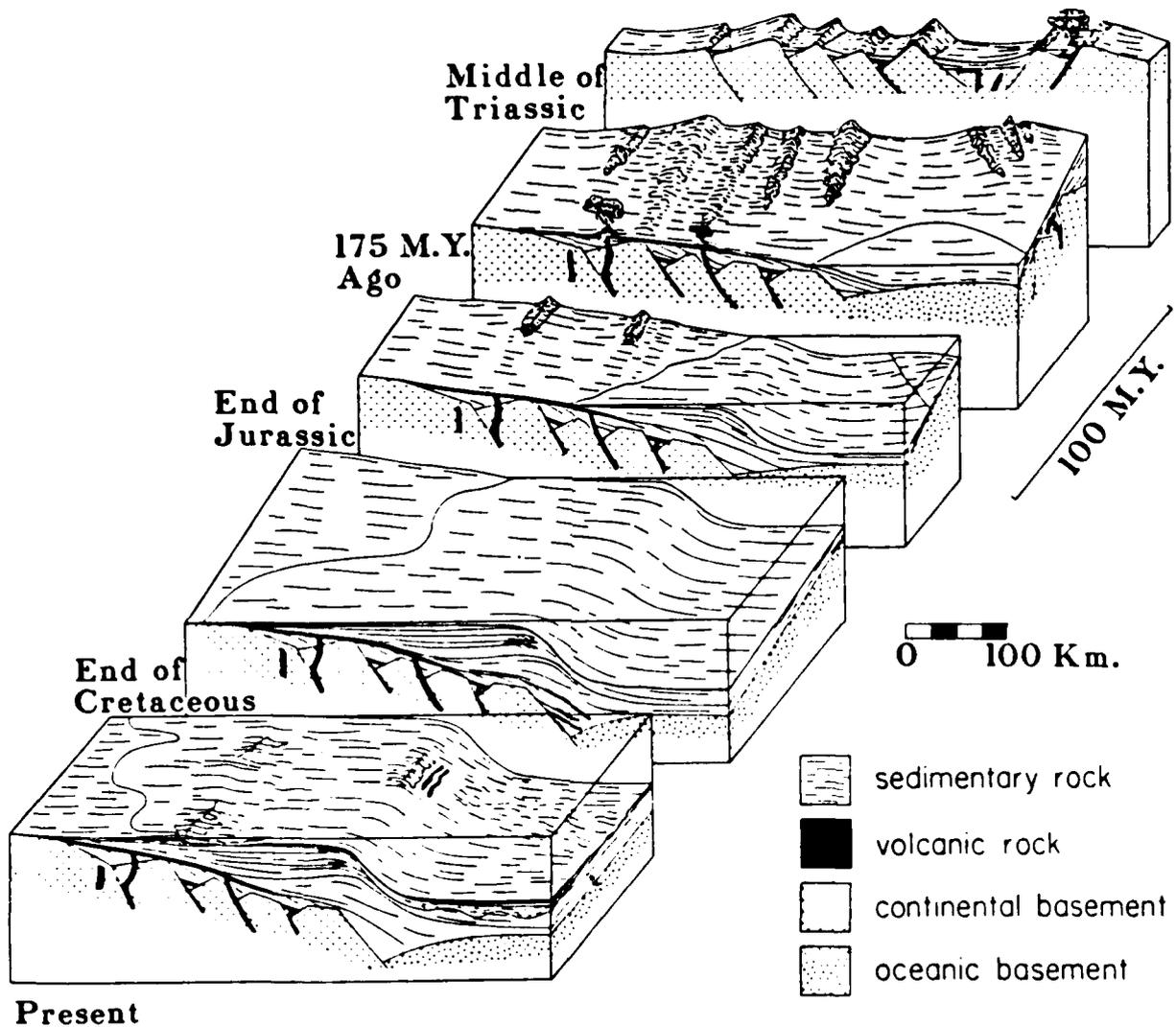
Figure 11. Block diagrams showing historic development of the continental
margin near Charleston, South Carolina (Dillon and others,
1979, Fig. 13)
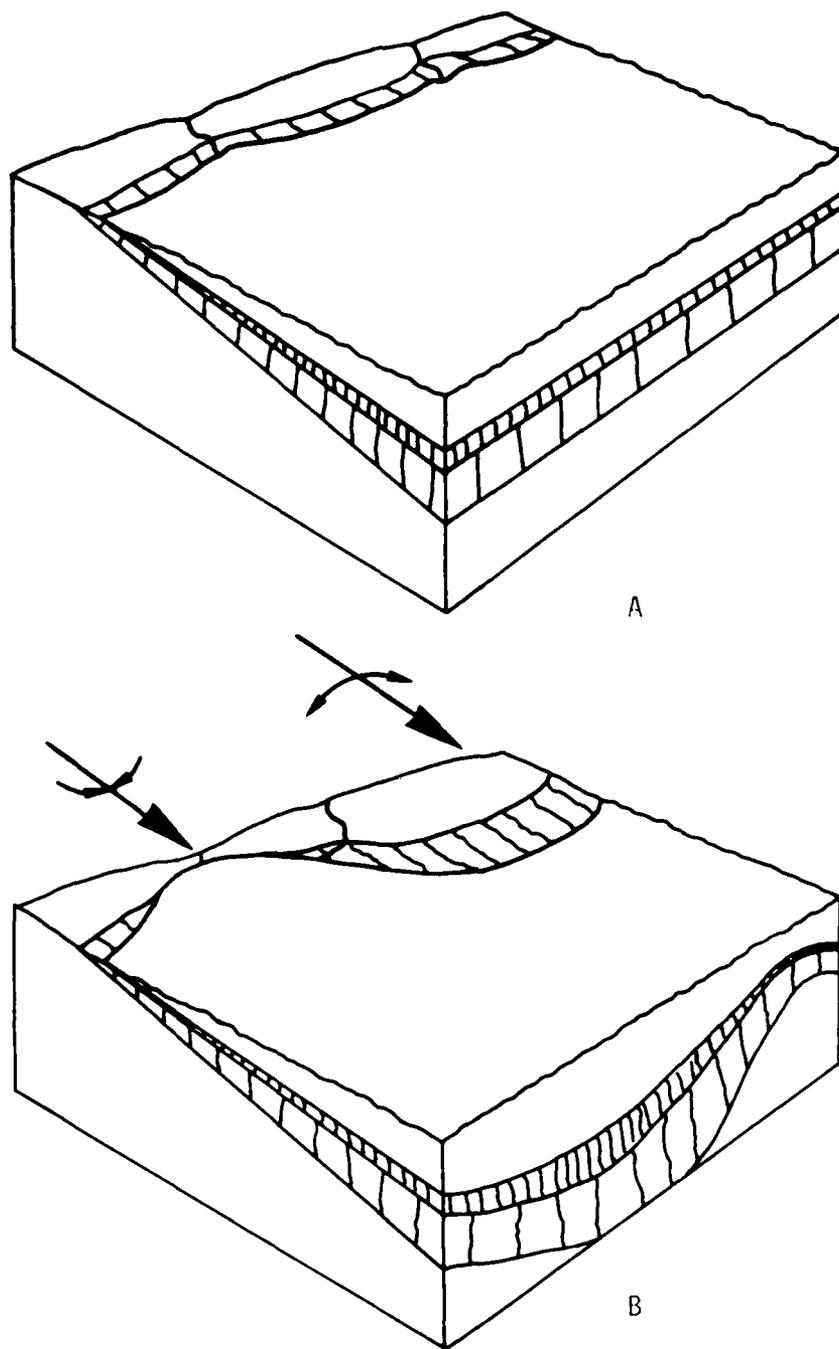
Figure 12. Block diagrams showing Coastal Plain with A) uniform sedimentation and B) variation due to transverse embayment and arch
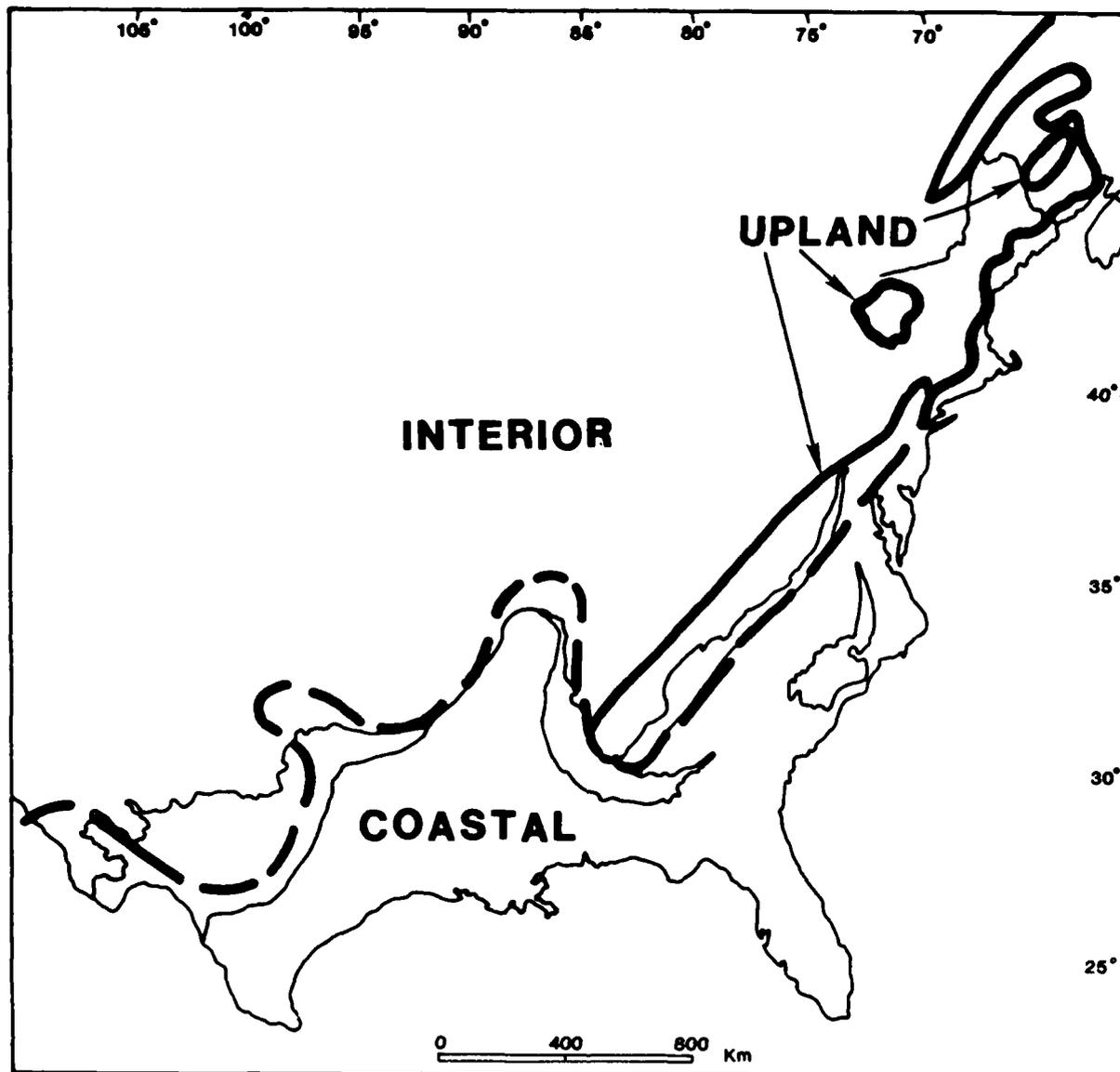
Figure 13. Map of the eastern United States and adjacent Canada showing approximate boundaries between areas of coastal, upland and interior zones of seismicity
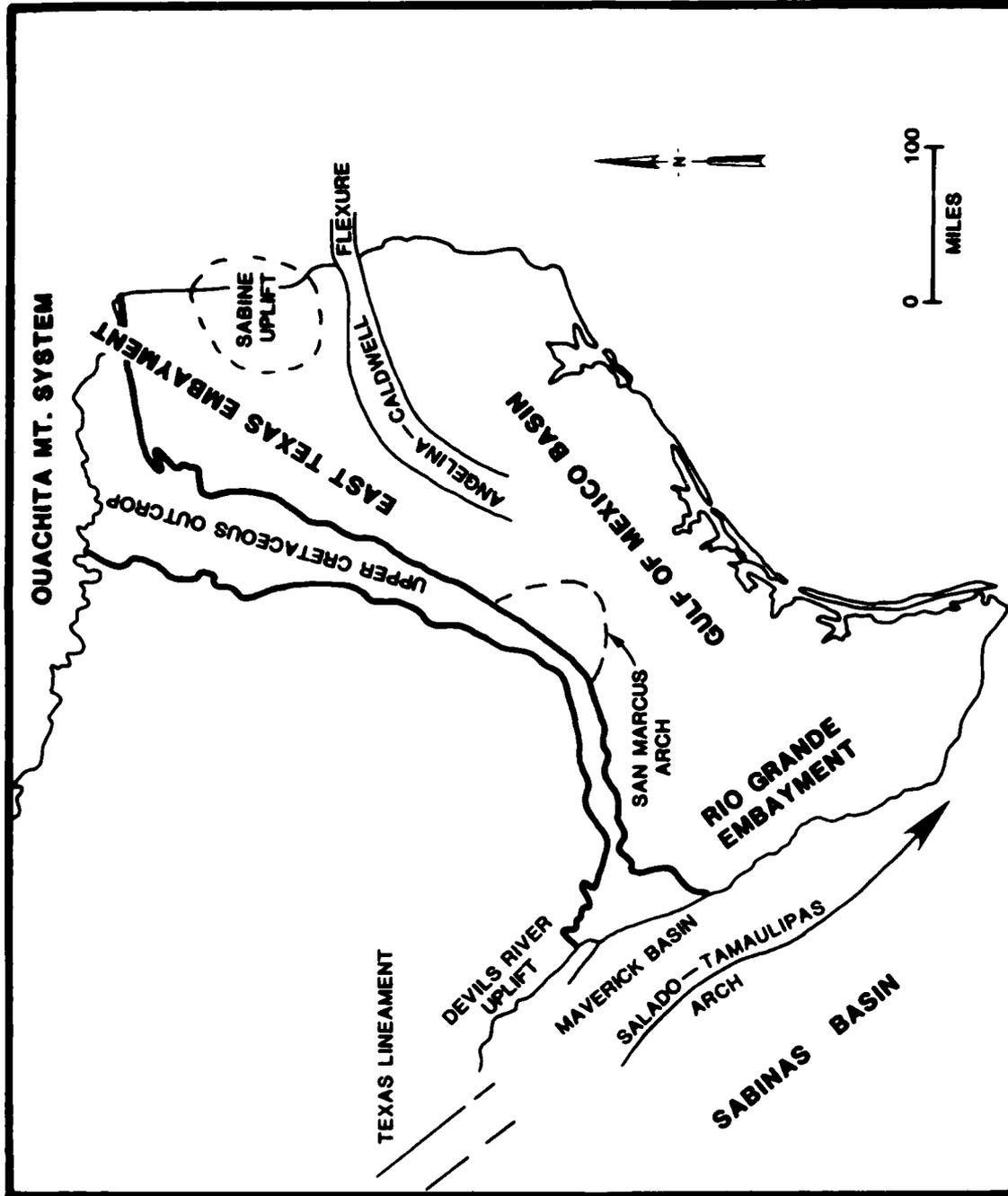
Figure 14. Map of Texas and adjacent area showing Late Cretaceous Embayments and selected geologic features (modified from Luttrell, 1977, Fig. 4)
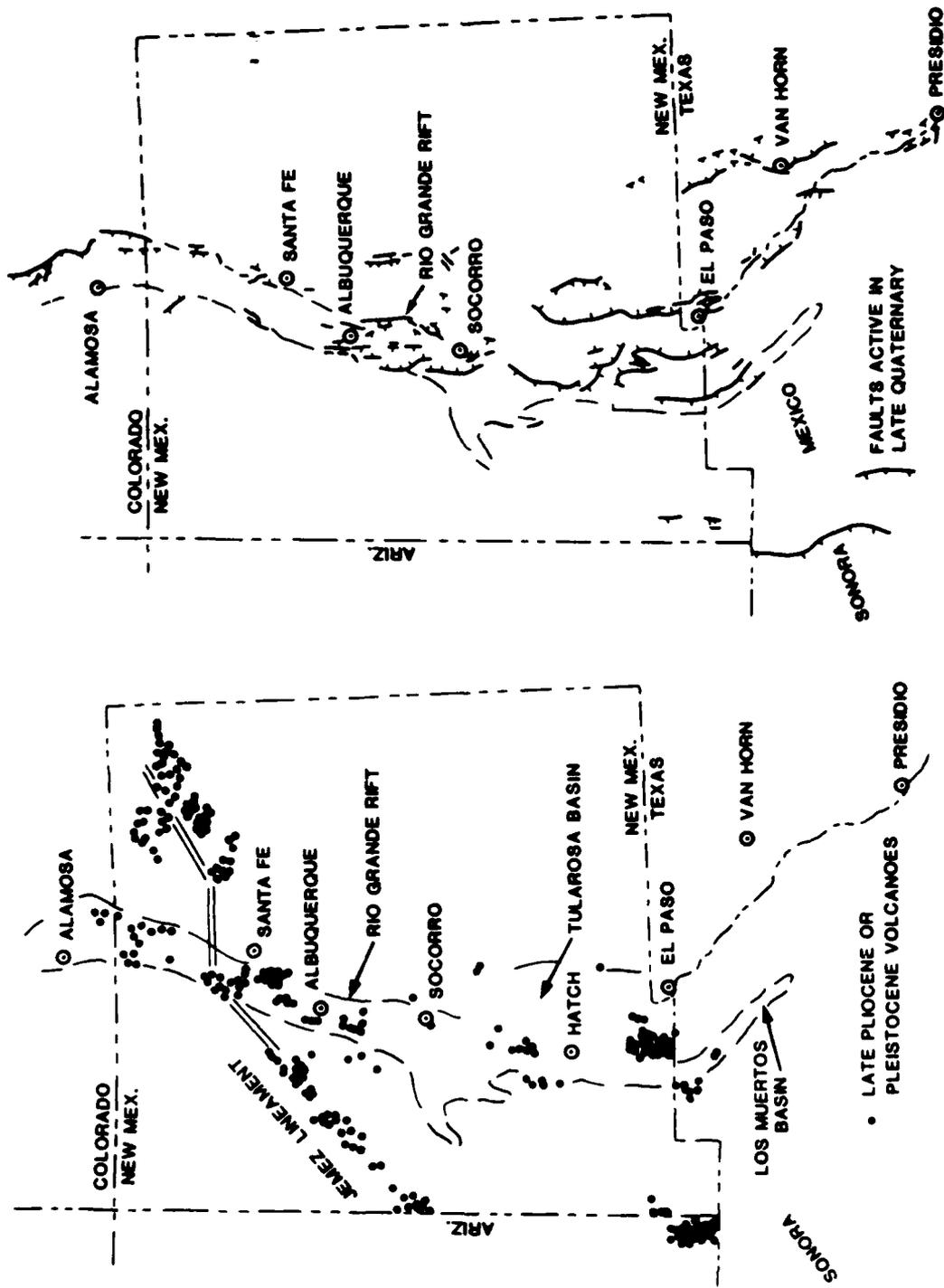
Figure 15. Map of New Mexico and adjacent area showing relation of late Quaternary faults and late Pliocene and Pleistocene volcanoes to Rio Grande Rift (modified from Seager and Morgan, 1979, Fig. 2)
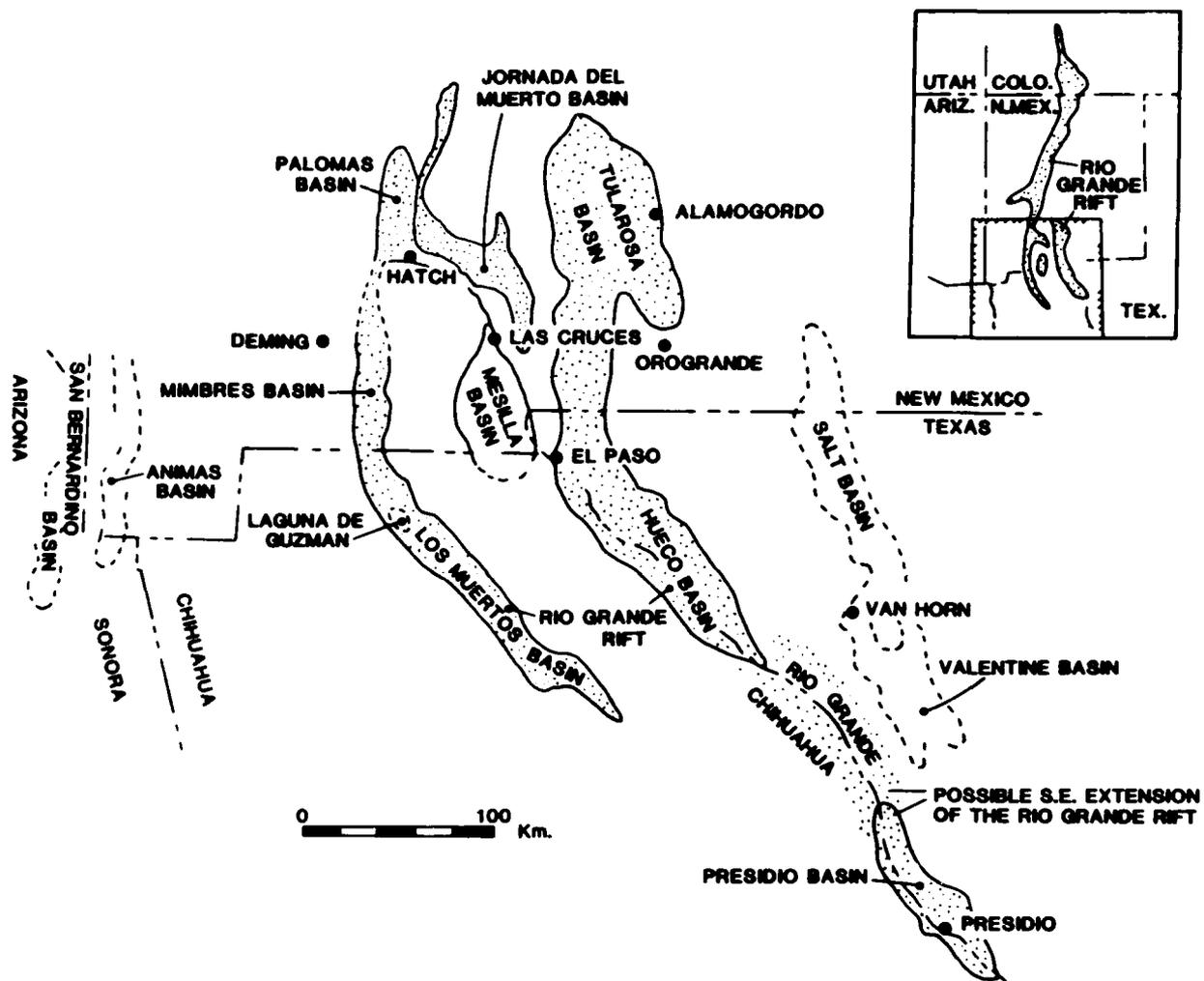
Figure 16. Index map of basins in the southern Rio Grande Rift and northwestern part of the Texas Lineament (Seager and Morgan, 1979, Fig. 1)
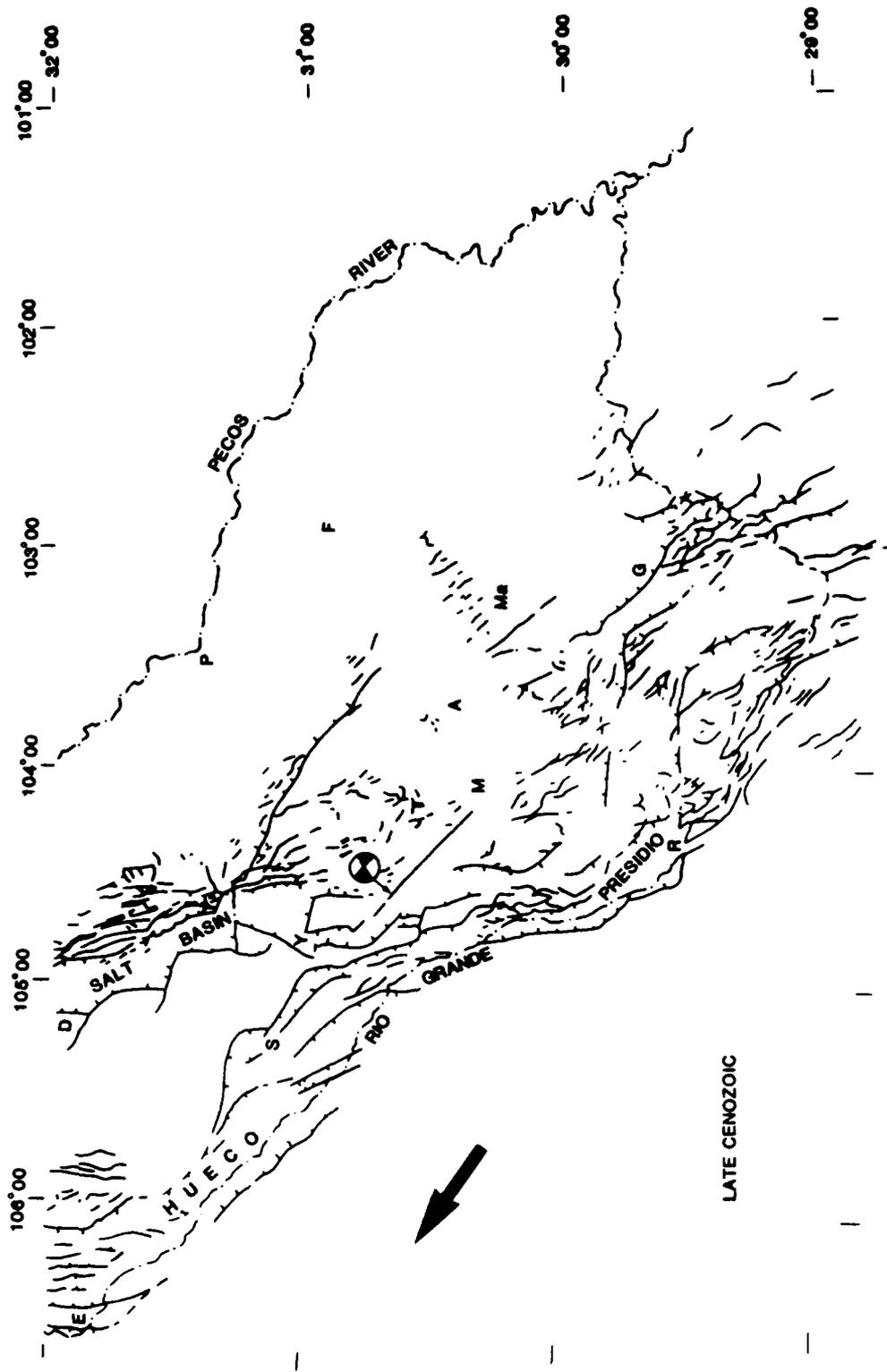
Figure 17. Map of West Texas showing known late Cenozoic faults. Tick marks on major graben border faults. Main sources: Geologic Atlas of Texas-Van Horn-El Paso, Pecos, Marfa, Ft. Stockton, and Emory Peak-Presidio sheets: King, 1937; Smith, 1970; Henry, 1979. Also shown is first-motion diagram from Dumas and others (1980) for Valentine earthquake; shaded quadrants=compression; arrow from diagram points to Valentine fault, the fault assumed to have moved during the August 16, 1931, Valentine earthquake (Muehlberger, 1980, Fig. 5)
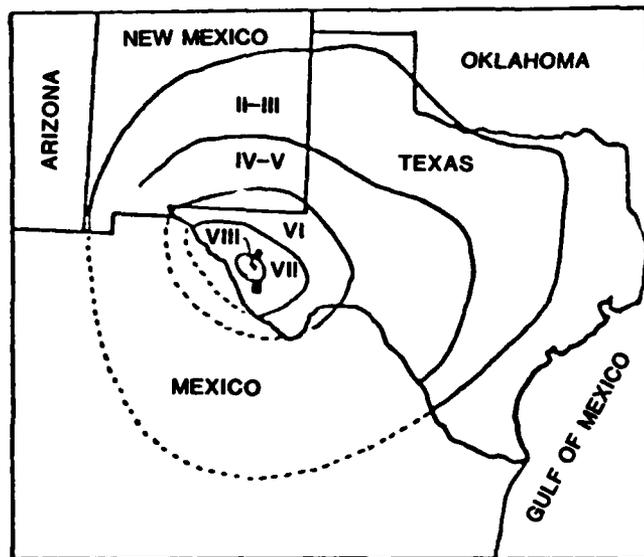
Figure 18. Isoseismal map for the August 16, 1931 Valentine, Texas, earthquake in the Rossi-Forel scale. Byerly's epicentral location indicated by (*) and the USGS location by triangles (Dumas and others, in press)
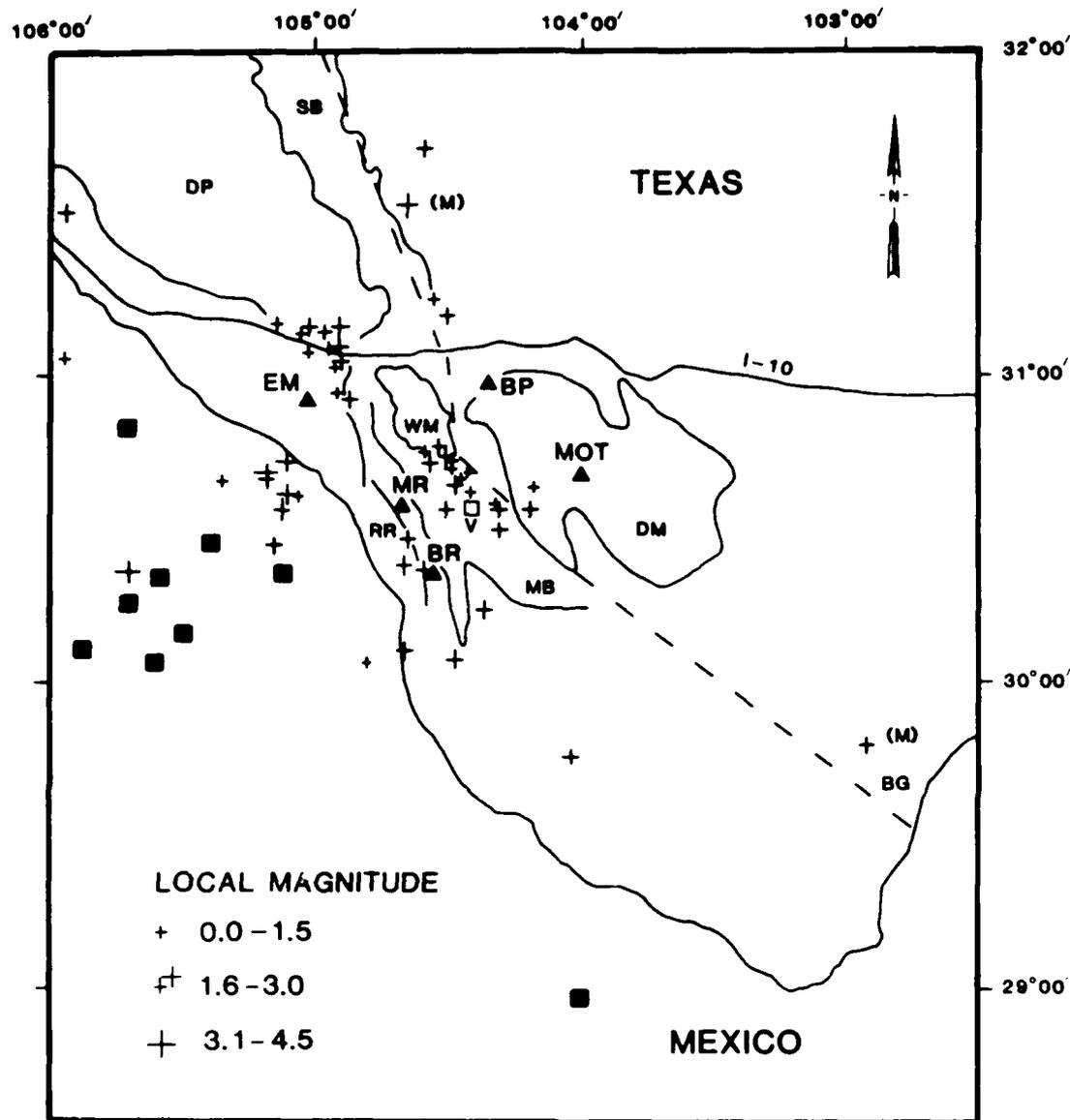
Figure 19. Seismicity map of the Basin and Range province of West Texas and adjacent Mexico. Crosses(+) indicate epicenters located by the five-station seismic array. The stations are indicated by triangles and locations and abbreviations are given in Table 1. Pre-1975 epicenters located by the USGS are indicated by solid squares. Abbreviations for structural features are: BG-Black Gap Area, DM-Davis Mountains, DP-Diablo Plateau, MB-Marfa Basin, RR-Rim Rock Fault, SB-Salt Basin Graben, and WM-Wylie Mountains. The dashed line marks Muehlberger's (1979) proposed eastern boundary of Basin and Range faulting. Multiple earthquakes with the same epicenter are indicated by (M). The town of Valentine (V) is indicated by the small open square (Dumas, 1980, Fig. 1)
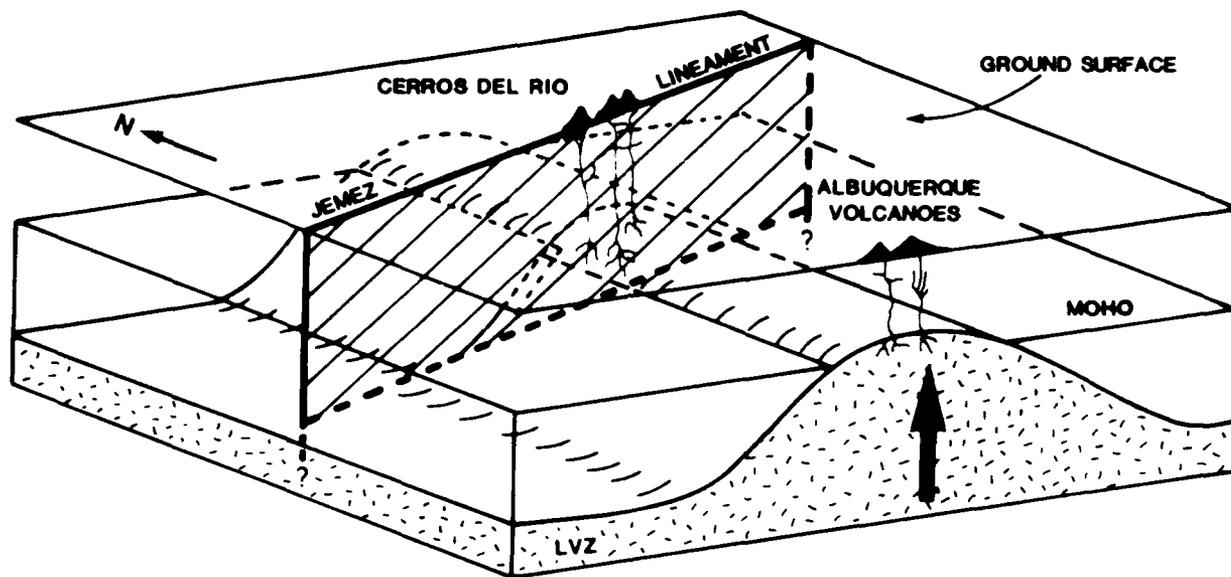
Figure 20. Schematic diagram of central Rio Grande Rift showing crustal thinning over the low velocity zone (LVZ) and volcanic fields, which are located along transverse offset of rift (shown here as Jemez Lineament) (Baldridge, 1979, Fig. 14)
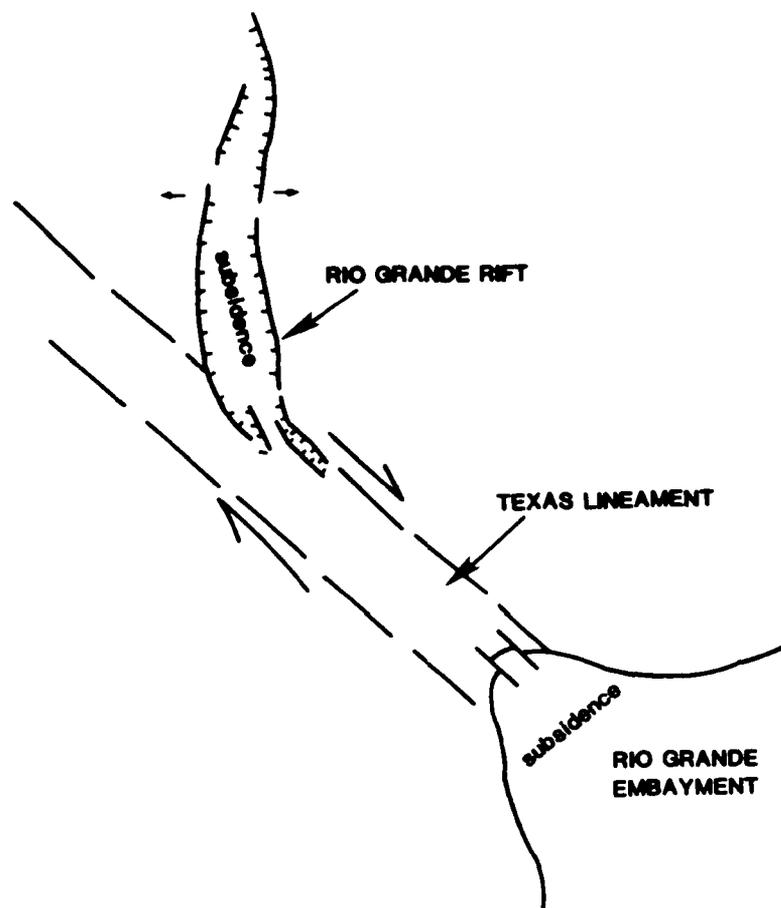
Figure 21. Sketch map showing relations of the Rio Grande Embayment, Texas Lineament and Rio Grande Rift and their probable relative movements
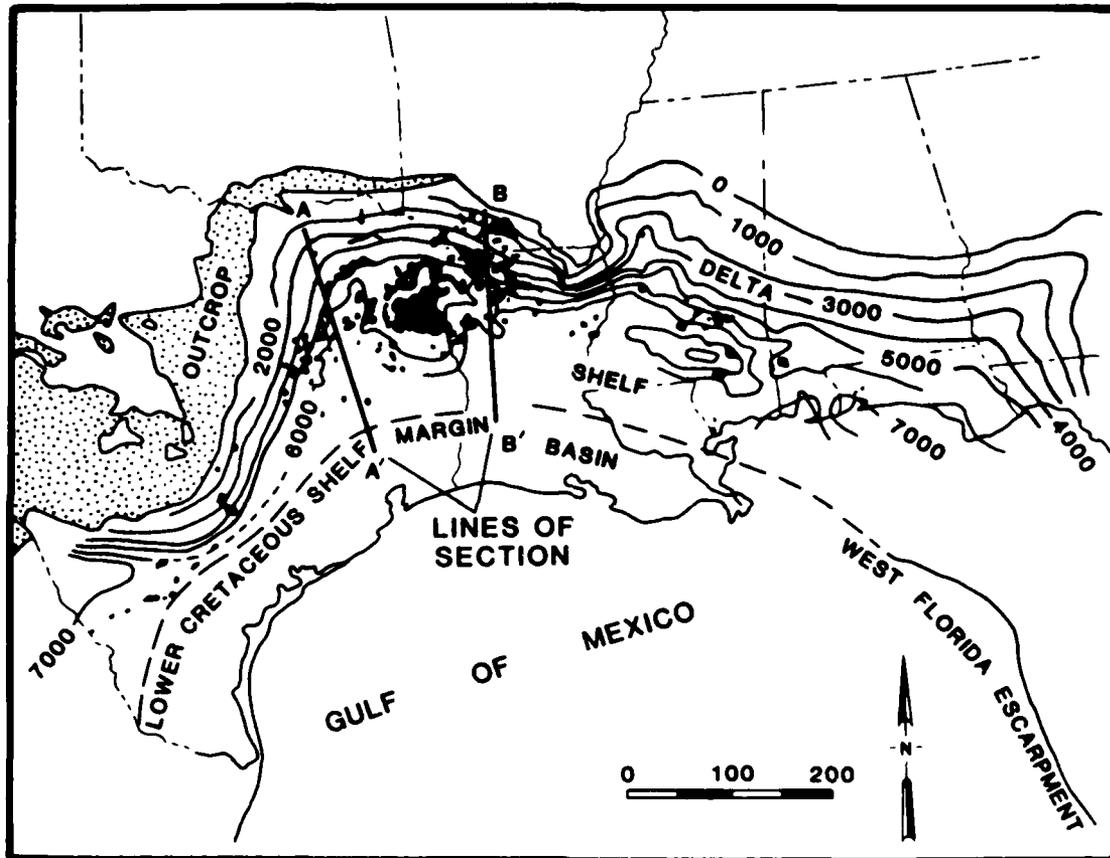
Figure 22. Map of the US Gulf Coast showing Lower Cretaceous thickness patterns (Rainwater, 1971)

Figure 23. Map showing basement structure of the Mid-Continent region (Rascoe and Adler, 1983, Fig. 1)
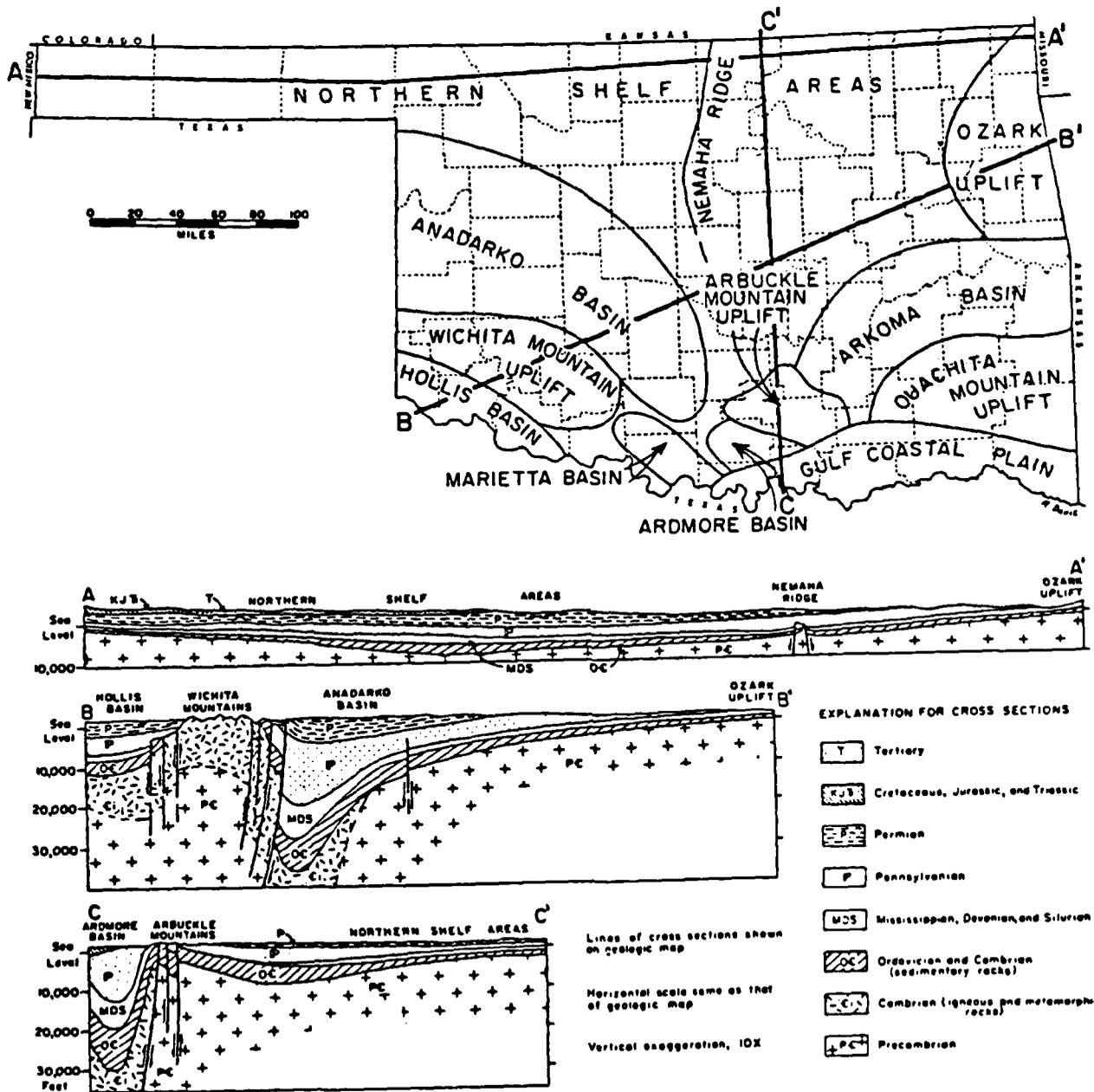
Figure 24. Map showing geologic provinces of Oklahoma and cross sections
(Oklahoma Geological Survey)
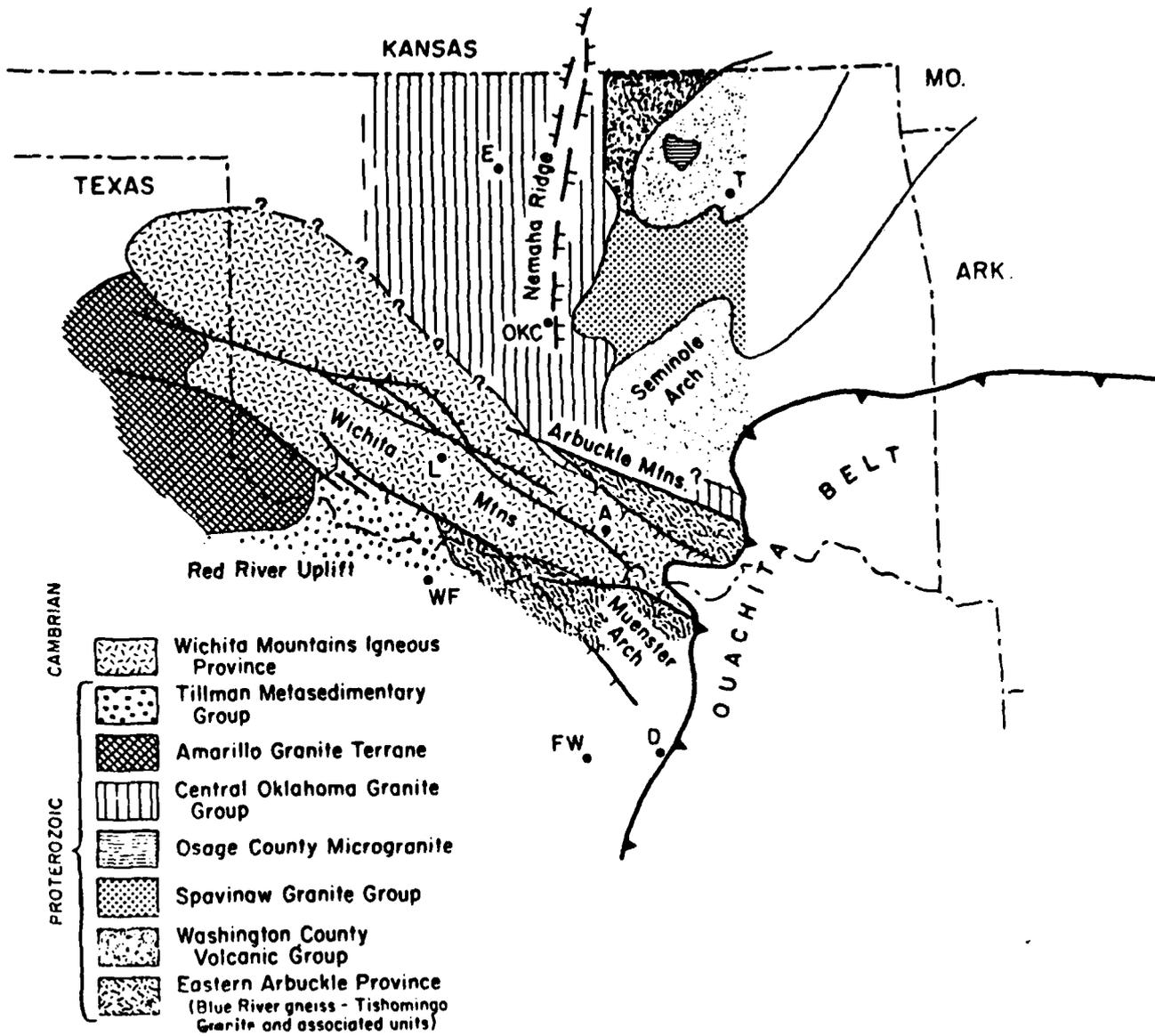
Figure 25. Map of the Southern Oklahoma Aulacogen region showing basement geology (Gilbert, 1982, Fig. 4)

SOUTH   NORTH

|← ~250 km →|

HOLLIS   BURCH   WICHITA   FRONTAL   ANADARKO
BASIN   FAULT   UPLIFT   FAULTS   BASIN

OM   OM
WM   WM

EARLY PERMIAN

No known volcanism

LATE MISSISSIPPIAN

≠ II

MID. CAMBRIAN

Basaltic ≠ 4 (mostly hidden)
Rhyolitic ≠ II (A-type)

≠ 2

MID CAMBRIAN? to MID PROTEROZOIC?

Basaltic ≠ 3 (wet tholeiitic)
[Basaltic ≠ 2 (spilitic)]
[Basaltic ≠ 1 (tholeiitic)]

CENTRAL OKLA
GRANITE
PLATFORM
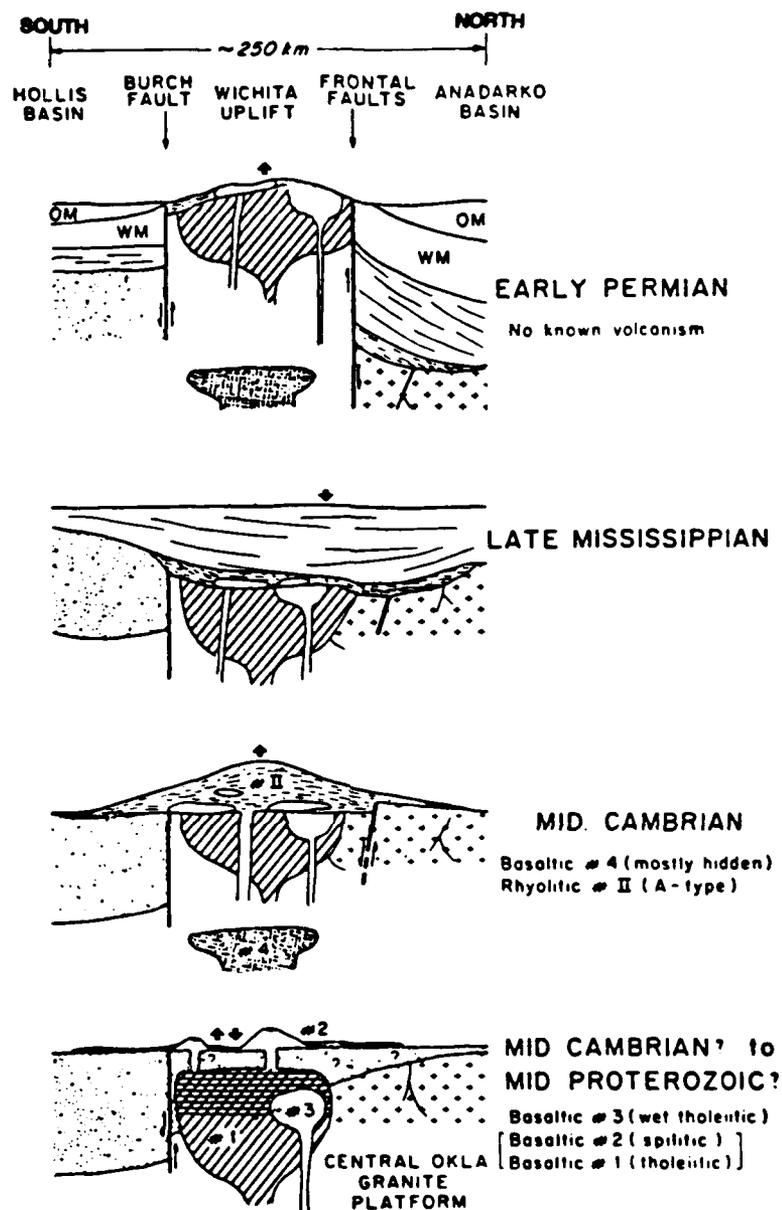
Figure 26.   Diagrams showing geologic development of the Wichita Mountains crustal block. View across axis of uplift. Most faults shown at high angle, although the frontal faults are known to dip 30°-40° to south (see Brewer, this guidebook). Strike-slip movement not precluded. Rhyolitic I = shallow older basement granites, perhaps equivalent to eastern Arbuckle province. Basaltic 1 = Glen Mountains Layered Complex. Basaltic 2 = Navajoe Mountain Basalt-Spilite Group. Basaltic 3 = Roosevelt Gabbros. Rhyolitic II = Carlton Rhyolite Group and Wichita Granite Group. Basaltic 4 = late diabase. Short, heavy arrows indicate net motion of land surface, by subsidence, uplift, or erosion. (Gilbert, 1982, Fig. 5)
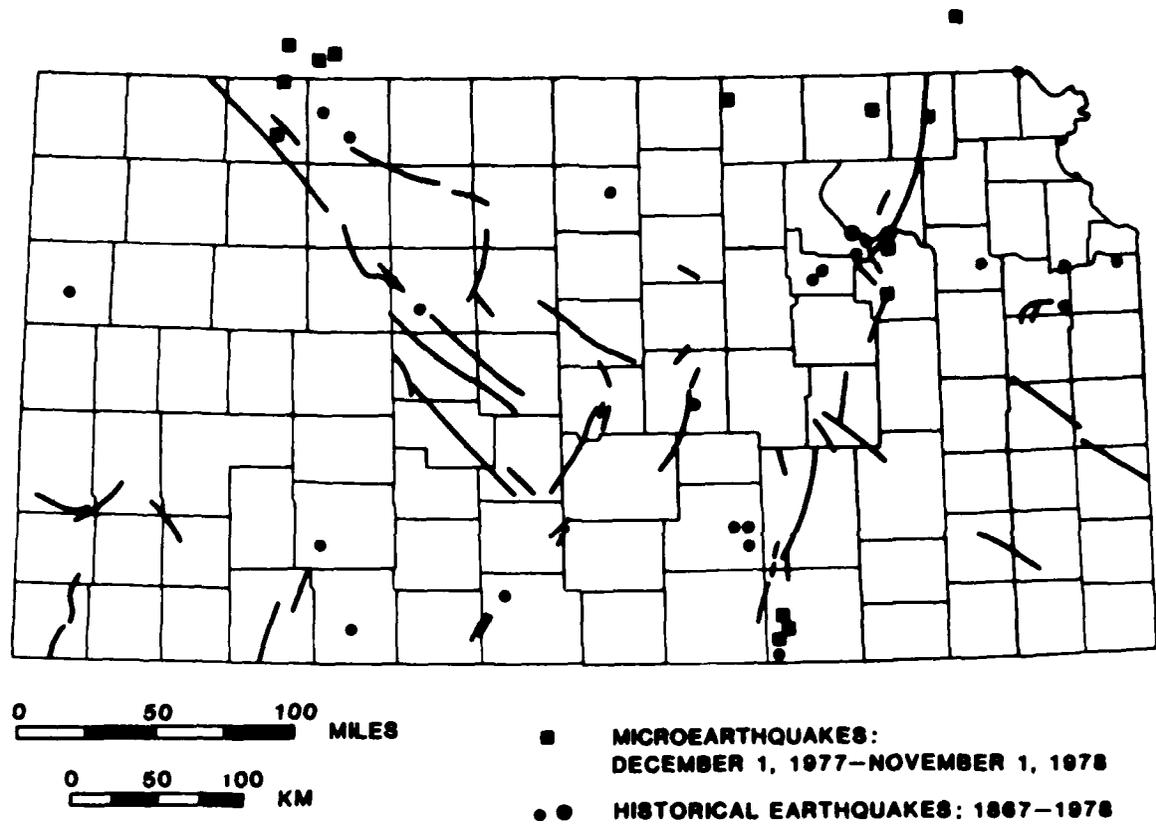
Figure 27. Map of Kansas showing known surface and subsurface faults and locations of historic earthquakes and recent microearthquakes (Wilson, 1979, Fig. 23)
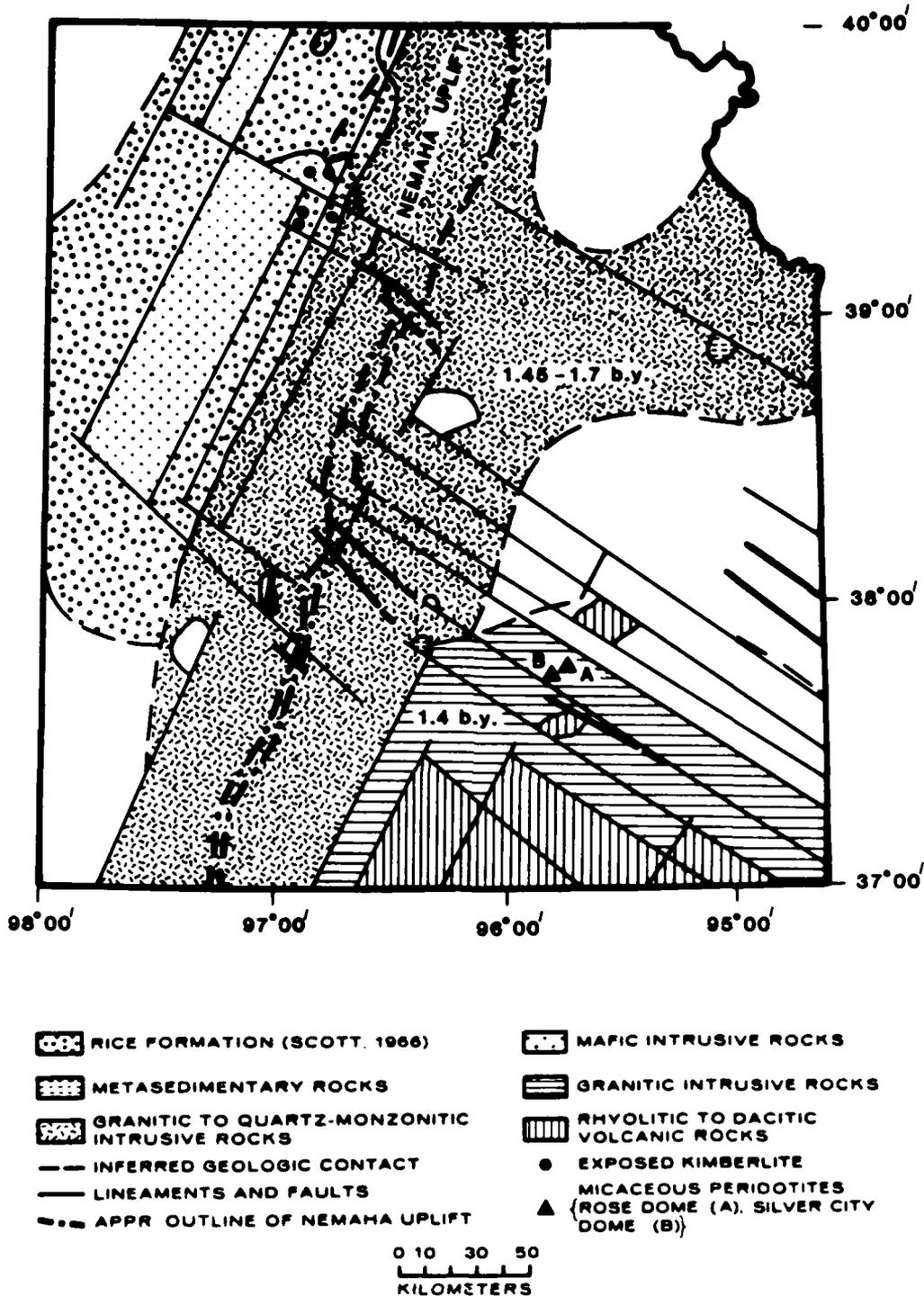
Figure 28. Map of eastern Kansas showing geologic interpretation of the Precambrian basement rocks. Blank spaces represent areas for which no information regarding basement rock types is available (Berendsen and others, 1981, Fig. 9)
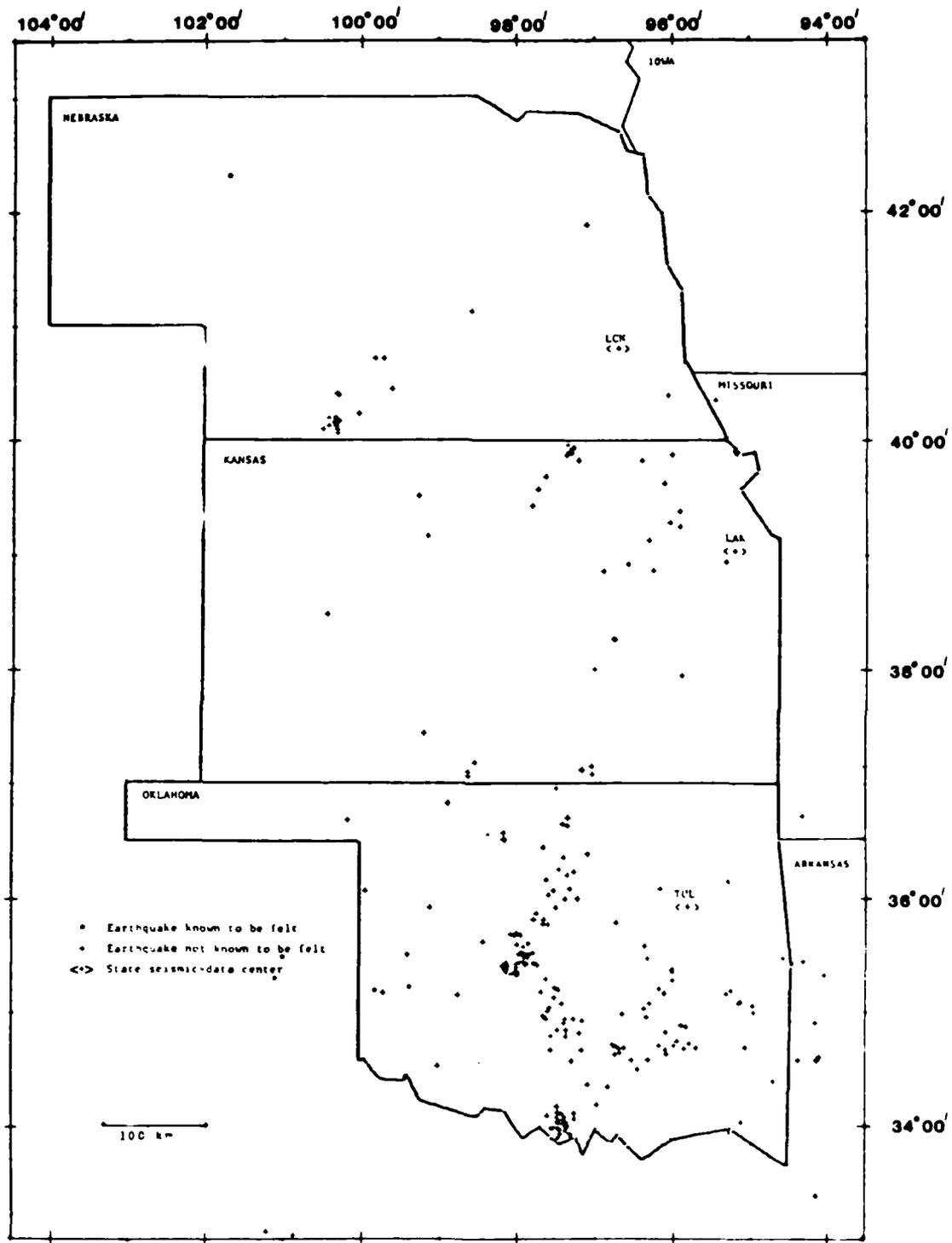
Figure 29. Map showing earthquake epicenters for western Iowa, Nebraska, Kansas, Oklahoma and surrounding area; January 1, 1977, through December 31, 1980 (Luza and Lawson, 1982, Fig. 10)
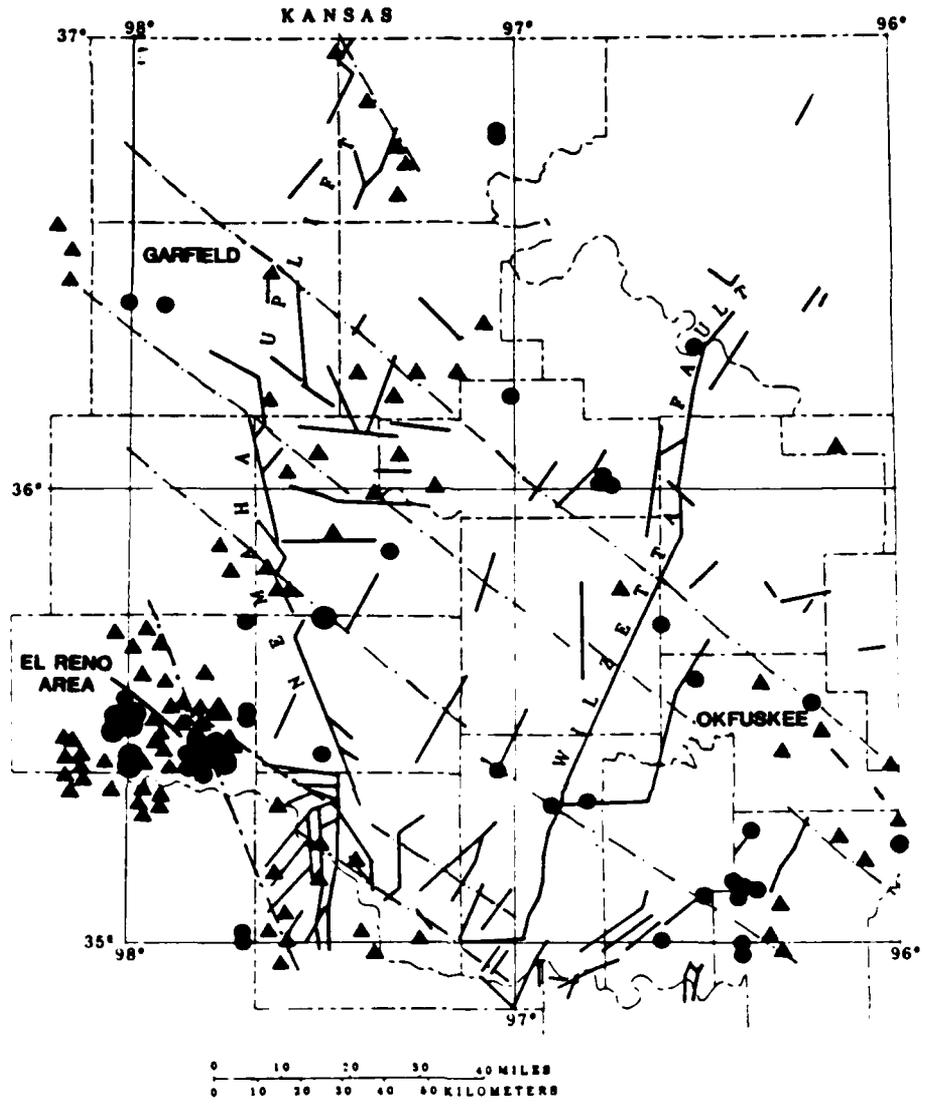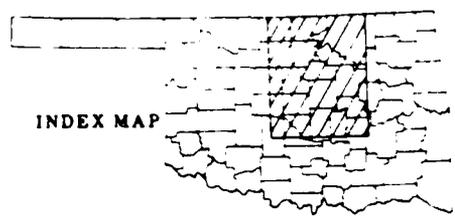
Figure 30. Map of north-central Oklahoma showing distribution of faults that cut pre-Pennsylvanian strata, earthquake epicenters and selected magnetic lineaments (modified from Luza and Lawson, 1982, Fig. 11)
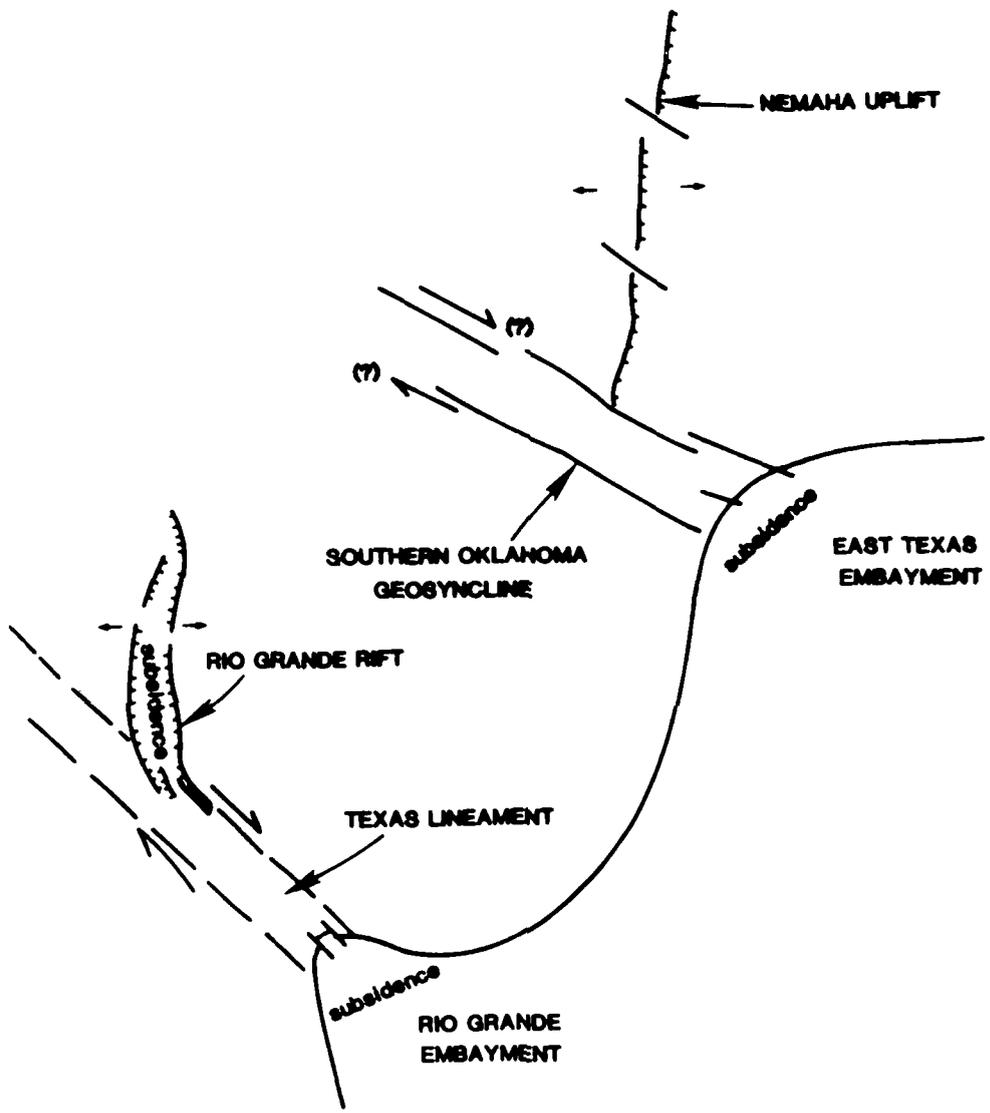
Figure 31. Sketch map showing Rio Grande and East Texas Embayments and major structures in areas of earthquakes

NORTHWEST

SOUTHEAST

MISSOURI | TENNESSEE

OZARK
PLATEAU

MISSISSIPPI

EMBAYMENT

HIGHLAND
RIM

SCALE
(meters)

CROWLEYS RIDGE

MISS. RIVER

MEMPHIS

RECENT AND
PLIO-PLEISTOCENE 0-10 M.Y.

300

MIDDLE AND UPPER EOCENE 40-50 M.Y.

SEA LEVEL

LOWER EOCENE 50-60 M.Y.

PALEOCENE 60-70 M.Y.

CRETACEOUS 70-100 M.Y.

300

PALEOZOIC

ROCK

300-600 M.Y.

600

0   40   80 KM

900
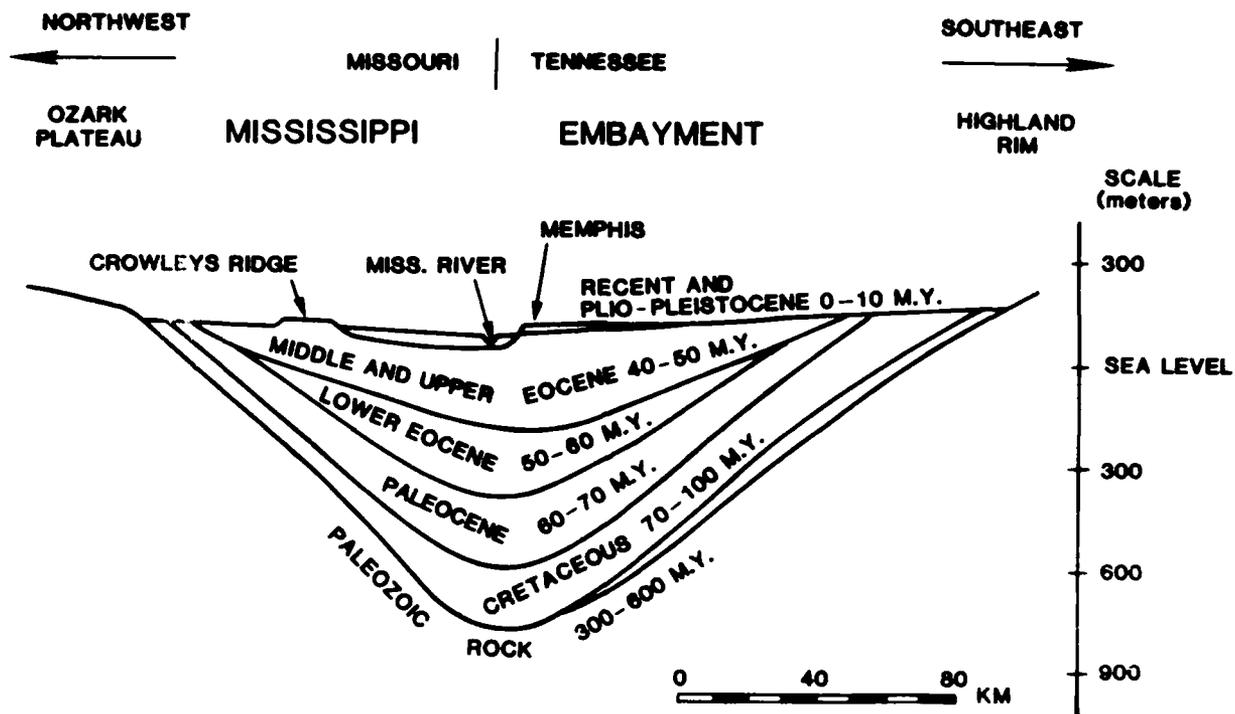
Figure 32. Cross section of the Mississippi Embayments showing
post-Paleozoic deposits (Nowak and Berg, 1981, Fig. 3)
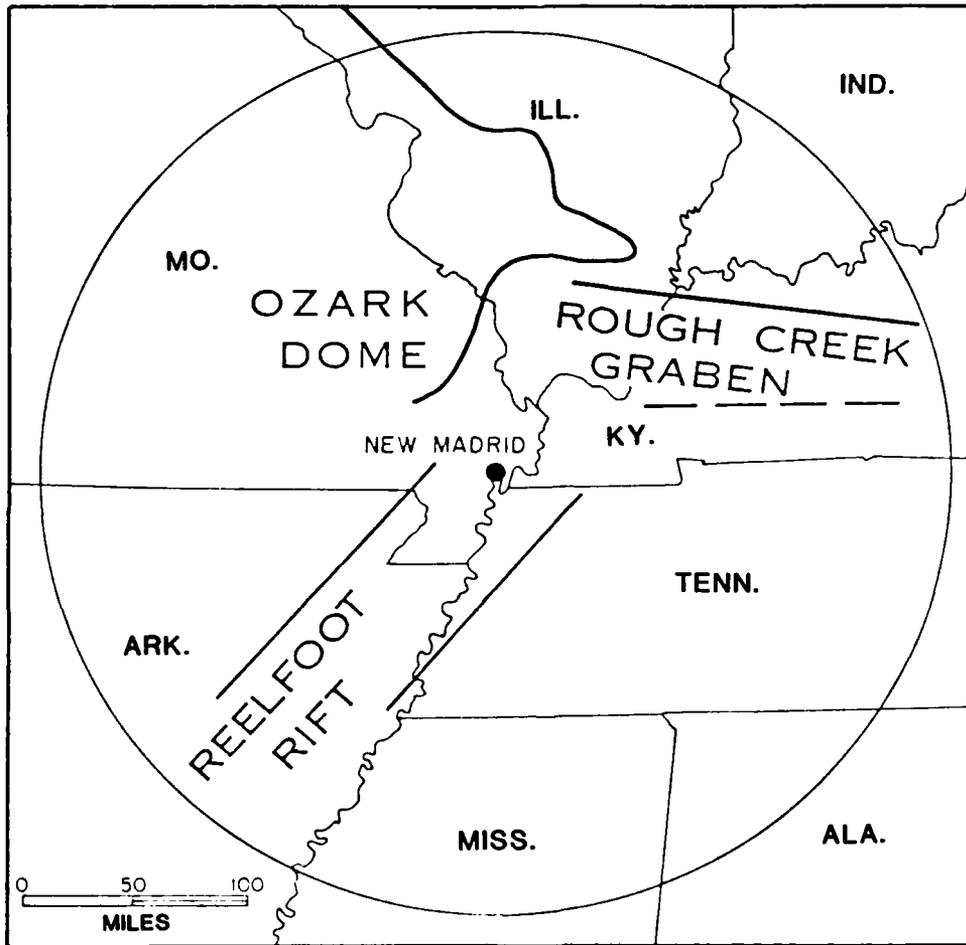
Figure 33. Sketch map showing tectonic features active in the New Madrid area in the late Precambrian and Cambrian time (Buschbach, 1980, Fig. A-2)
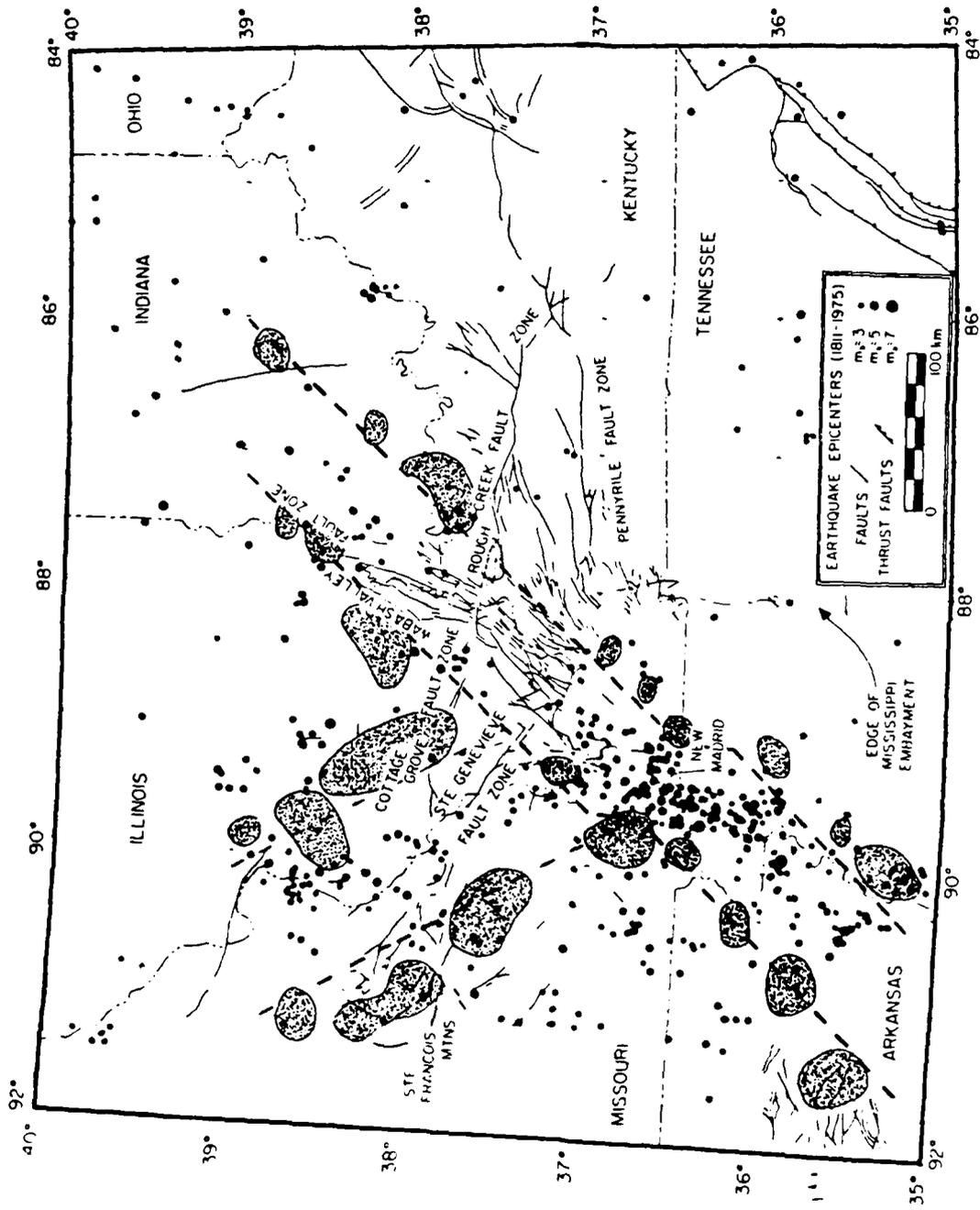
Figure 34. Map of the northern Mississippi Embayment and adjacent area showing earthquake epicenters, mapped faults, inferred mafic intrusions (stippled pattern) and boundaries of linear basement structures (heavy slashed lines) (Braile and others, 1981, Fig. 53)
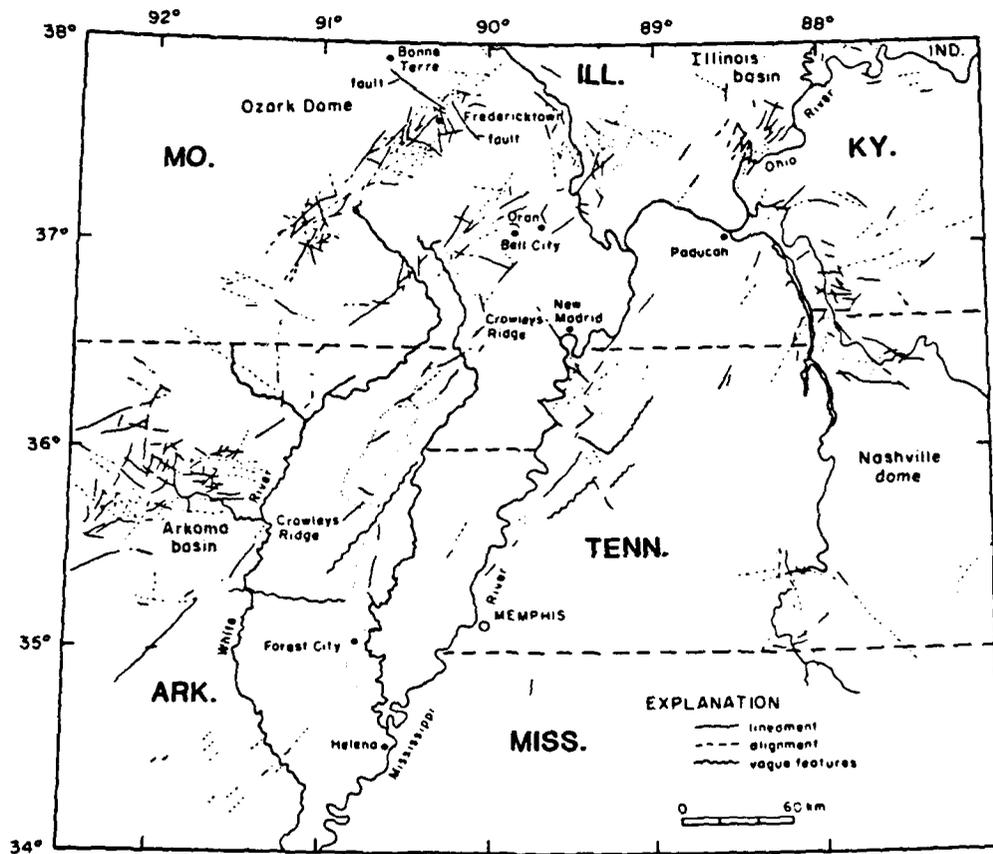
Figure 35. Map of the Mississippi Embayment and adjacent region showing linear features found from analysis of SLAS image strips (O'Leary and Hildenbrand, 1981, Fig. 1)
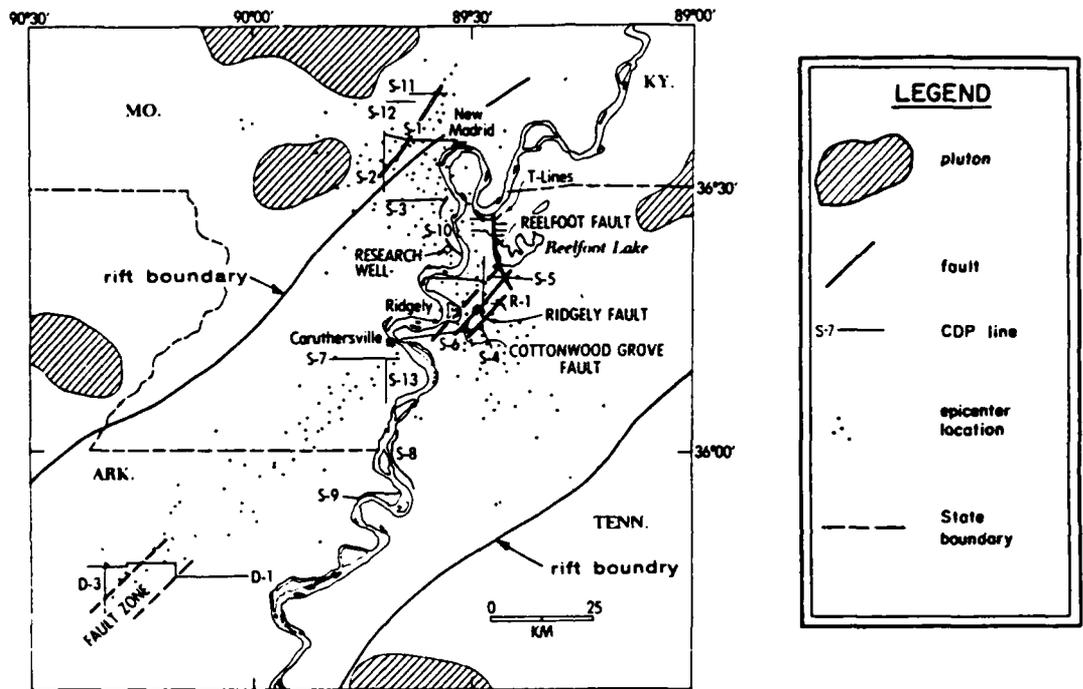
Figure 36. Map of the northern Mississippi Embayment region showing boundaries of the Reelfoot Rift and plutons in the basement, probable Holocene faults, earthquake epicenters and positions of seismic reflection (CDP) lines (Russ, 1981, Fig. 5)
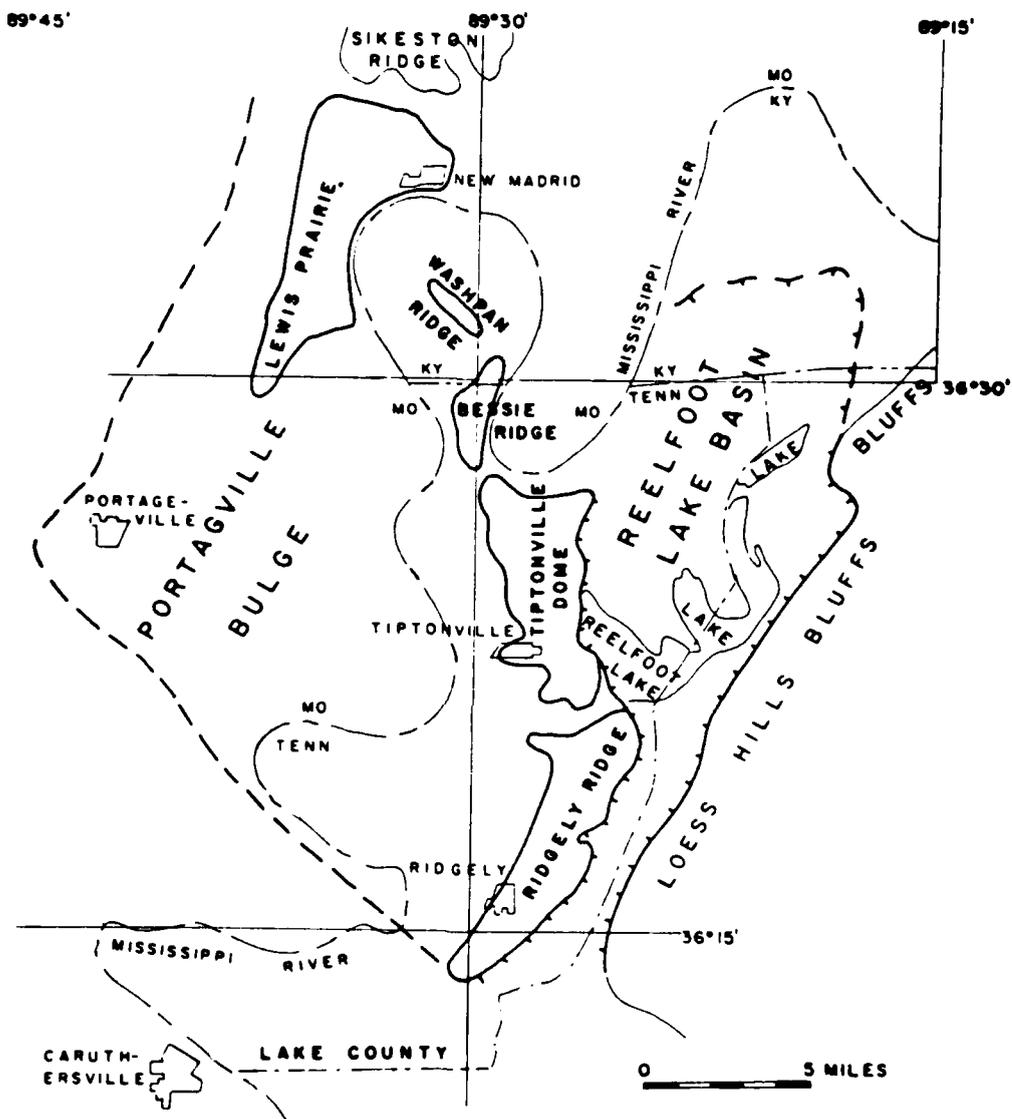
Figure 37. Map of the head of the Mississippi Embayment showing topographic features considered to have or probable have a tectonic origin (Stearns, 1979, Fig. 3)
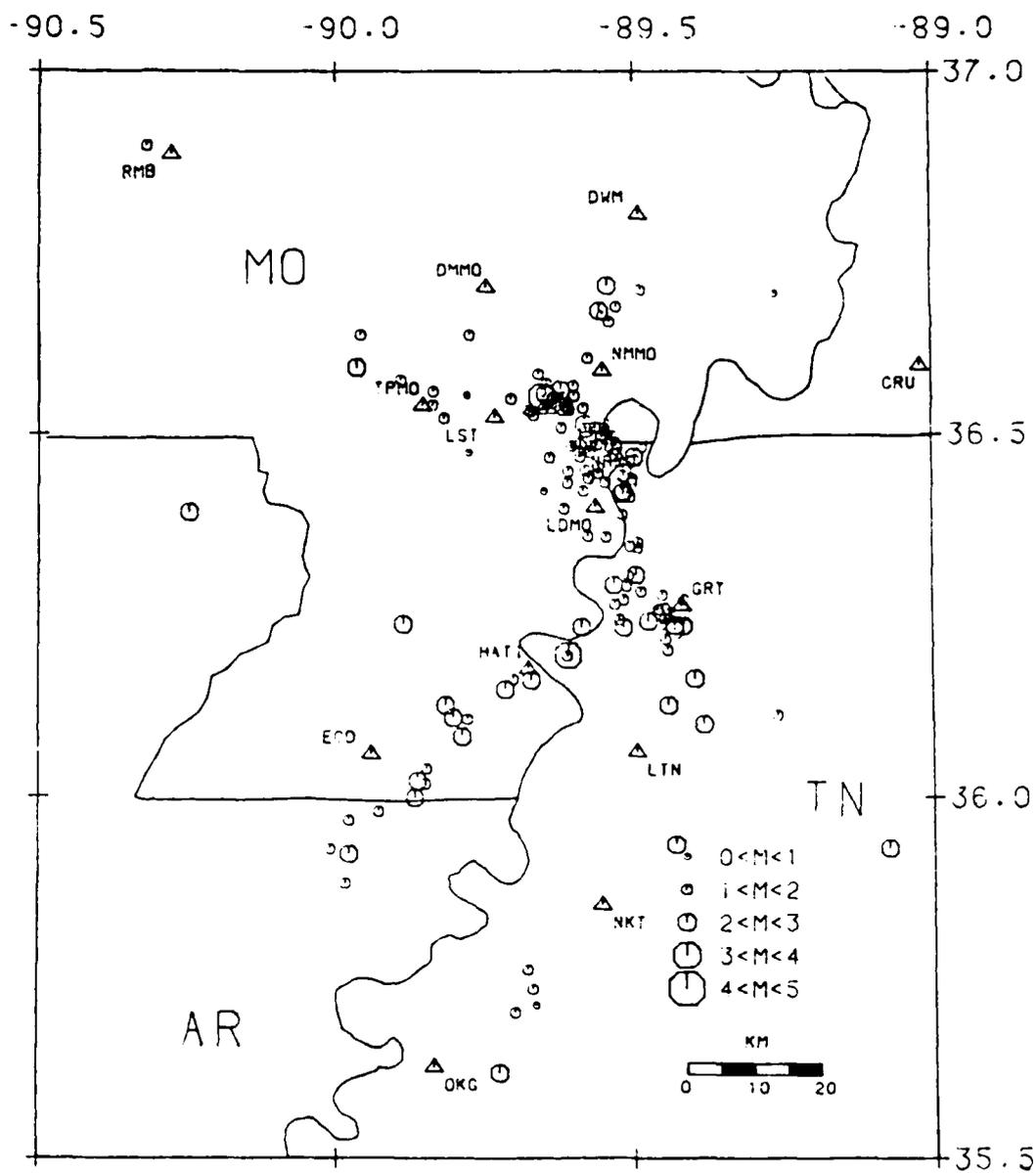
Figure 38. Epicentral map of the region surrounding New Madrid, Missouri, showing earthquakes from April 1982 through March 1983 (Stauder and others, 1983, Fig. 3)
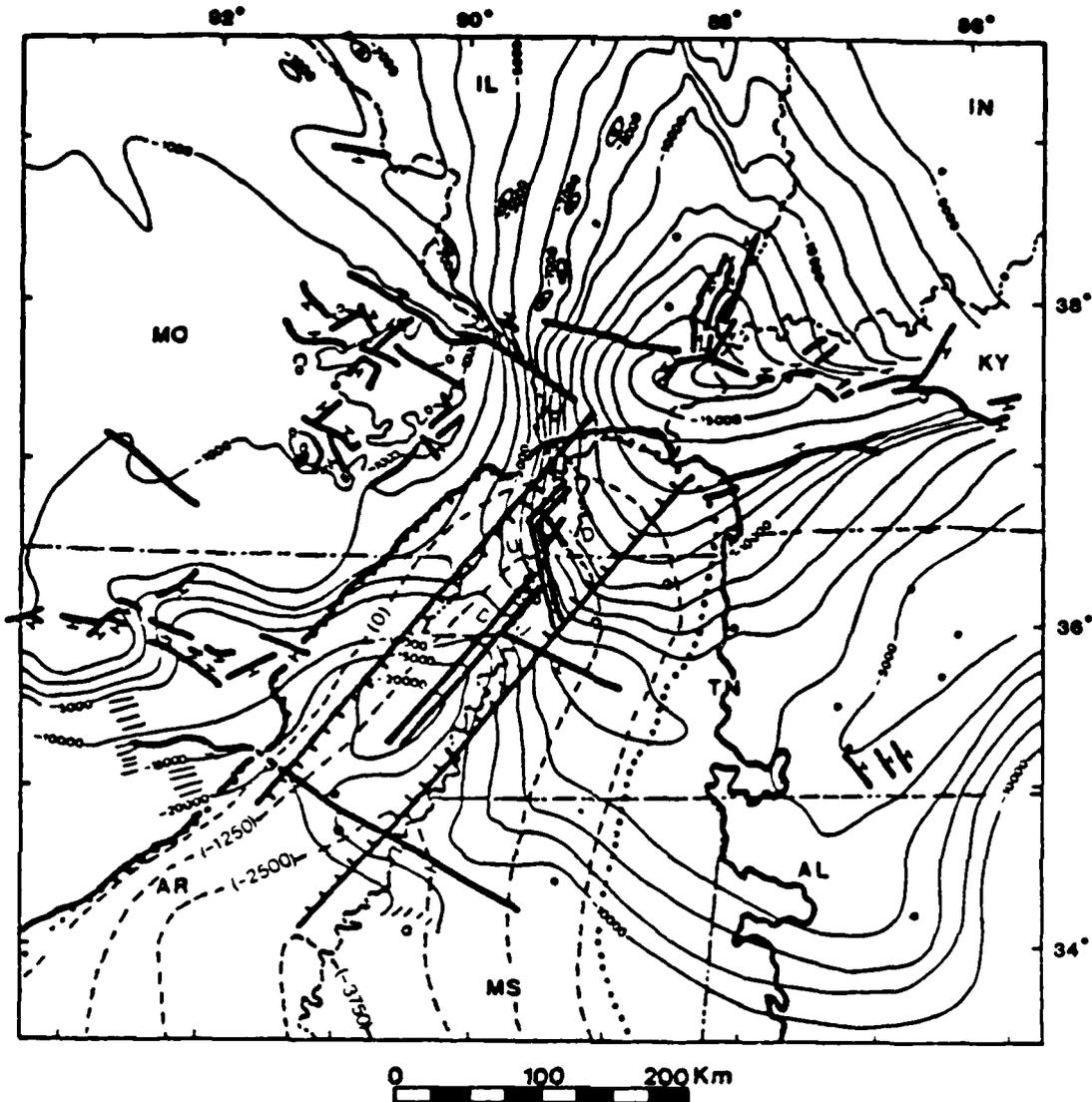
EXPLANATION

• Borehole reaching Precambrian igneous rocks.

○ Other significant deep borehole.

⎯-2000⎯ Structural contours drawn on top of Precambrian igneous rocks, showing elevation below sea level.

Border of Coastal Plain rocks of the Mississippi embayment.

·········· Contact of Tertiary rocks.

--12501-- Structural contour drawn on top of Cretaceous rocks showing elevation below sea level.

Generalized border of graben in Precambrian rocks interpreted from geophysical data, ticks indicate downthrown side.

Trend of seismicity; letters indicate relative movement (D, down; U, up) based on slope of Precambrian surface, if seismicity occurs on faults.

X New Madrid, Missouri, approximate epicenter of 1811-1812 earthquakes.

Figure 39. Map of the New Madrid seismic area showing the configuration of the top of Precambrian rocks within a 200 mile radius of New Madrid (Buschbach, 1980), and the base of Tertiary rocks (Cohee, 1961), borders of the Precambrian Reelfoot Rift (Pinckney, 1980), and the presently active seismic zone (Herrmann, 1980)
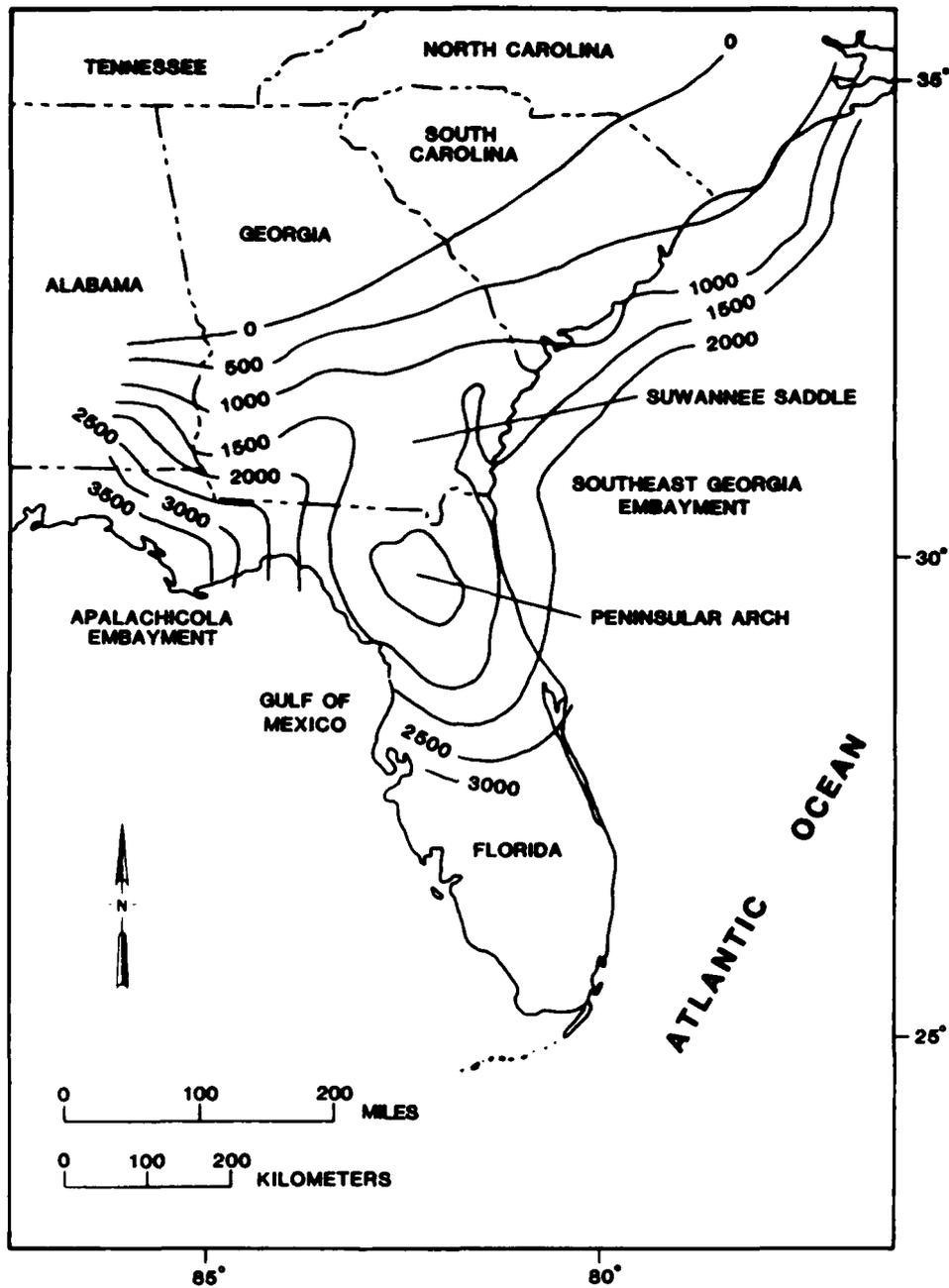
Figure 40. Map of the southeastern United States showing contours of the unconformity beneath the Coastal Plain deposits in depth below sea level in meters (Chowns and Williams, 1983, Fig. 1)
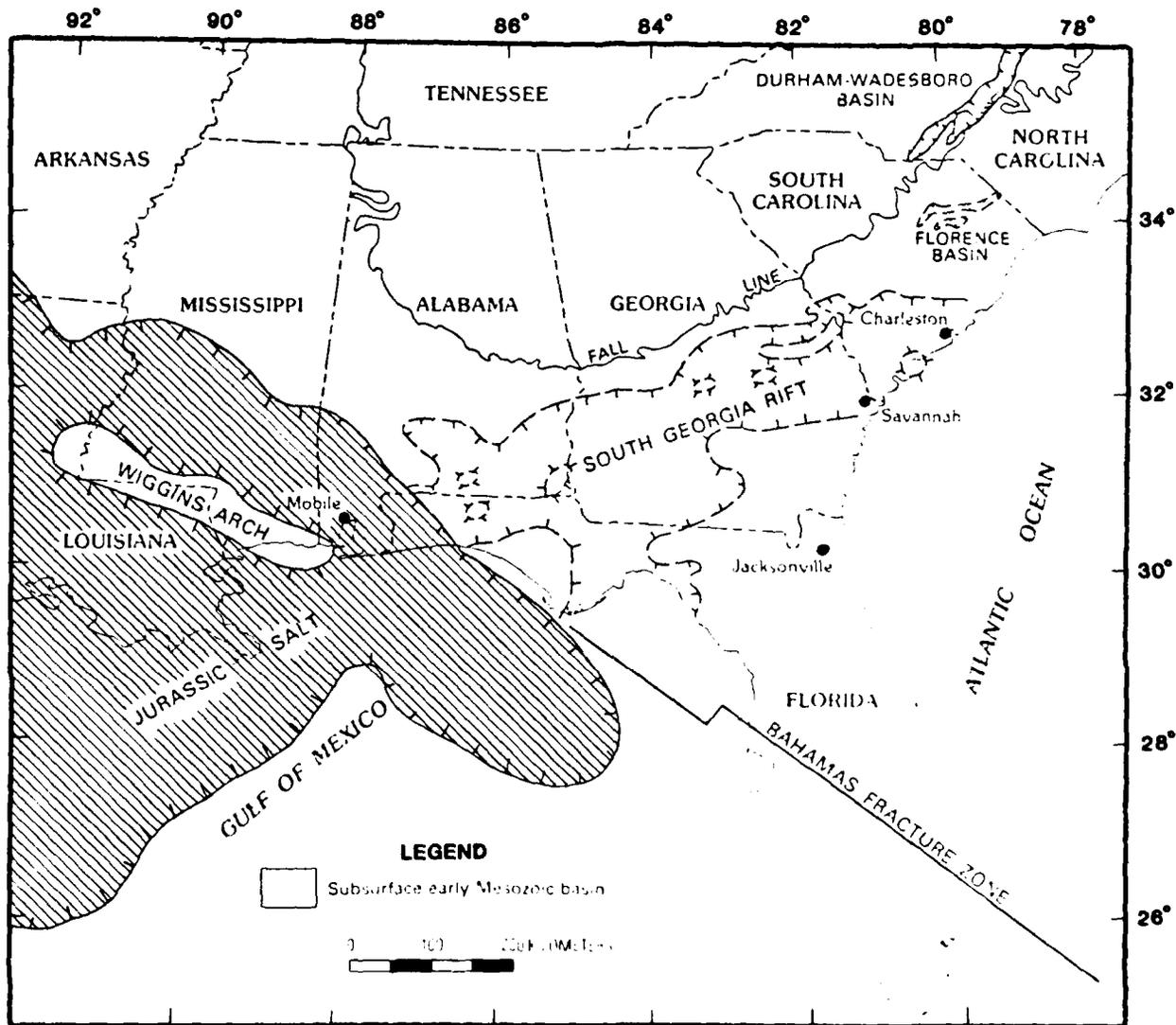
Figure 41. Map of southeastern United States showing interpreted distribution of subsurface early Mesozoic basins, extent of Jurassic salt and the Bahamas fracture zone (Daniels and others, 1983, Fig. 2)
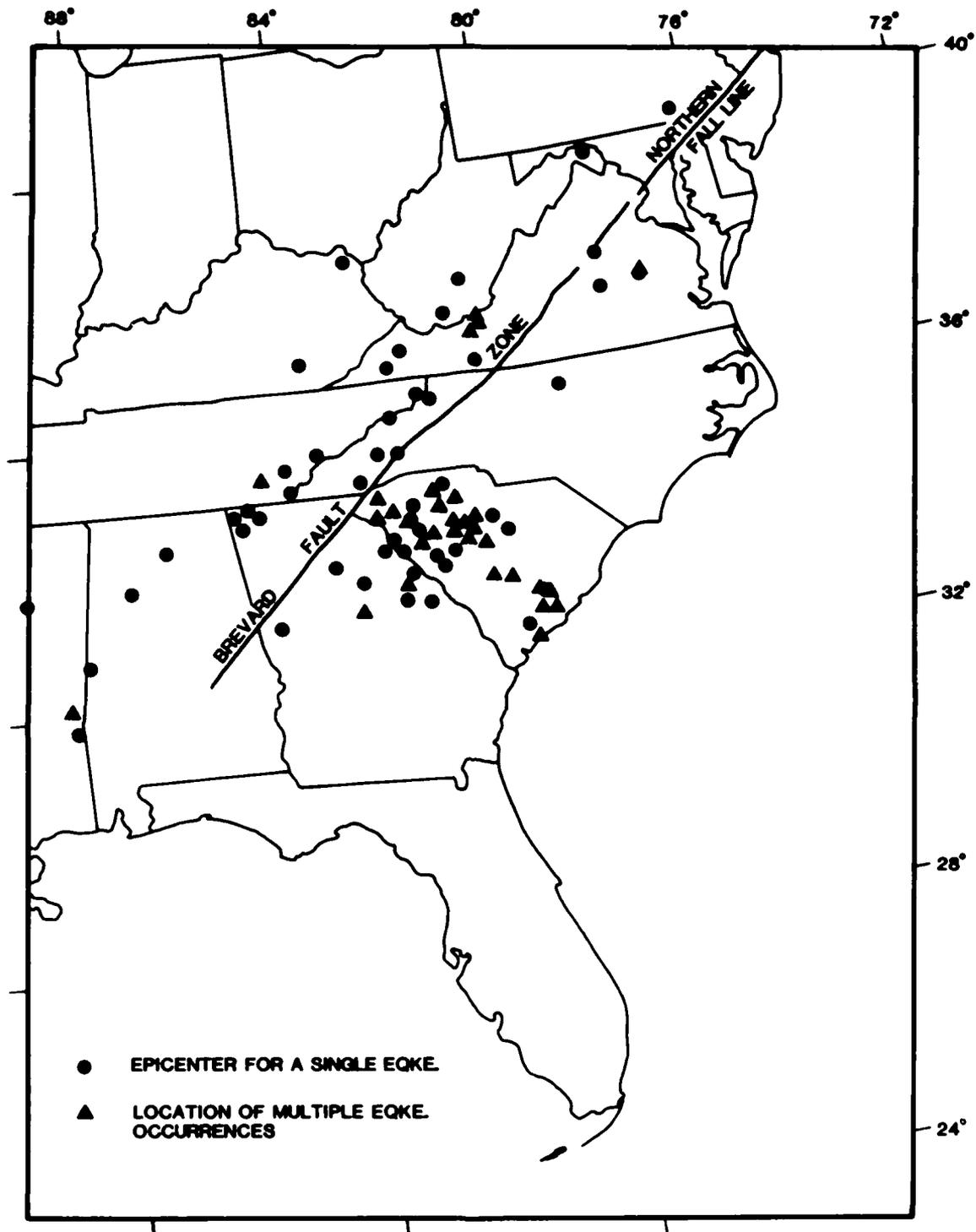
Figure 42. Epicentral map of southeastern United States showing earthquakes recorded during the period July 1977 through December 1979 (modified from Bollinger and Mathena, 1980)
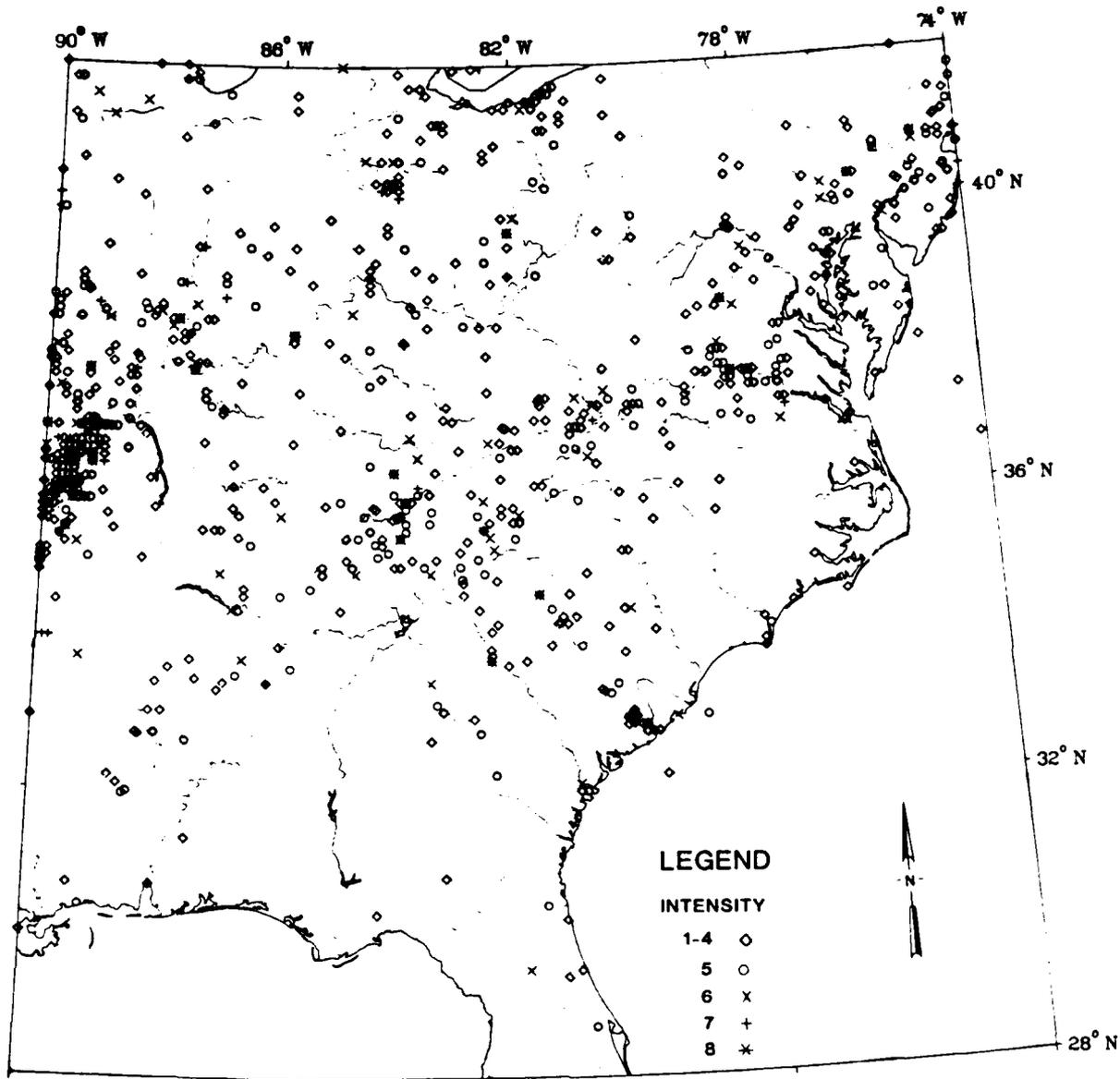
Figure 43. Epicentral map of southeastern United States showing historic and recorded earthquakes through 1980 (Rinehart, 1983)
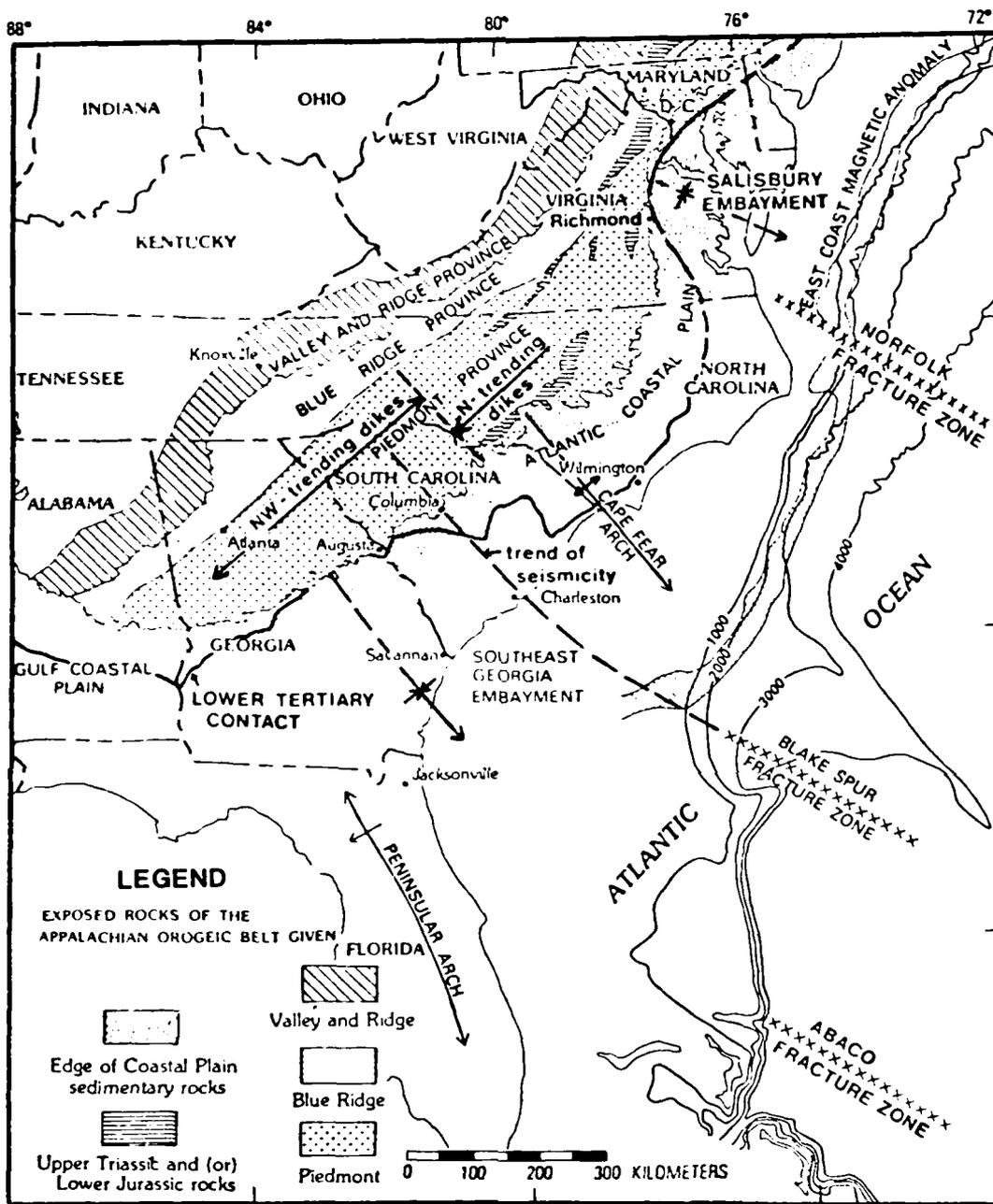
Figure 44. Map of the southeastern United States showing selected geologic features (modified from Rankin, 1977) (Barosh, 1981a, Fig. 10)
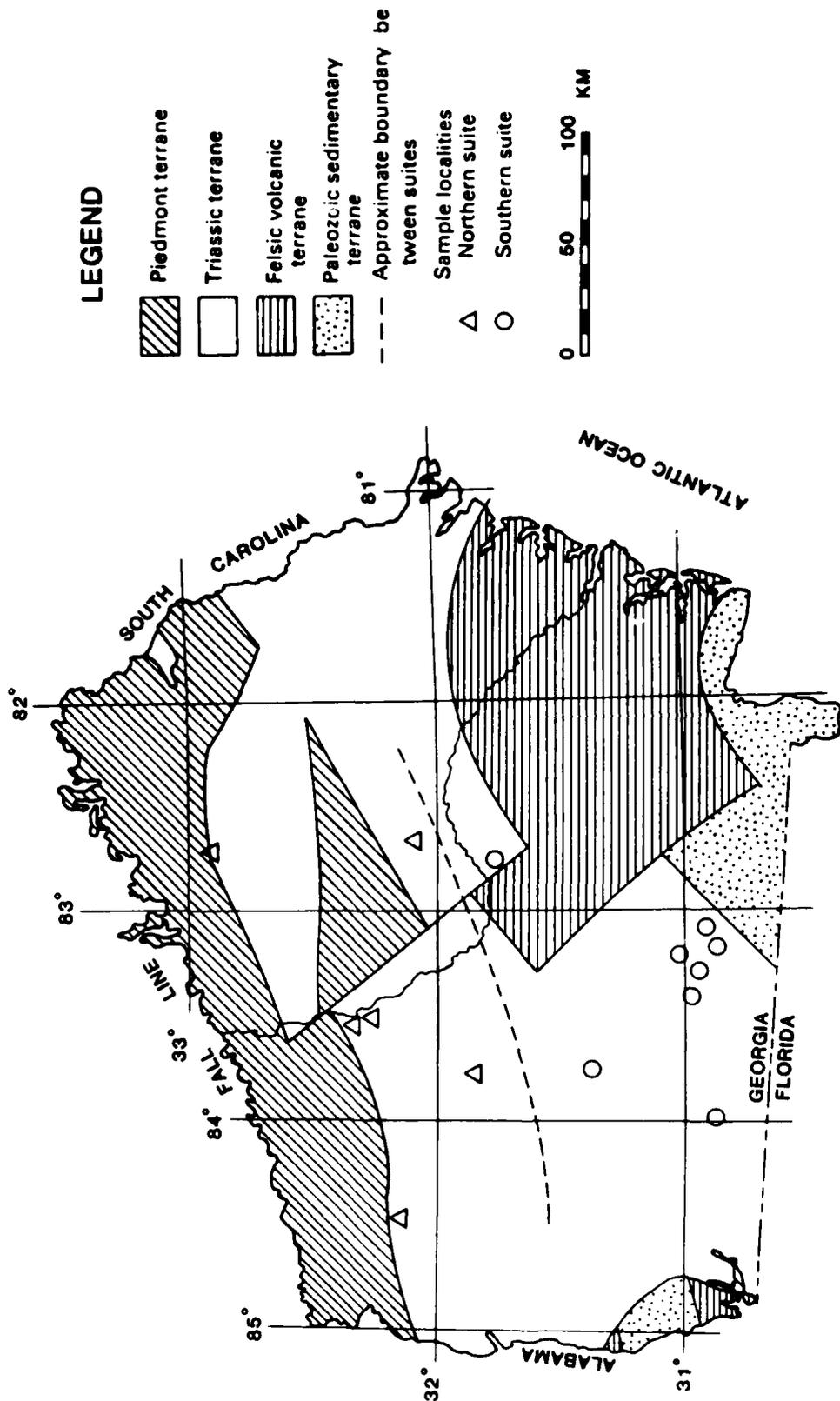
Figure 45. Generalized geologic map of the pre-Coastal Plain deposits of southern Georgia showing extent of the South Georgia Graben (Triassic terrane) (Chowns and Williams, 1983, Fig. 6)
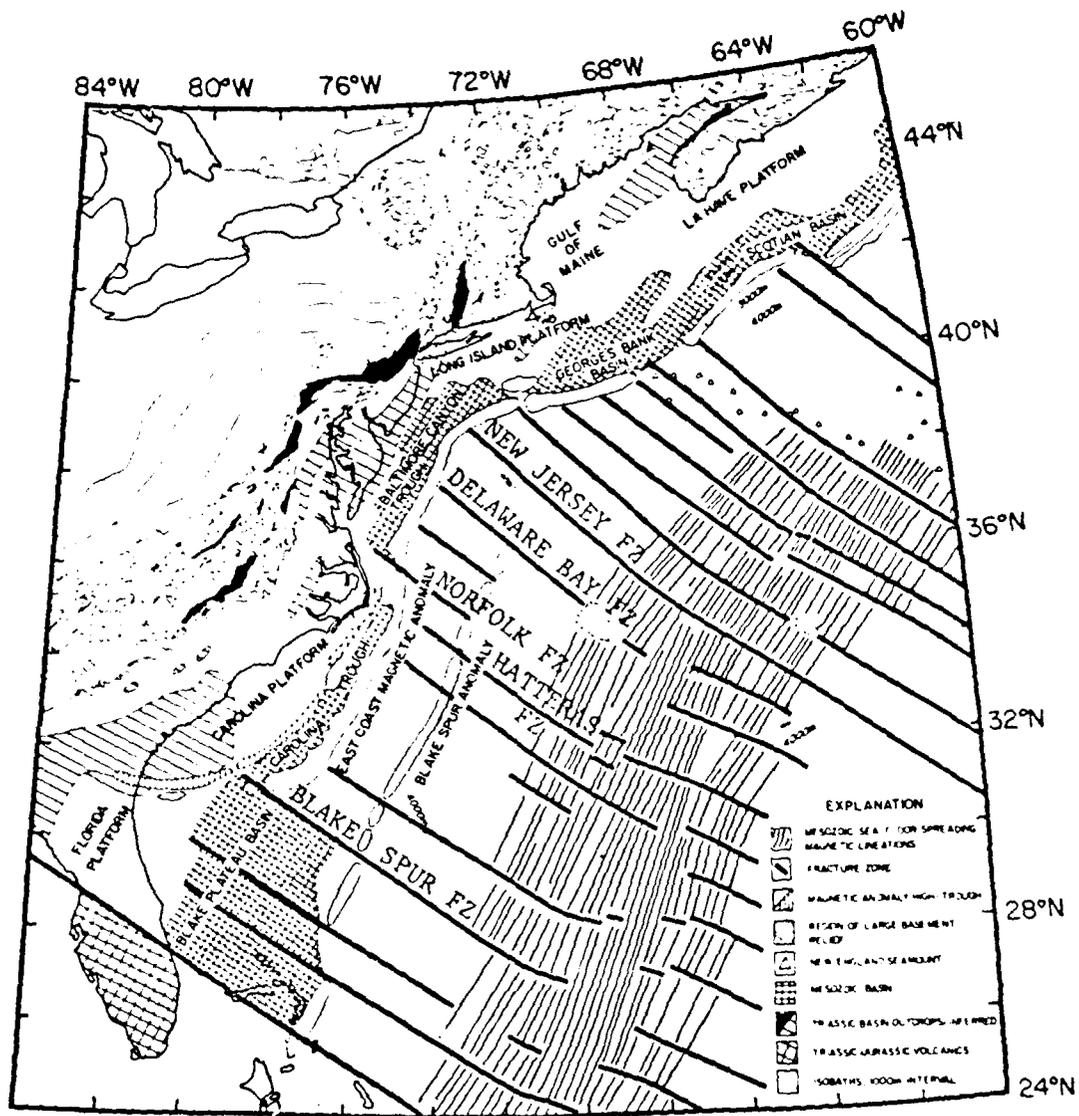
Figure 46. Map of the western North Atlantic Basin adjacent to the East Coast of the United States and Maritime Canada showing oceanic fracture zones and selected geologic features (Klitgord and Behrendt, 1979, Fig. 2)
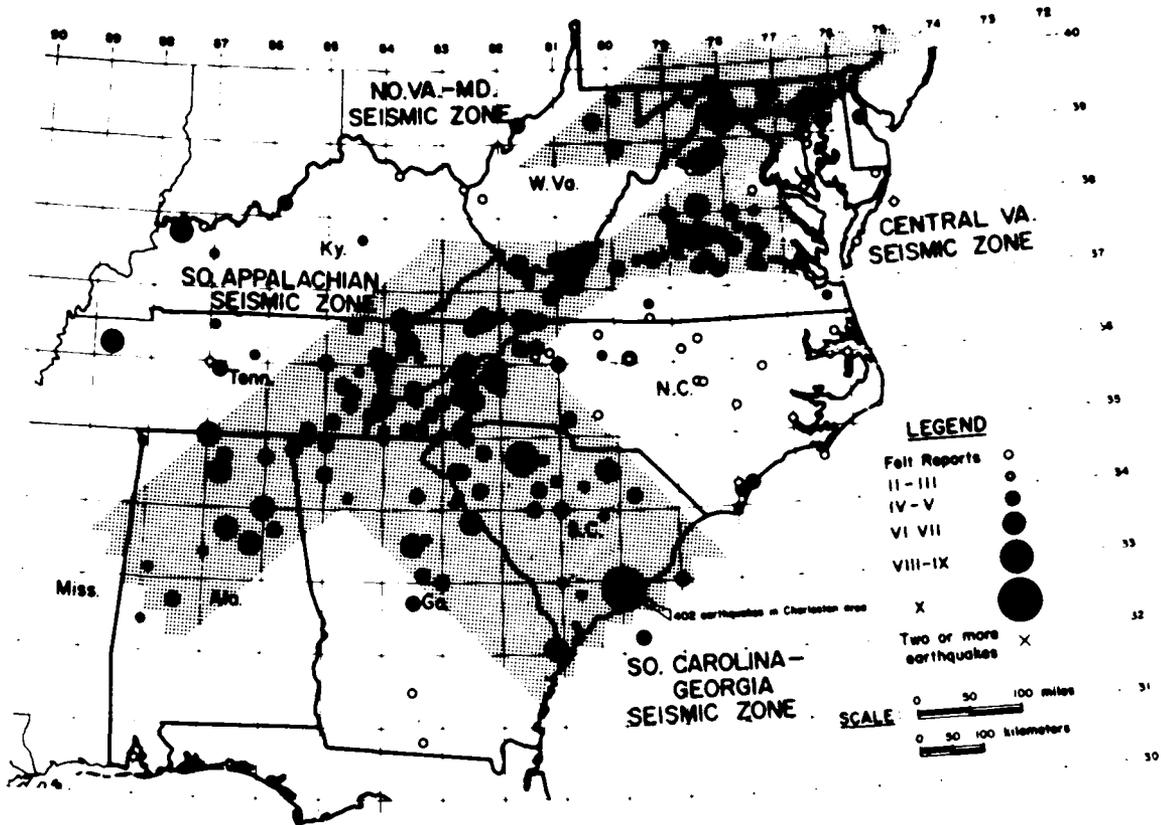
Figure 47. Epicentral map of the southeastern United States showing
seismic zones (Bollinger, 1973, Fig. 3)
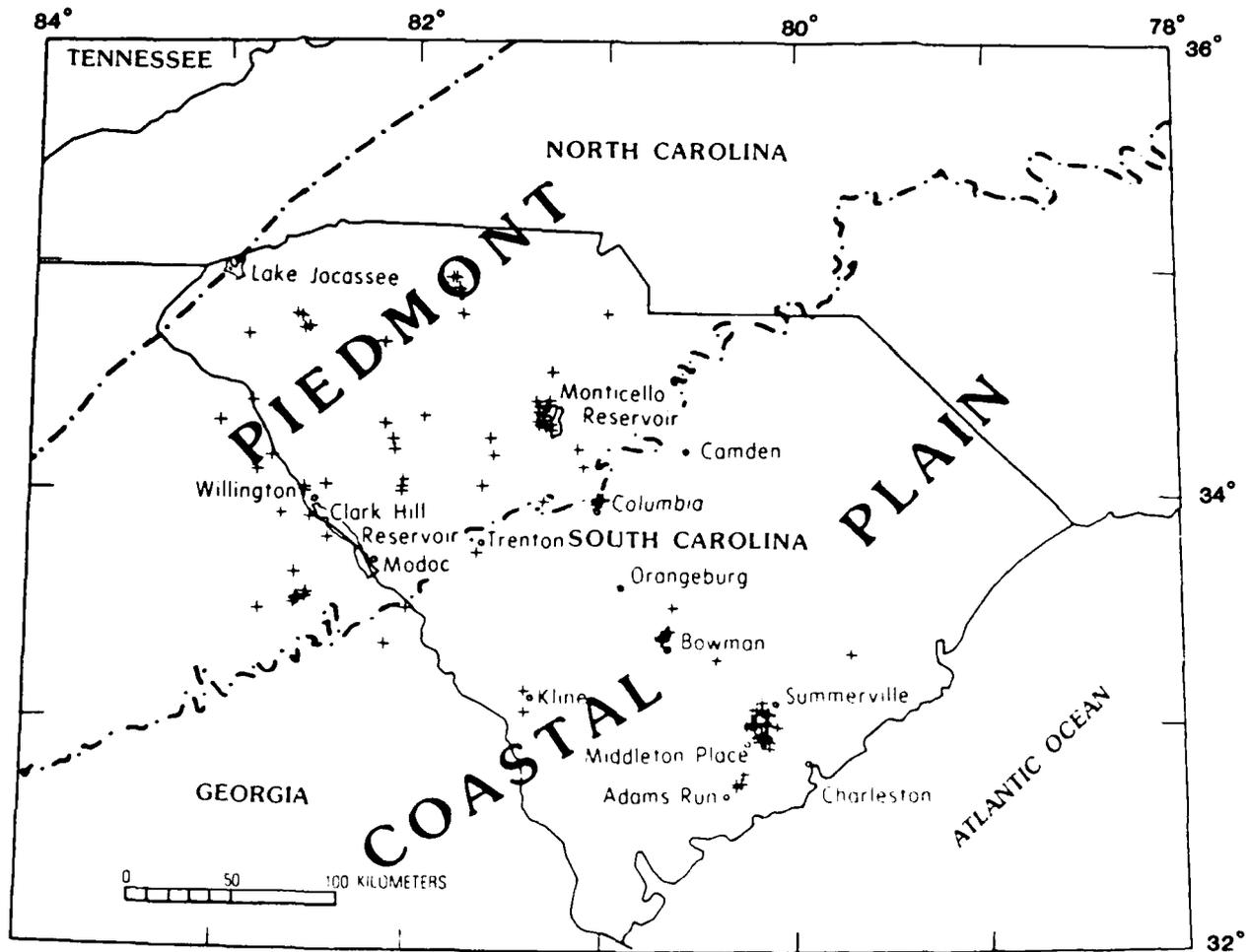
Figure 48. Epicentral map of South Carolina and adjacent area showing earthquakes recorded from 1973 through 1979 (Tarr and Rhea, 1983, Fig. 5)
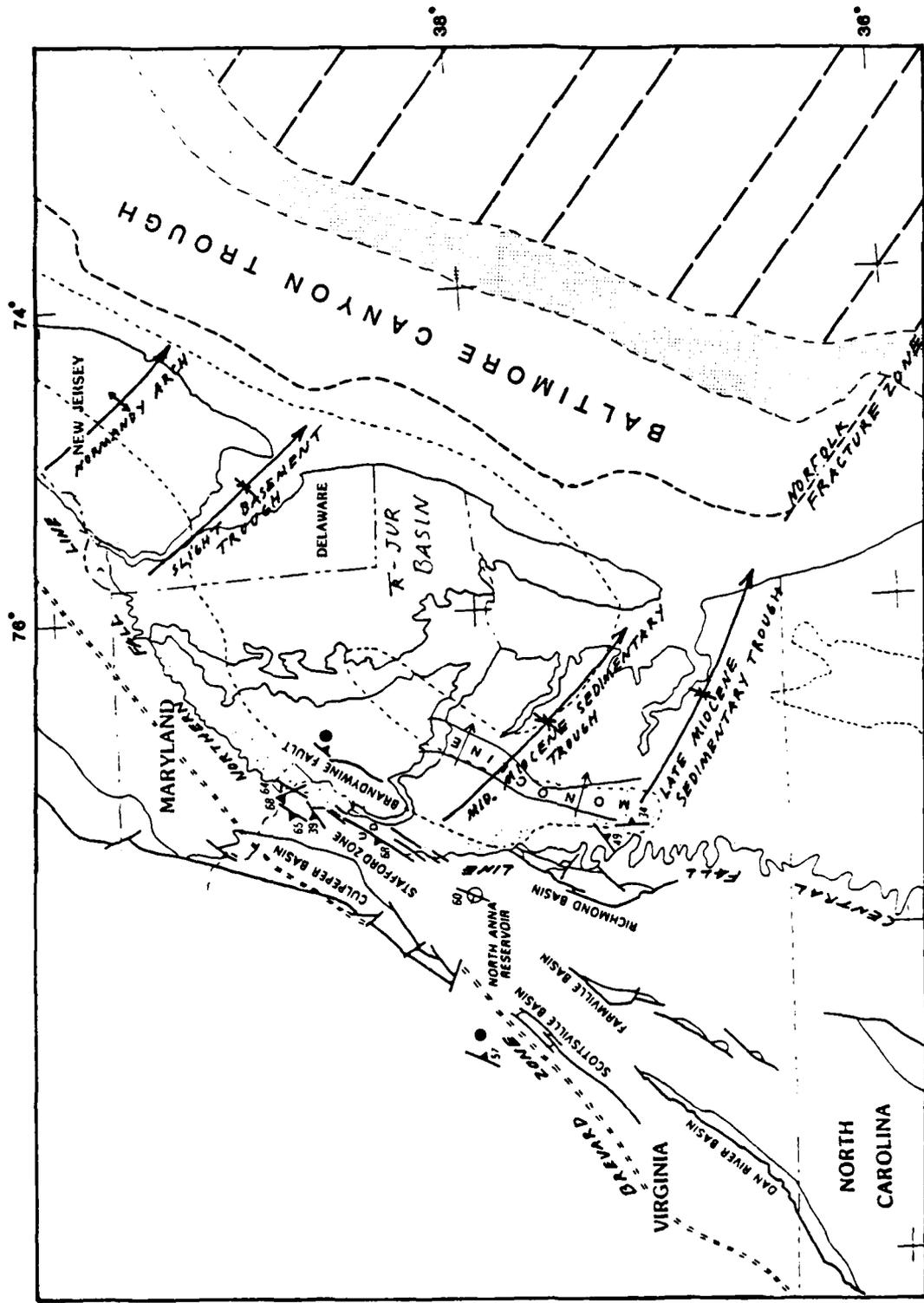
Figure 49. Map of the Chesapeake-Delaware Embayment and surrounding region showing early Mesozoic basins, oceanic fracture zones, and selected geologic features (modified from Wentworth and Mergner-Keefer, 1983)
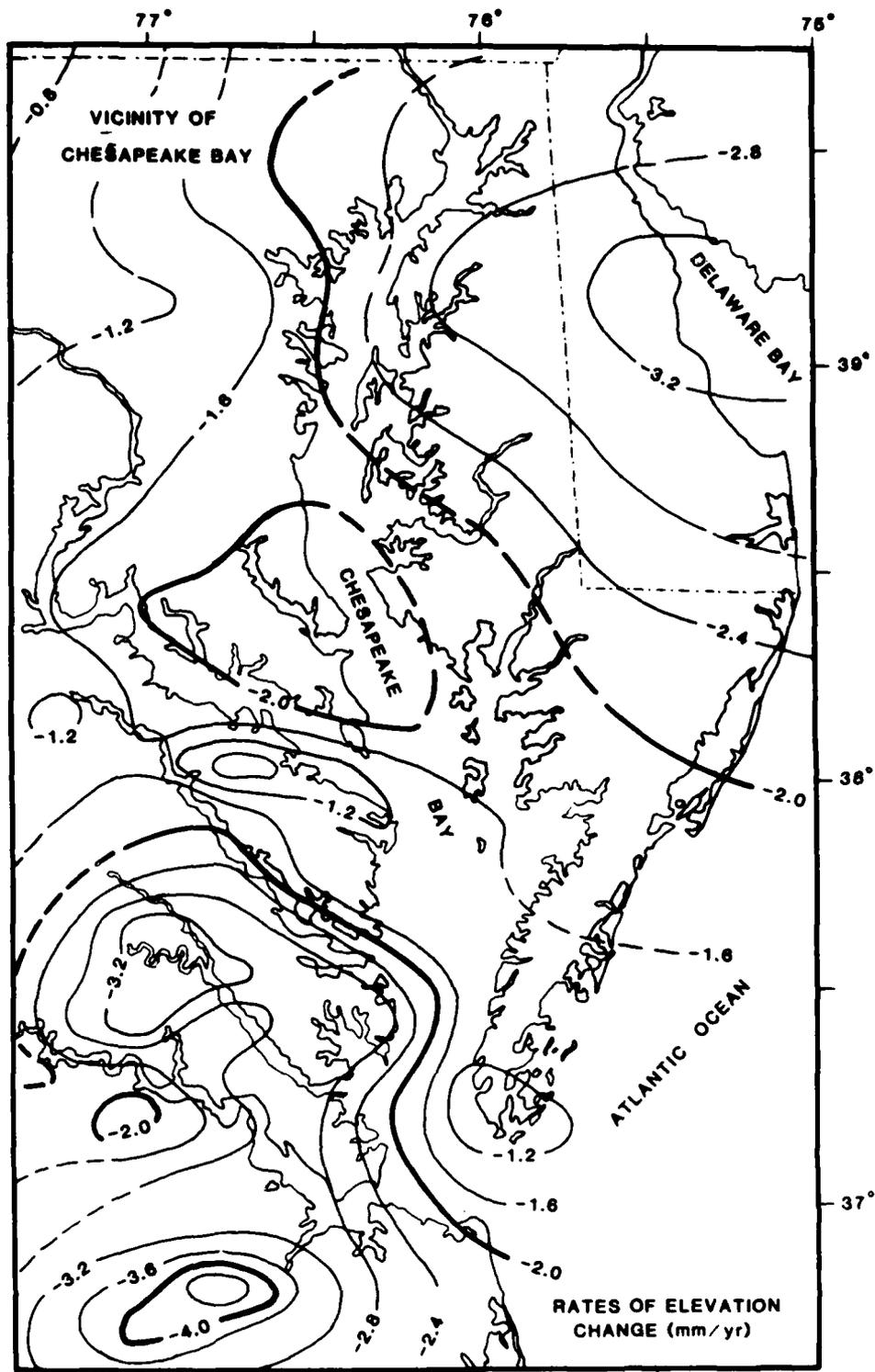
Figure 50. Map of the Chesapeake Bay area showing the relative rate of present-day vertical movement derived from leveling surveys (Holdahl and Morrison, 1974)
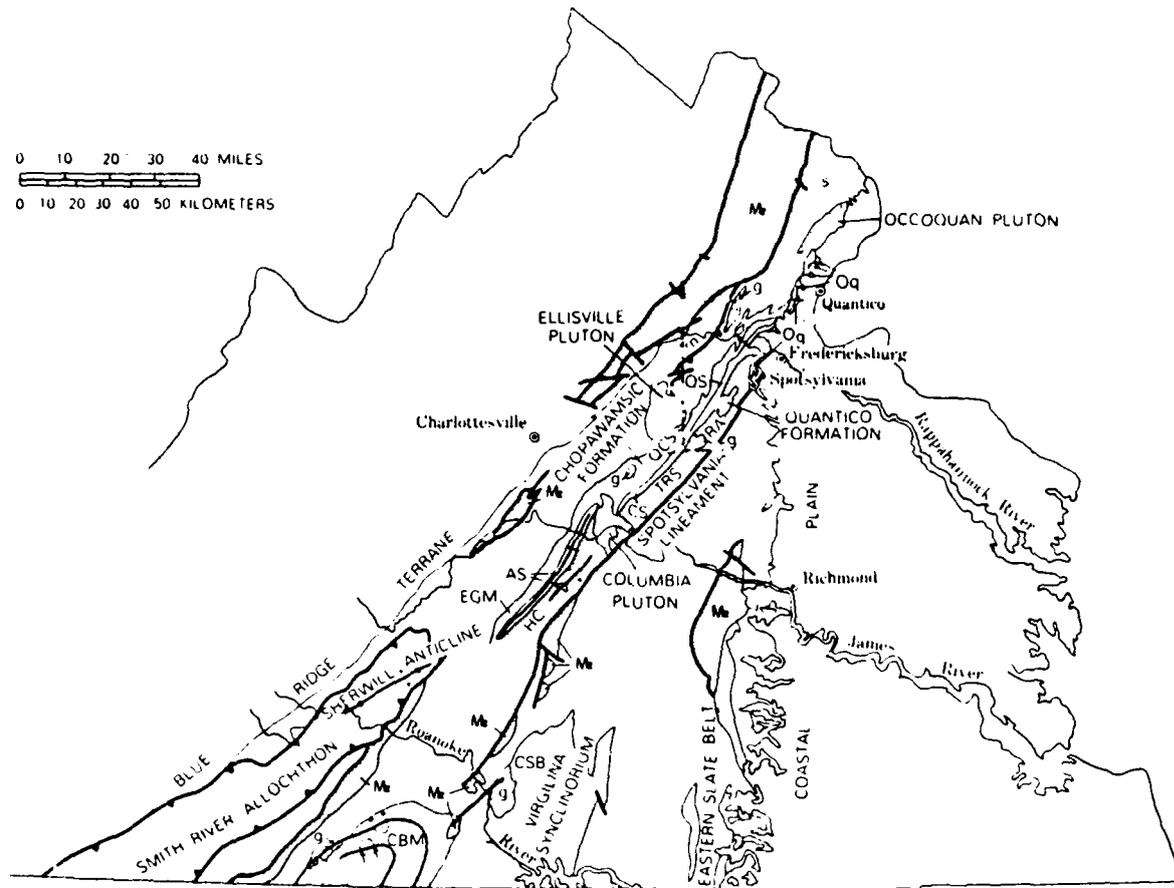
Figure 51. Map of Virginia showing early Mesozoic basins and selected geologic features (Pavlides, 1981, Fig. 1)
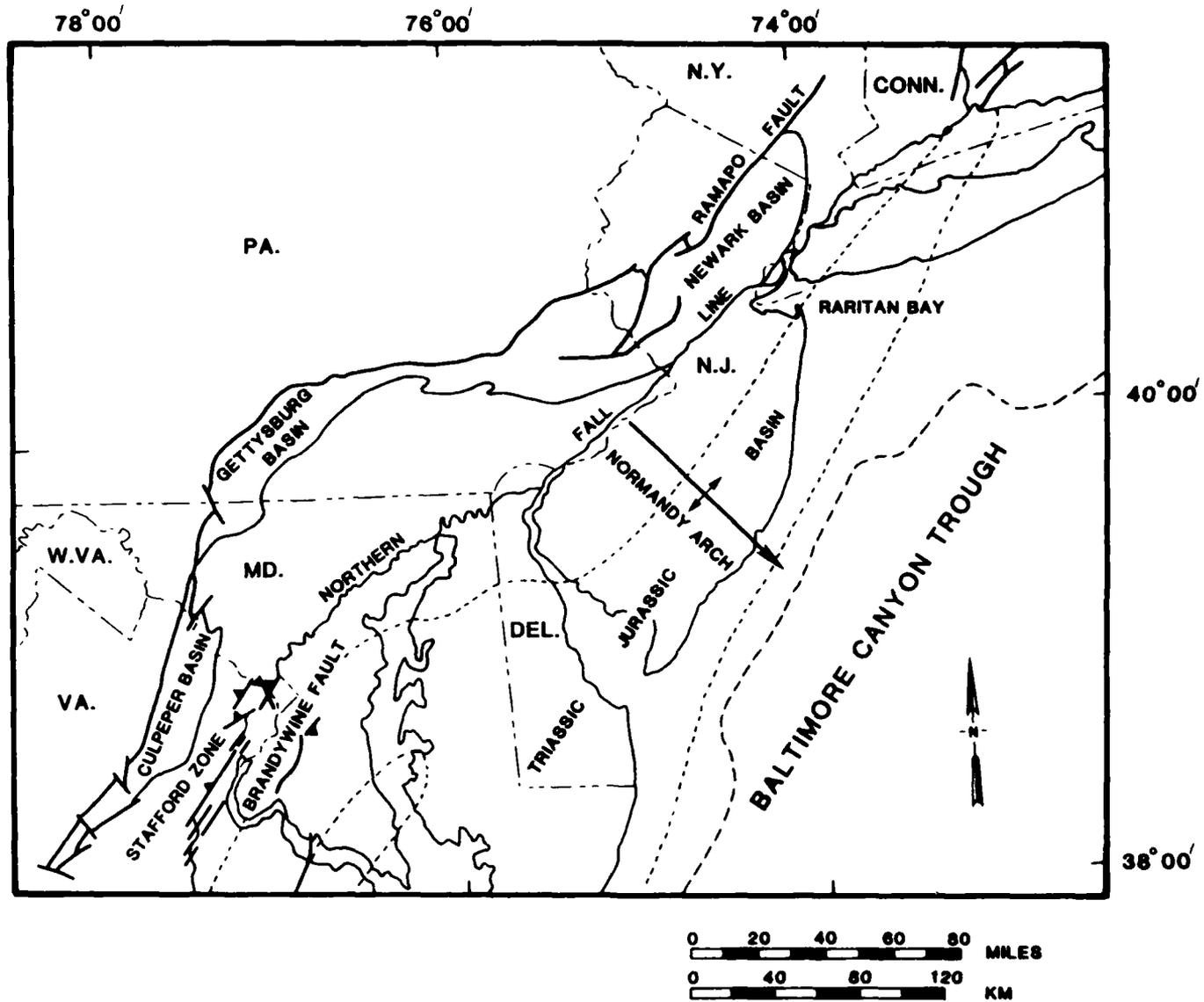
Figure 52. Map of Raritan Embayment and surrounding region showing location of early Mesozoic basins, oceanic fracture zones and selected geologic features (Wentworth and Mergner-Keefer, 1983)
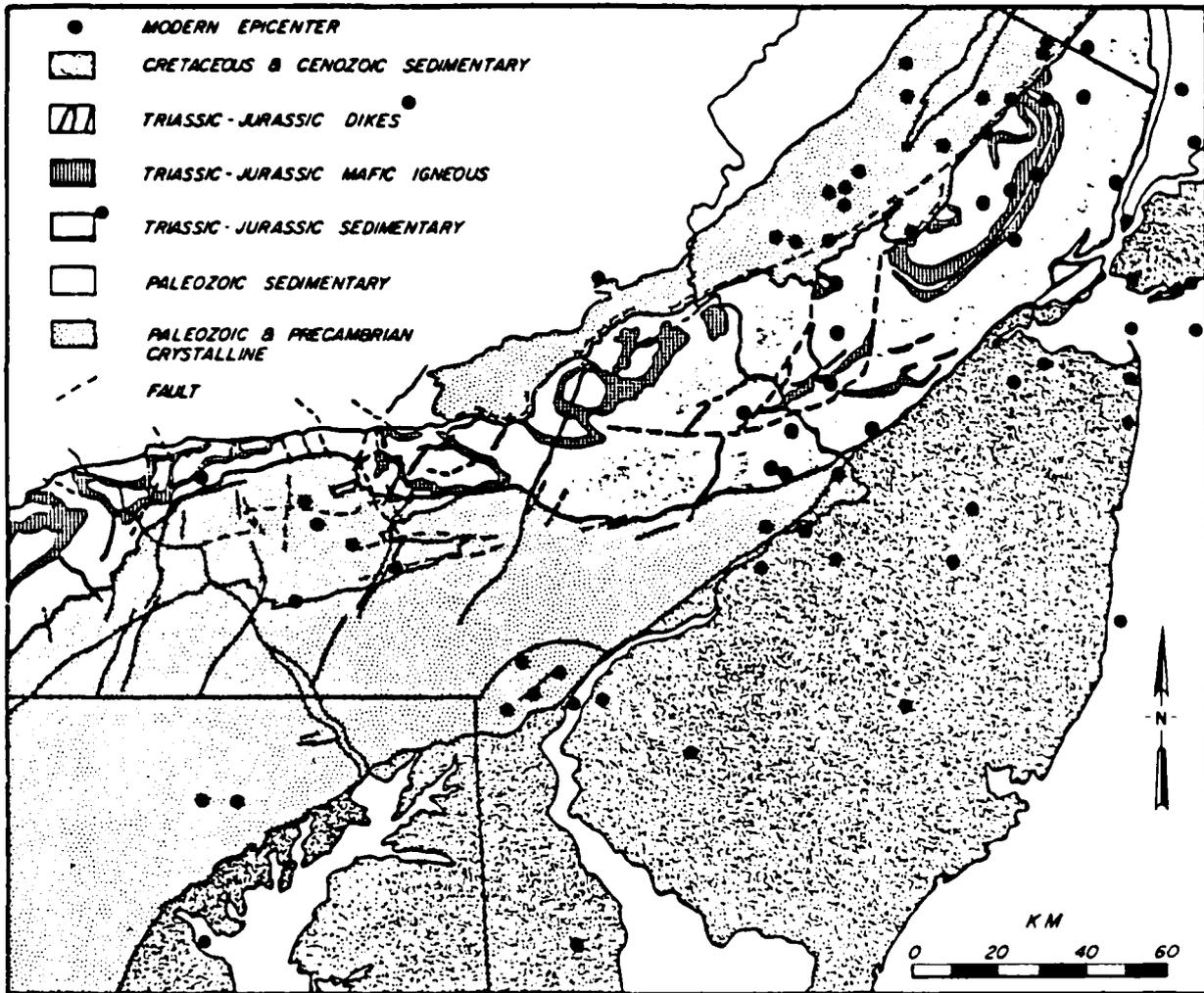
Figure 53. Map of New Jersey and surrounding region showing relation of earthquake epicenters to geologic features (Thompson, 1983, Fig. 4)
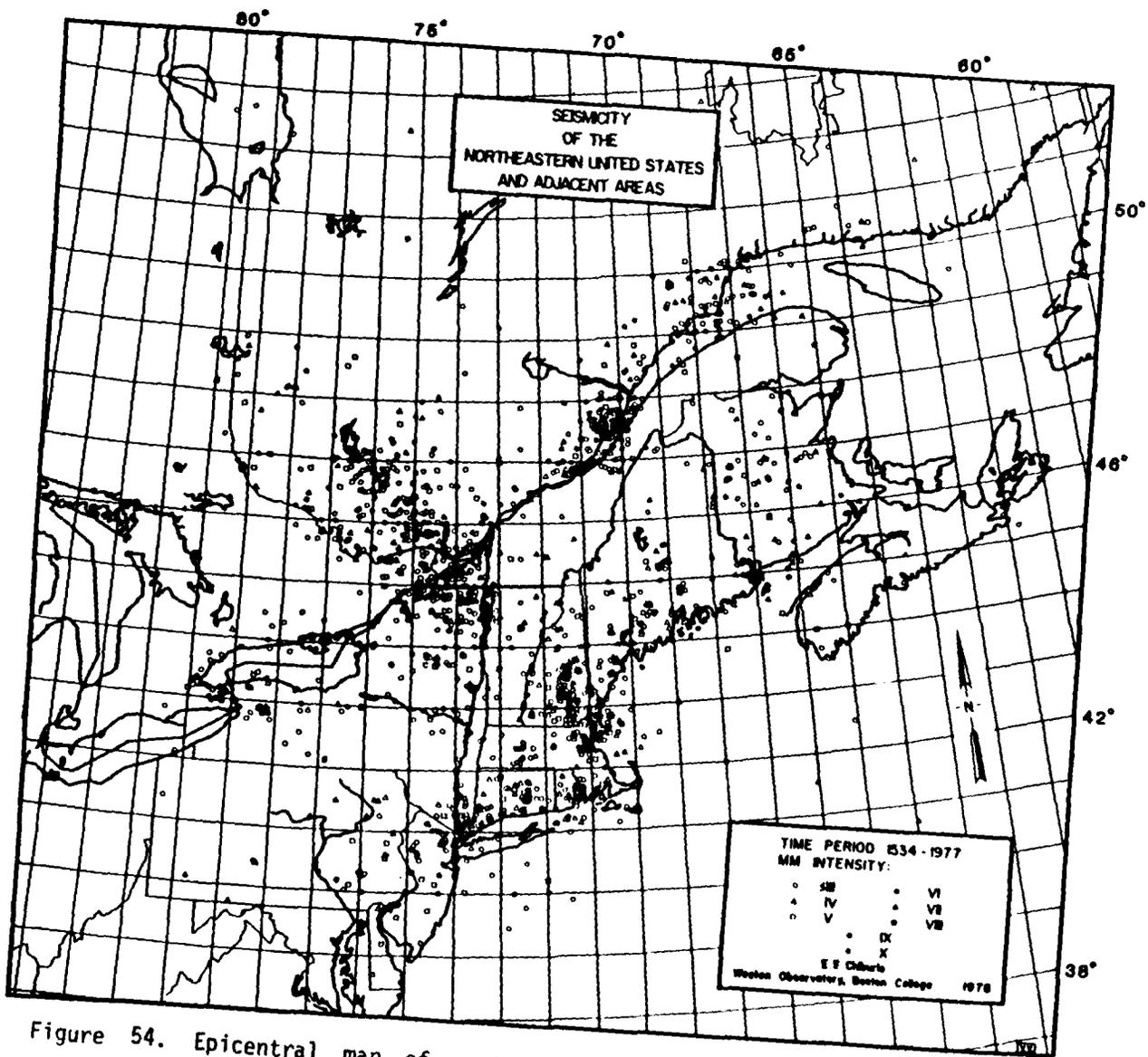
Figure 54. Epicentral map of northeastern United States and adjacent Canada for earthquakes from 1534 through 1977 (Chiburis and others, 1980)

Figure 55. Epicentral map of Raritan Bay and surrounding region for historic and recorded earthquakes through 1980 (Nottis and Mitronovas, 1983)

Figure 56. Map showing structural provinces and the border faults in the region around Moodus, south-central Connecticut (Barosh, 1982c, Fig. 4)
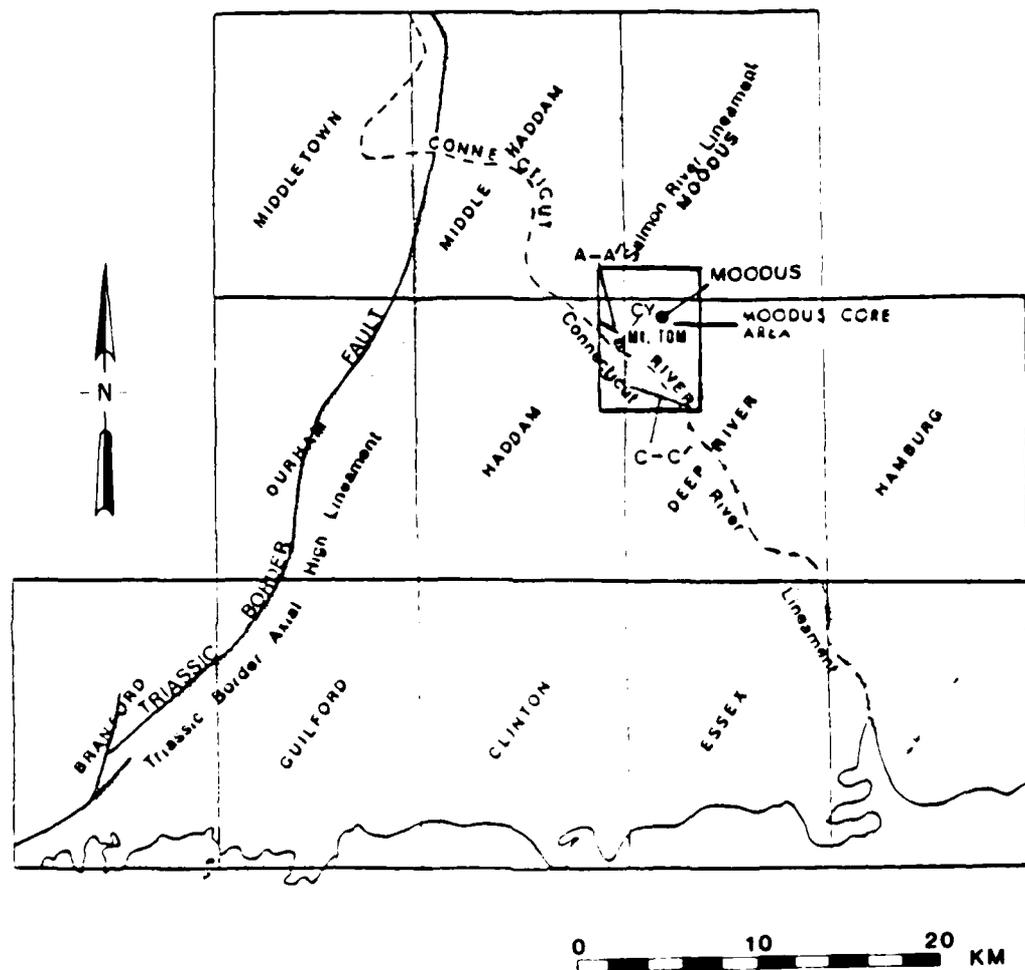
Figure 57. Index map of the Moodus area, south-central Connecticut, showing 7.5-minute quadrangles, selected gravity lineaments, and position of cross sections. Explanation: CY, Connecticut Yankee Nuclear Power Plant; strippling gravity lineament; A-A' and C-C', cross sections (Barosh, London, and de Boer, 1982, Fig. 3)

SYMBOLS

FAULTS

Dashed when inferred,
teeth where thrust fault

Intruded

FOLDS

Anticlinal axis, arrow showing plunge

Synclinal axis, arrow showing
plunge

Synclinal axis, overturned

DIKES

Diabasic (Early Jurassic)

● Ultramafic (Paleozoic)

FIELD STOPS

X2 Stops in road log

INITIALS

FAULTS

B.F.	Border fault
M.H.	Middle Haddam
B.B.	Bonemill Brook
I.H.	Injun Hollow
C.H.	Cremation Hill
F.R.	Falls River
S.N.	Seldon Neck
H.H.	Honey Hill

FOLDS

G H	Great Hill syncline
K D	Killingworth dome
M	Moodus anticline
V P	Vincent Pond basin
L	Lyme anticline

DIKE SYSTEM

H.D.	Higganum dike

0   10   20 KM

Figure 58. Generalized geologic structure map of south-central Connecticut
(after unpublished data from M.H. Pease, Jr., P.J. Barosh,
Jelle de Boer, David London, R.P., Wintsch, Brian Koch, R.J.
Fahey and others, and Sawyer and Carroll, 1982) (Barosh,
London, and de Boer, 1982, Fig. 6)
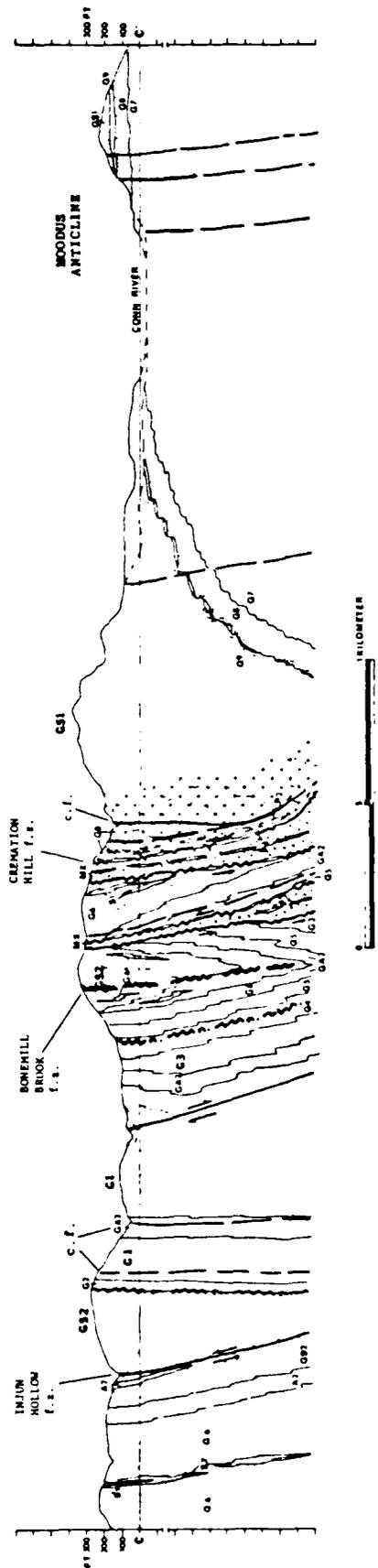
Figure 59. Geologic cross section through the southern part of the Moodus core area view north (see Fig. 57) (after London in prep.) Explanation: F.z., fault zone; c.f., cross fault; T and A, P, pegmatite; MX, migmatitic zone (Barosh, London, and de Boer, 1982, Fig. 8)
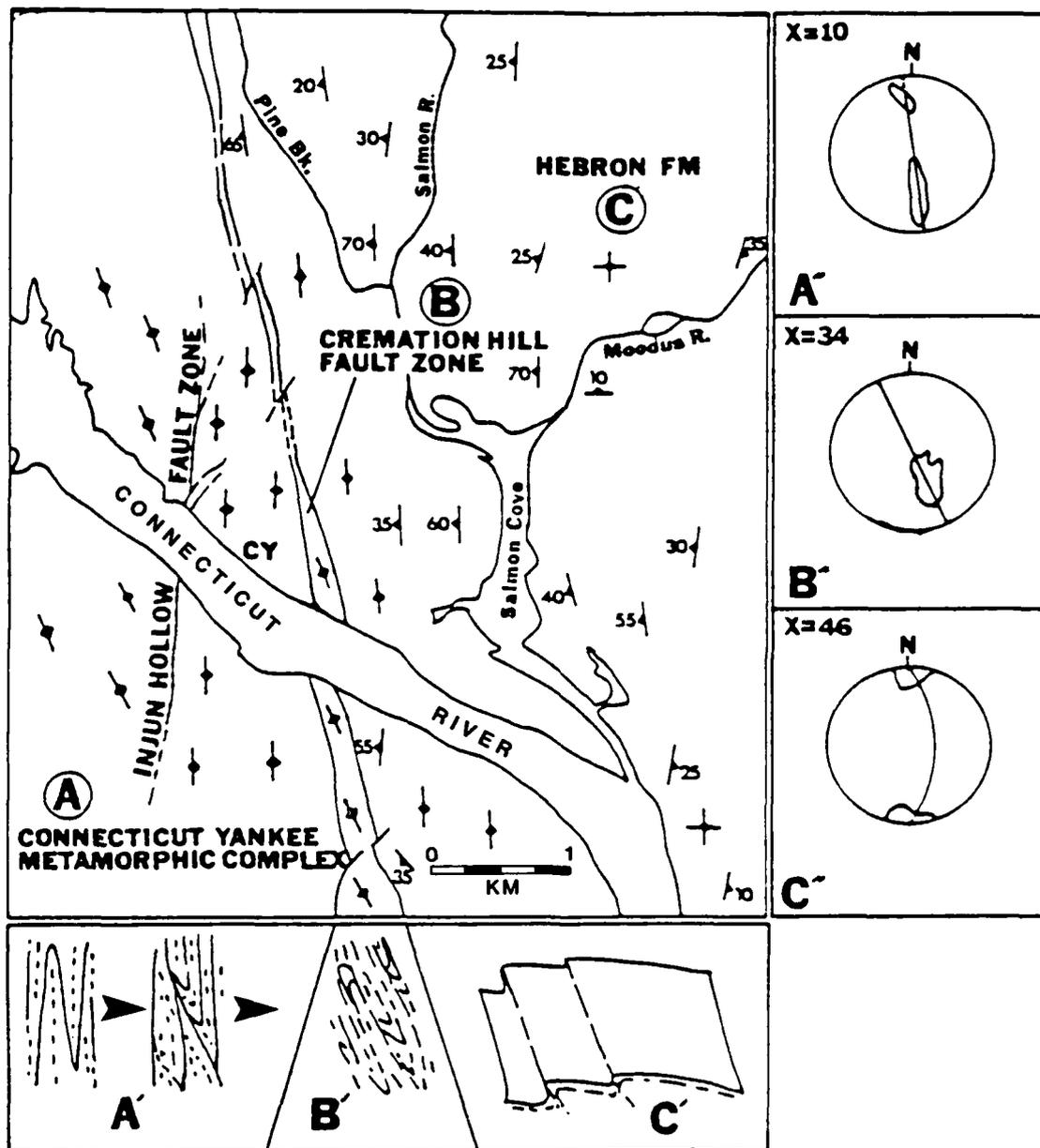
Figure 60. Map showing foliation and small-scale folds in the Moodus area,
south-central Connecticut (Barosh London, and de Boer,
1982, Fig. 10)
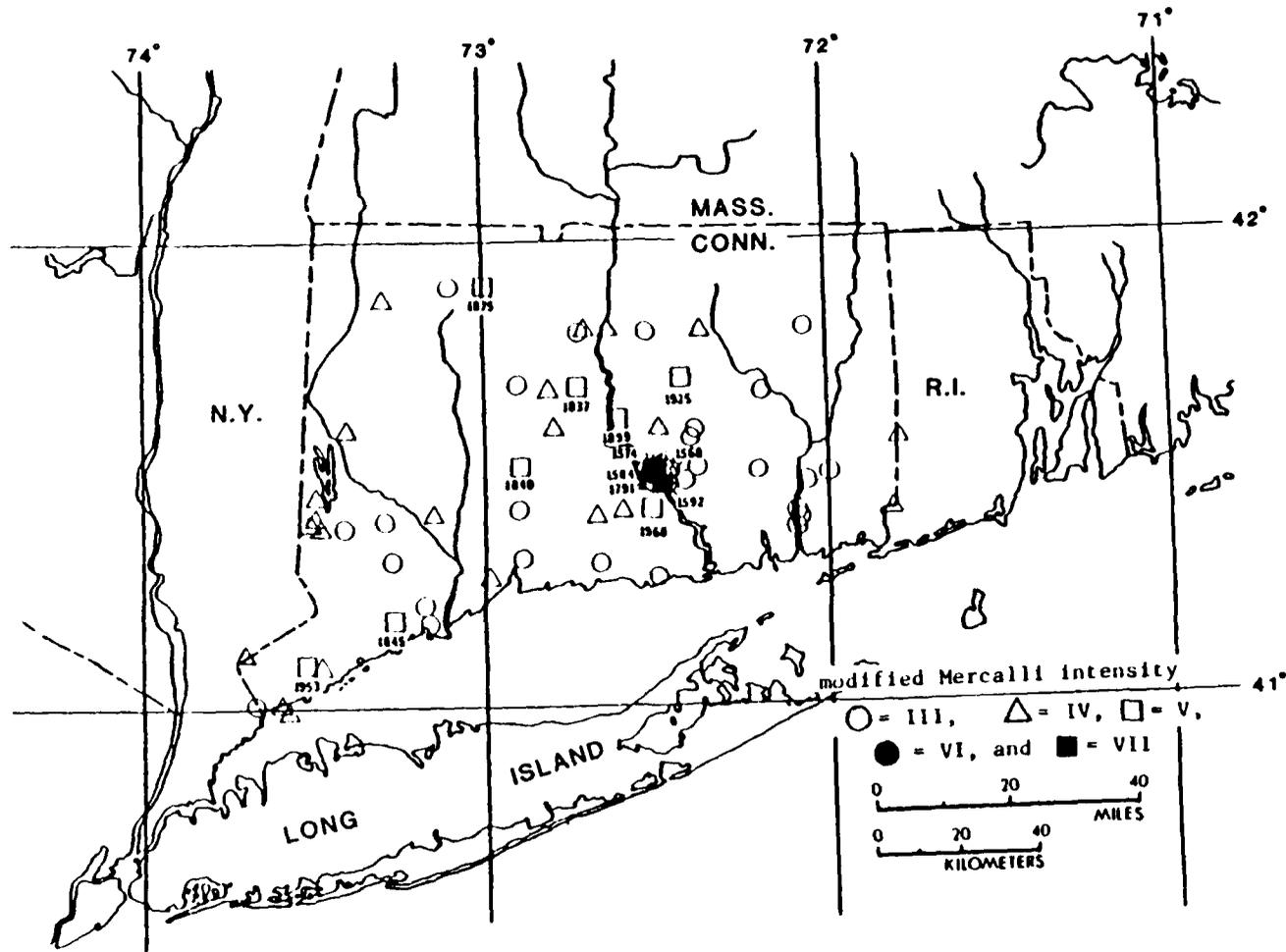
Figure 61. Epicentral map of Connecticut through 1977, the symbols show the maximum Modified Mercalli intensity for the events. Moodus is located in the area of concentration of events on the map (derived from Chiburis and others, 1980) (Barosh, London, and de Boer, 1982, Fig. 1)
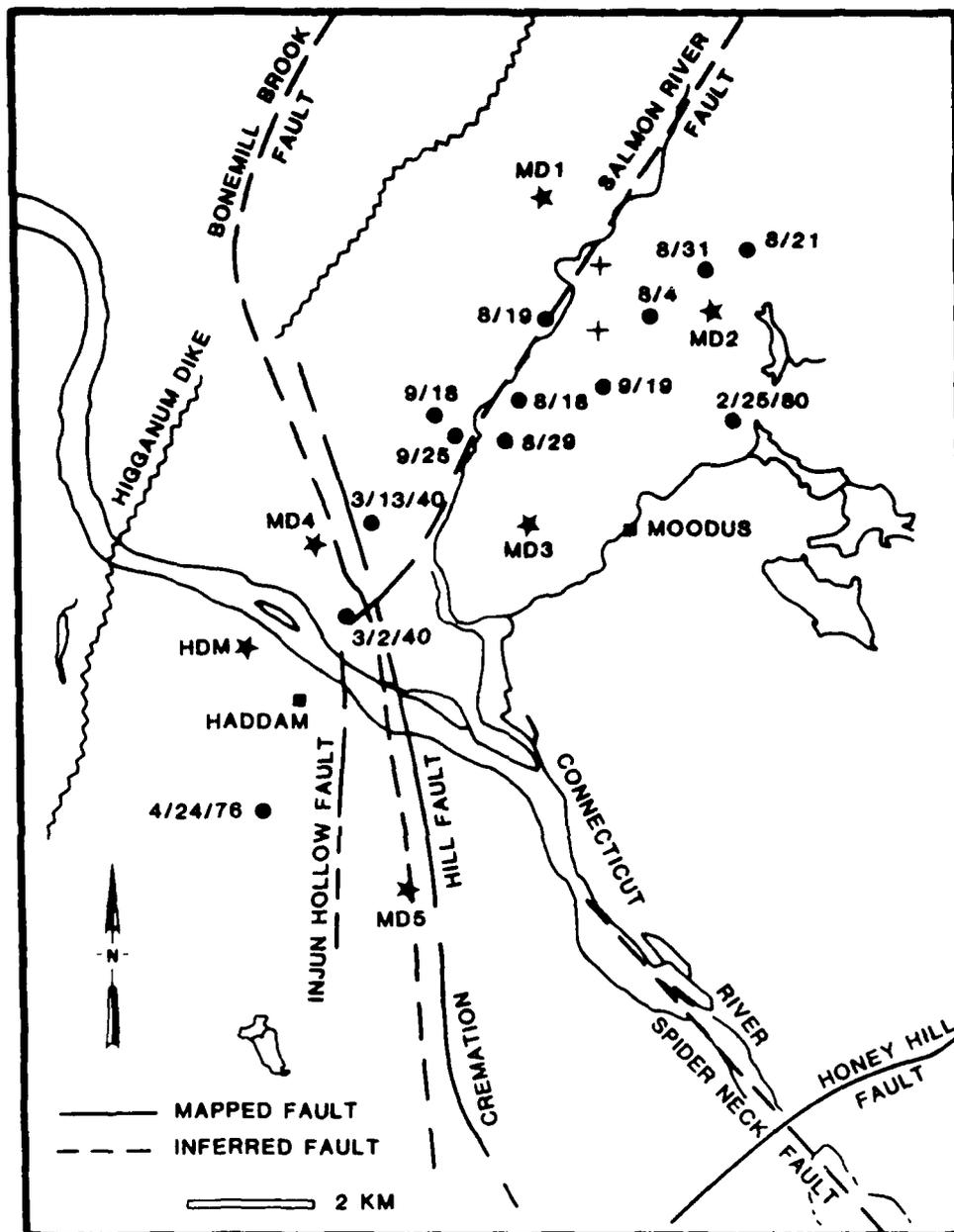
Figure 62. Map of the Moodus area, south-central Connecticut showing
locations of recent earthquakes and selected fault zones. The
dates of the locations of events from 1982 swarm are shown, as
well as dates and locations from previously located events.
Locations of the Moodus network stations are shown as stars,
and sites of the two temporary stations which were installed
after the August 4 event are shown as crosses (Modified from
Ebel and others, 1982) (Barosh, London, and de Boer, 1982,
Fig. 2)
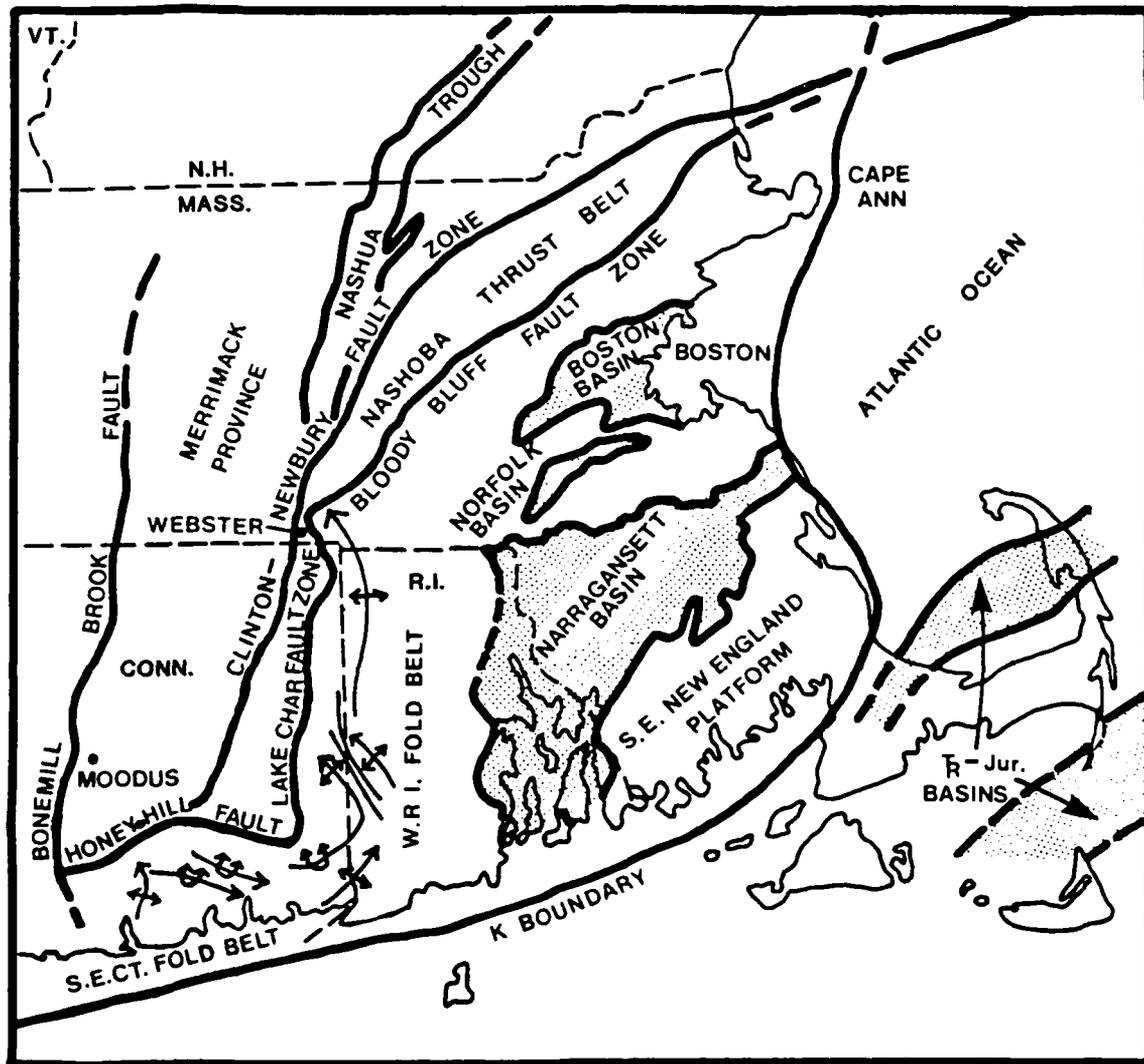
Figure 63. Map of southeastern New England showing major geologic provinces and few selected geologic features (Barosh, 1982a, Fig. 1)
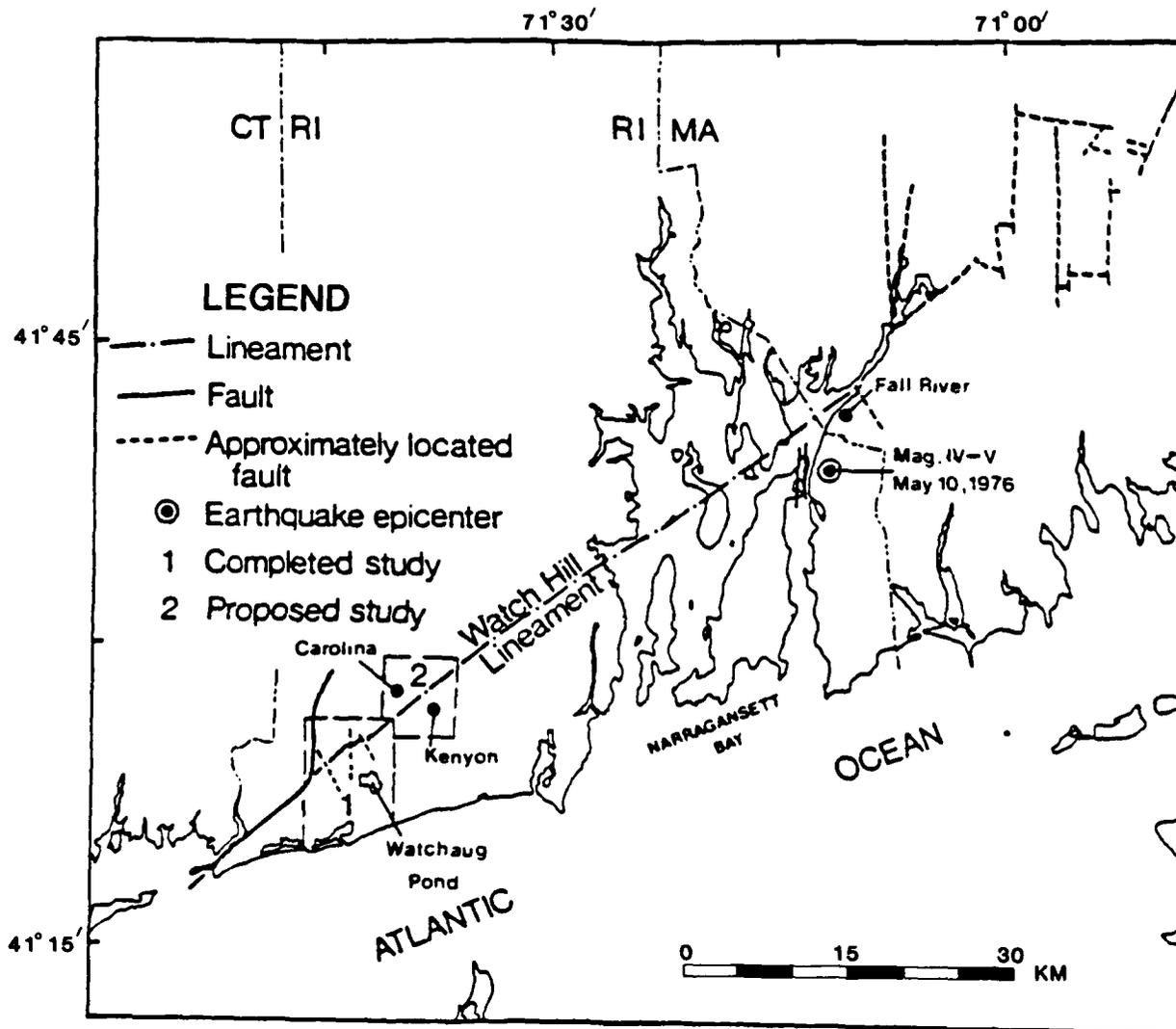
Figure 64. Sketch map of Rhode Island and southeastern Massachusetts showing the locations of the Watch Hill lineament and the epicenter of the 1976 Fall River earthquake (Barosh and Smith, 1982, Fig. 85)
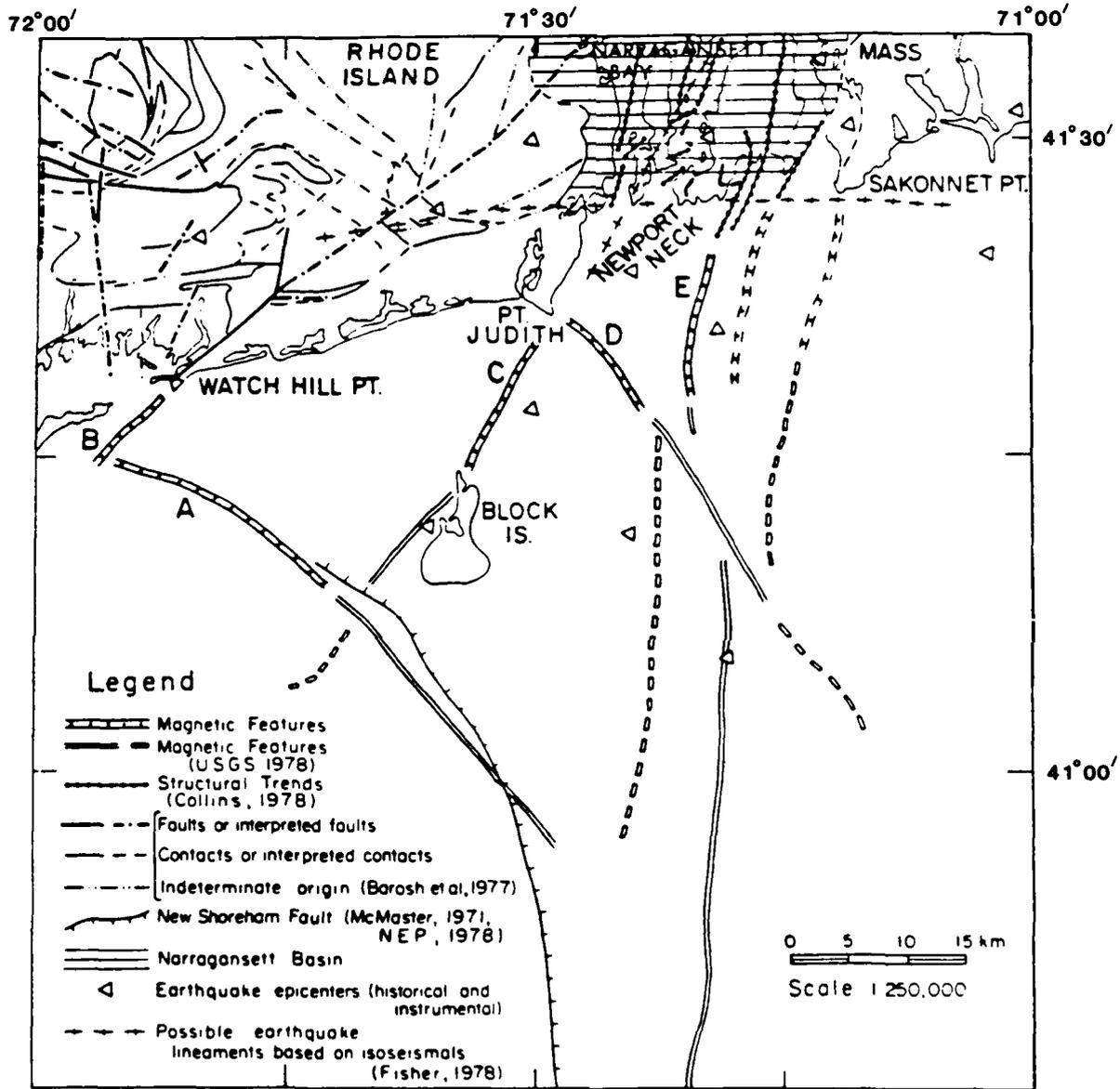
Figure 65. Map of southern and offshore Rhode Island showing land aeromagnetic lineaments, offshore magnetic lineaments (A-E), New Shoreham fault, and seismicity (from McMaster, de Boer, and Collins, 1980)
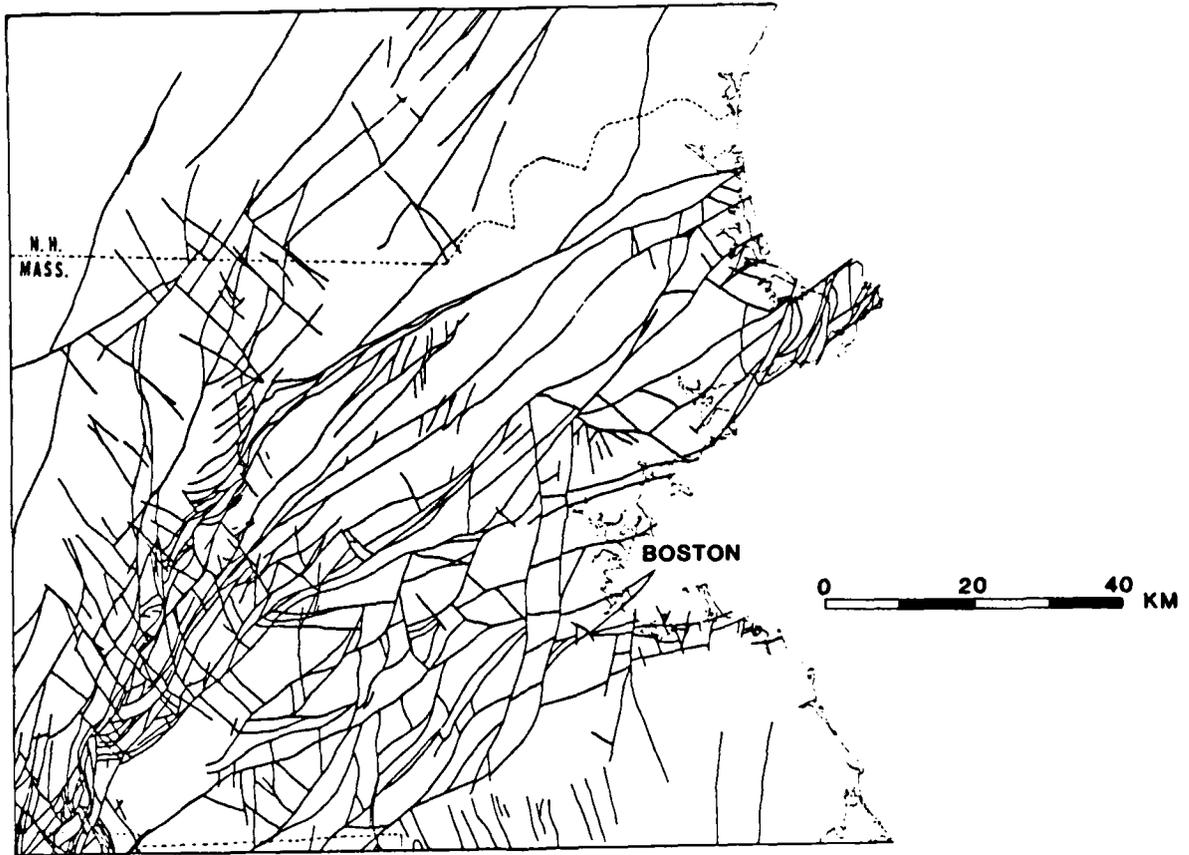
Figure 66. Mapped faults of eastern Massachusetts and southern New Hampshire
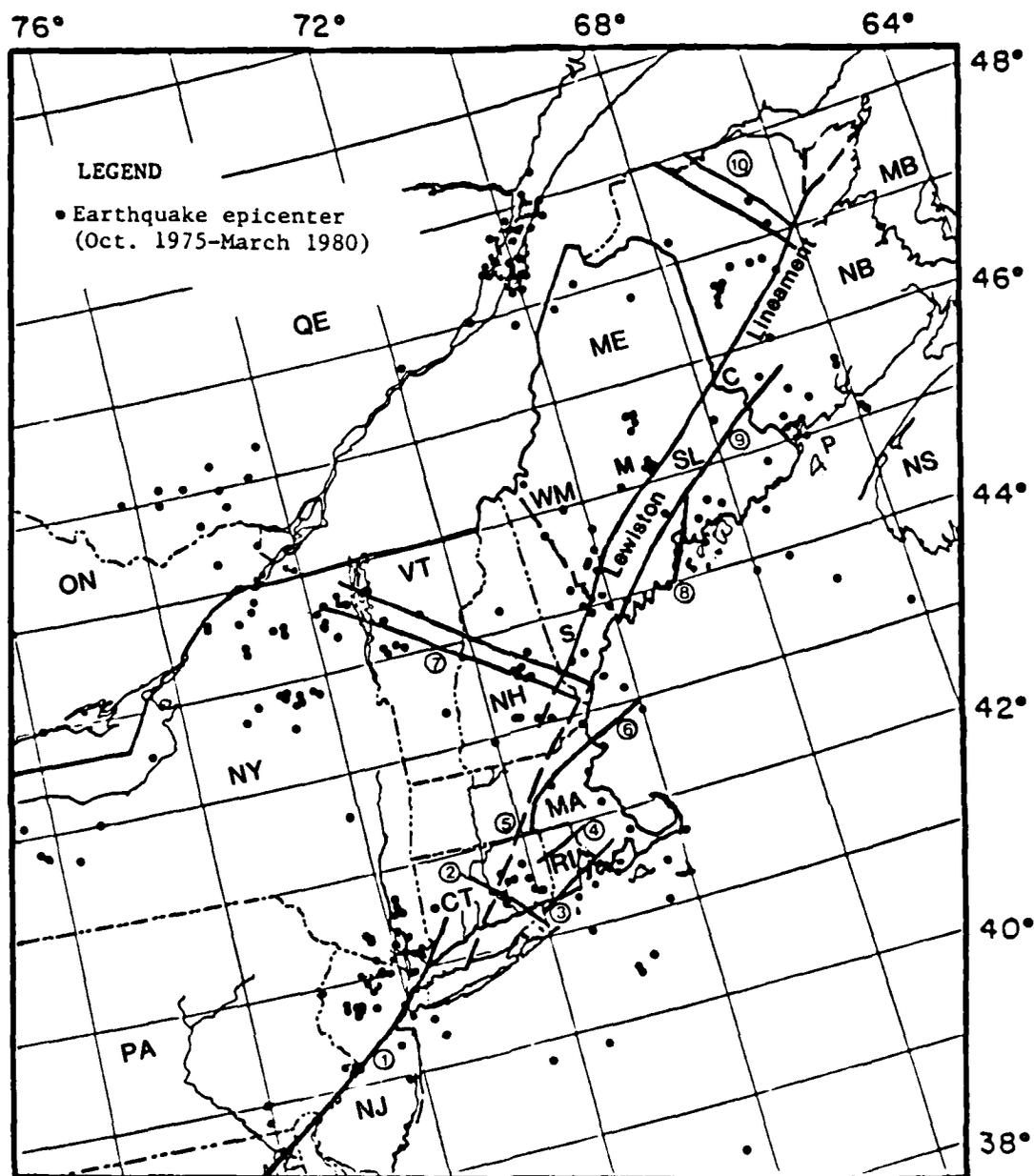
Figure 67. Map of the northeastern U.S. and adjacent Cana a showing the Lewiston gravity lineament and other selected interpreted structures found or greatly extended on the basis of magnetic and gravity data. 1. Northern Fall Line; 2. Connecticut River lineament; 3. Watch Hill lineament; 4. North Scituate-Blackstone lineament zone; 5. Higganum dike system; 6. Clinton-Newbury fault zone; 7. Winnipesaukee-Winnooski lineament zone; 8. Penobscot lineament; 9. Norumbega fault zone; and 10. Upsalquitch lineament zone. Some geographical locations are: S, Sebago Lake; L, Lewiston; M, Medford; SL, South Lincoln; P, Passamaquoddy Bay; C, Chiputneticook Lakes; MB, Miramichi Bay; WM, West Maine (modified from Barosh, 1982d)
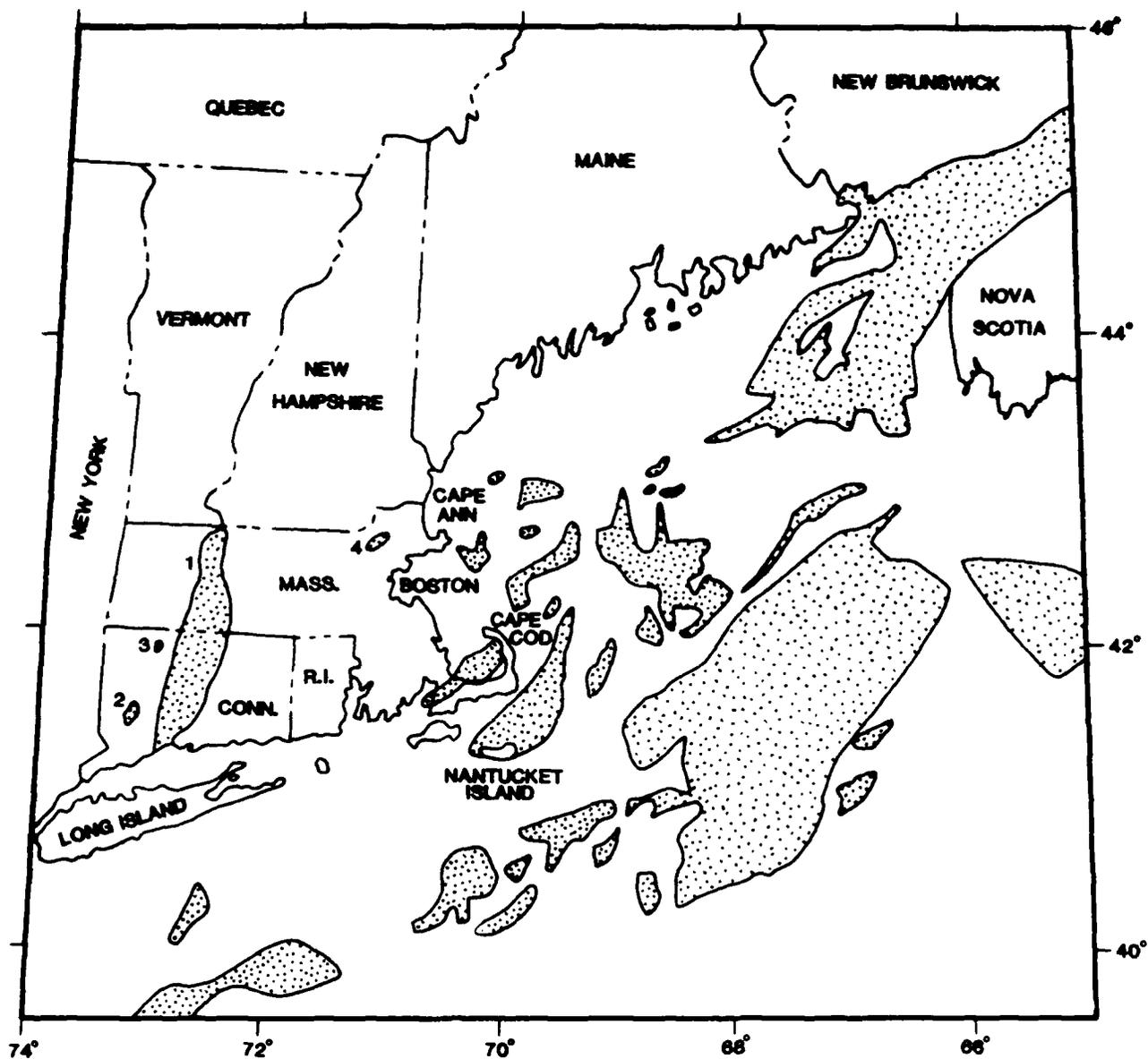
Figure 68. Map of New England and adjoining New York and Canada showing locations of basins containing Newark Group (stippled). Offshore basins from Ballard and Uchupi (1975). Onshore basins are the Connecticut Valley basin (1), Pomperaug Valley basin (2), Canton Center Basin (3), and Middleton basin (4) (Kaye 1983a, Fig. 1)
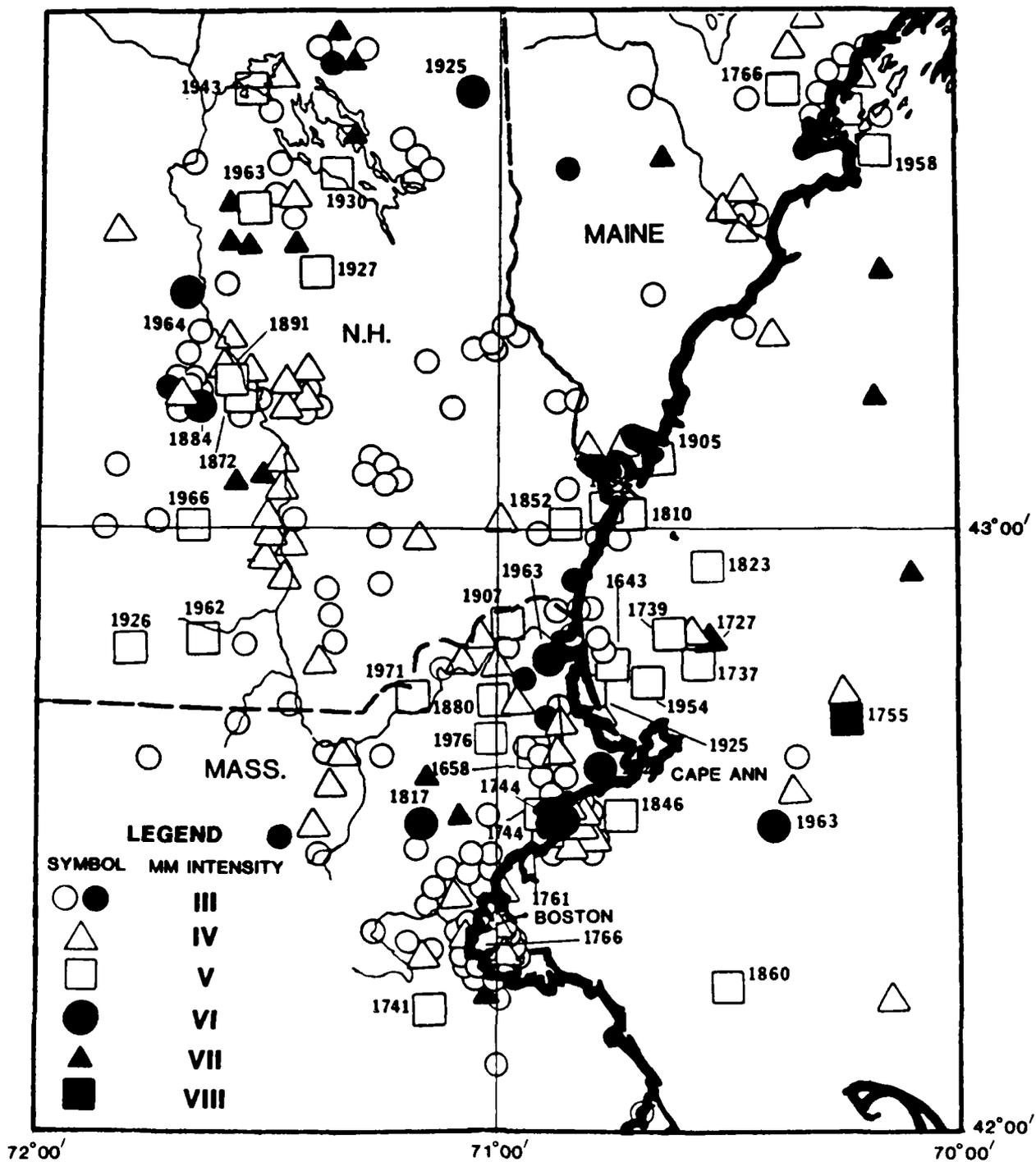
Figure 69.   Epicentral map of southeastern New England for earthquakes through 1980 (Nottis and Mitronovas, 1983)
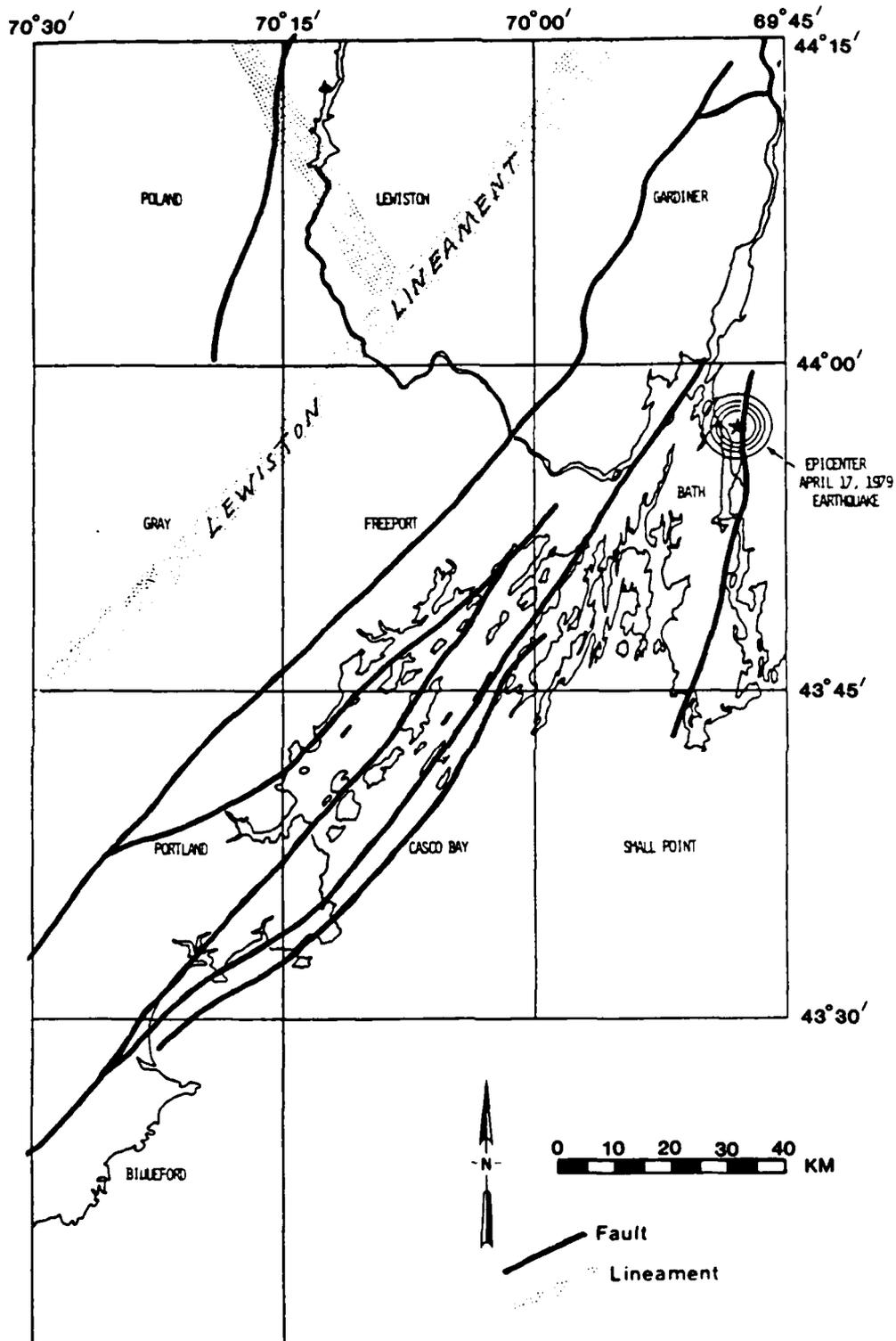
Figure 70. Map showing known or inferred major faults in the Lower
Androscoggin River-Casco Bay area, Maine, and epicenter of the
April 17, 1979, earthquake (Barosh, 1982b, Fig. 62)
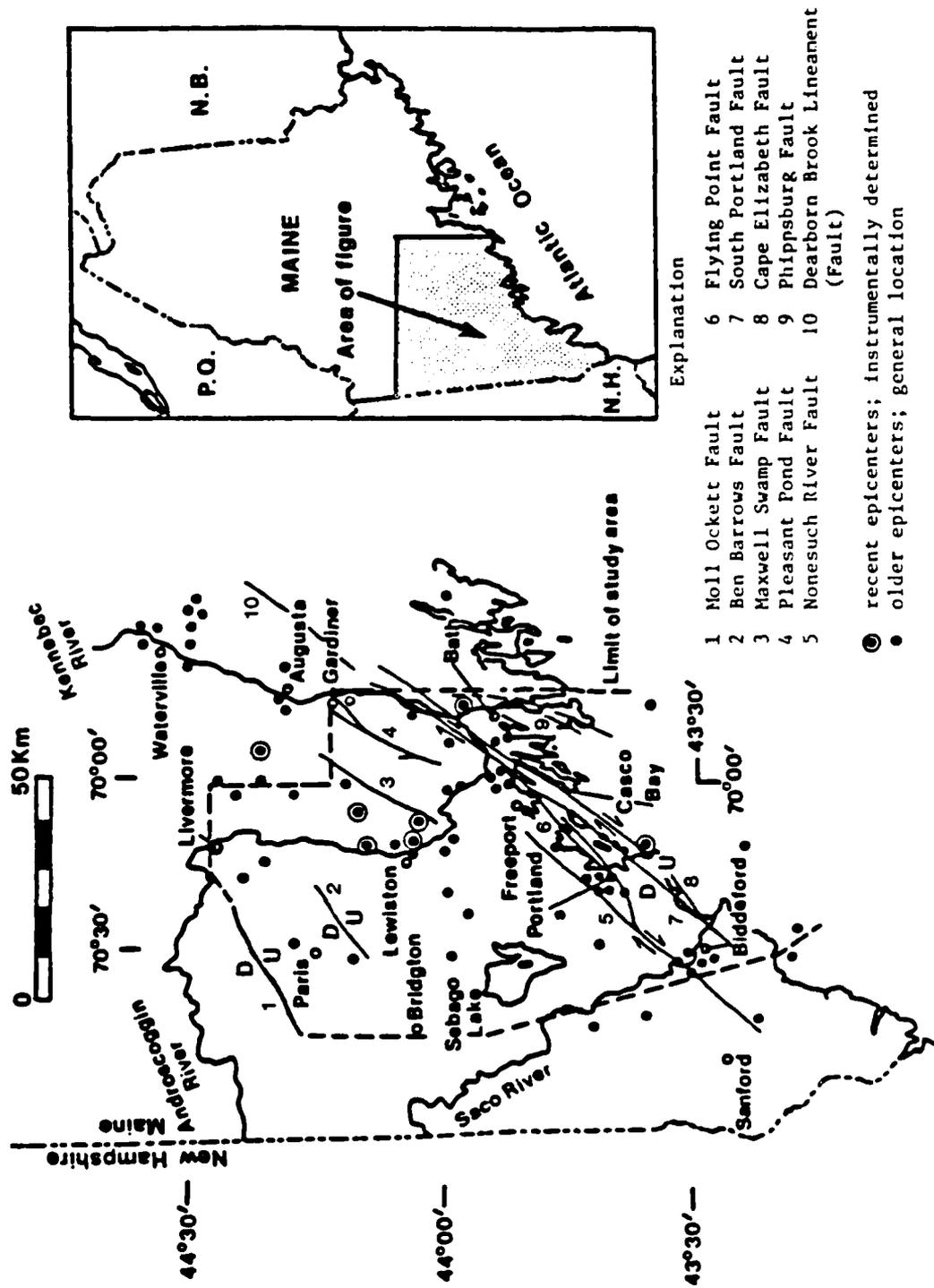
Figure 71. Locations of earthquake epicenters (1776-1979) and post metamorphic faults in the lower Androscoggin Valley, Casco Bay area, Maine (Hussey, 1983, Fig. 57)
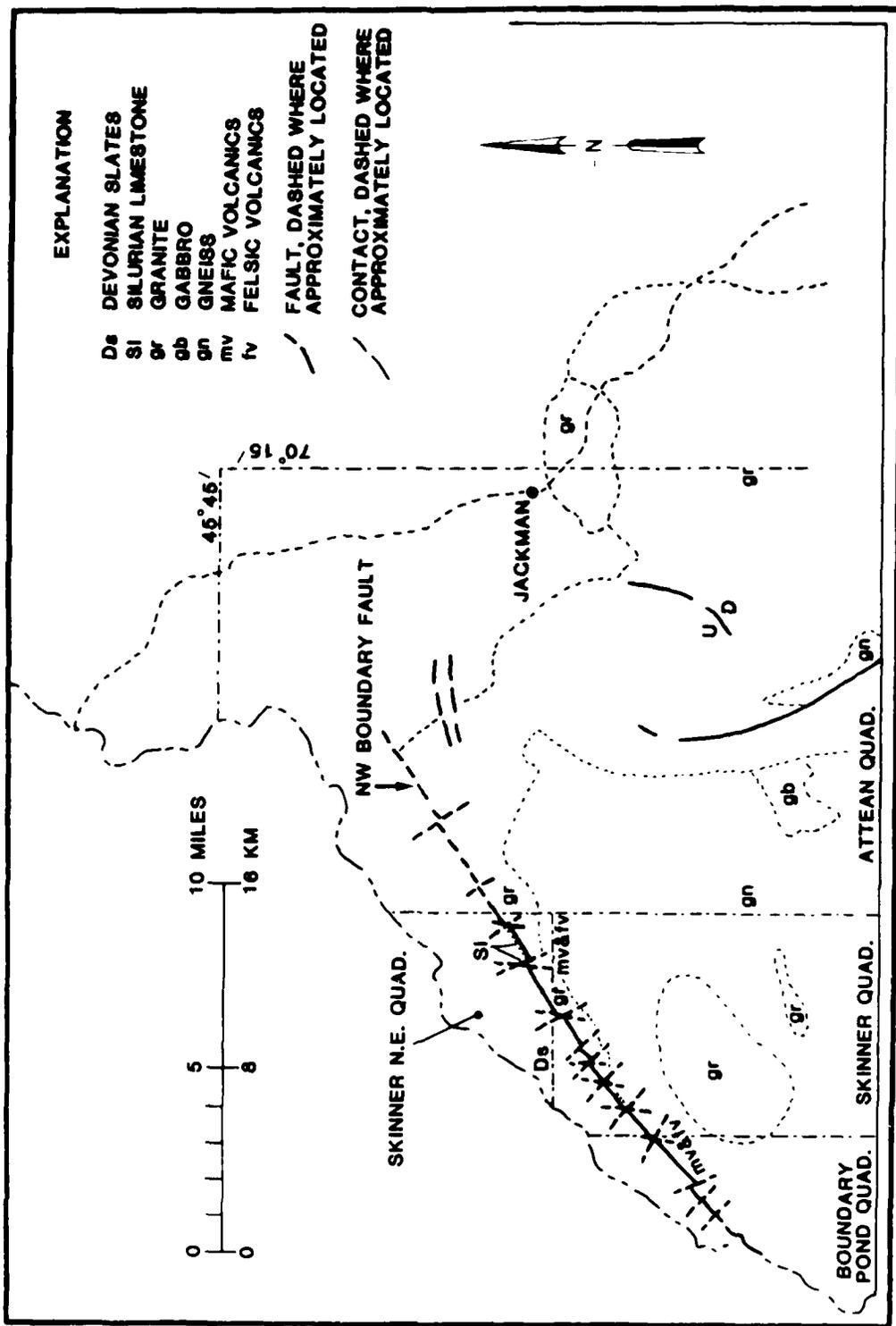
Figure 72. Sketch map of the Northwest Boundary Fault of the Boundary
Mountain Anticlinorium (Westerman, 1981a, Fig. 41)
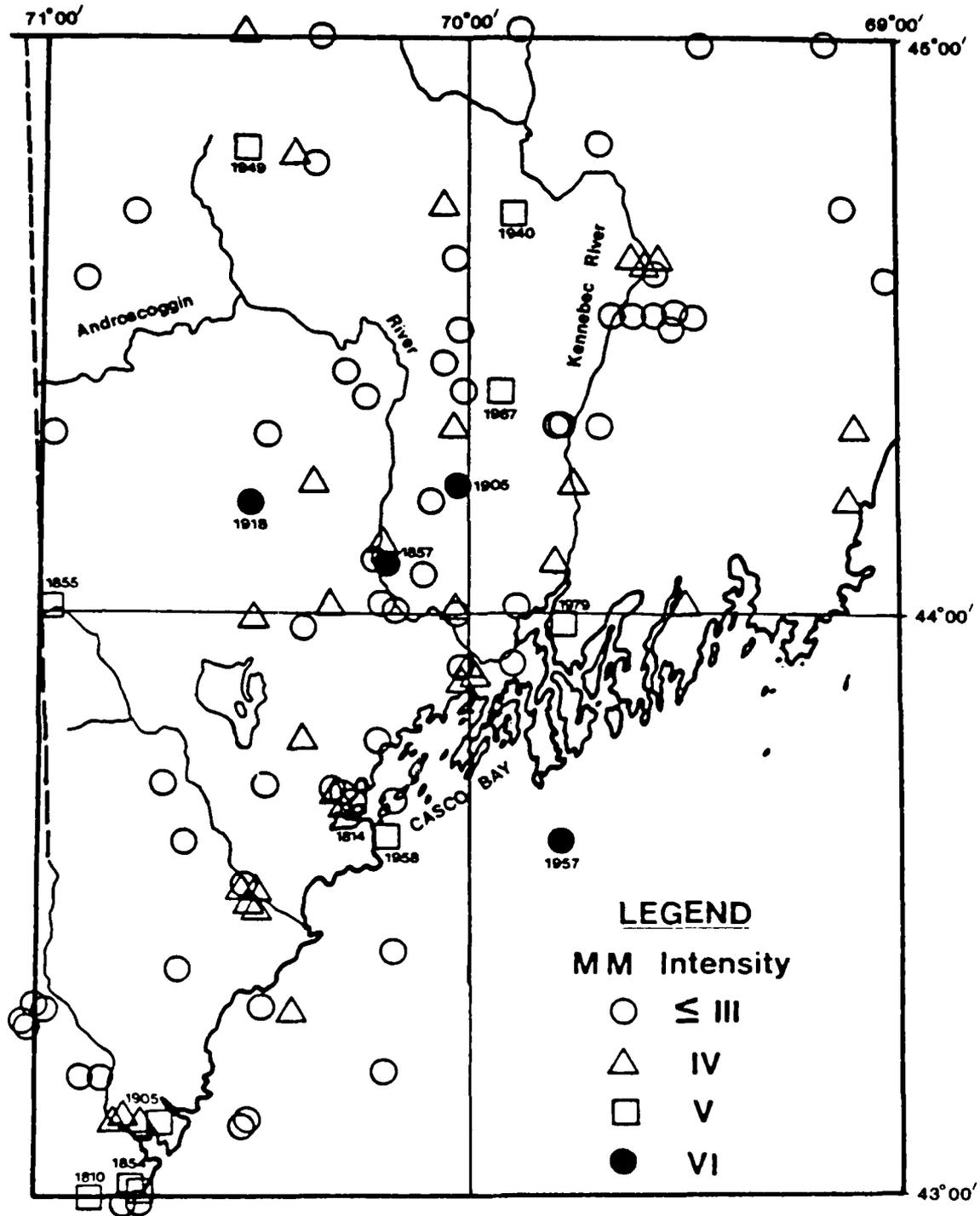
Figure 73. Sketch map of Lower Androscoggin River-Casco Bay area, southern coastal Maine, showing locations of epicenters and Modified Mercalli (MM) intensities of recent and historical earthquakes (from Chiburis and others, 1980) (Barosh, 1982b, Fig. 61)
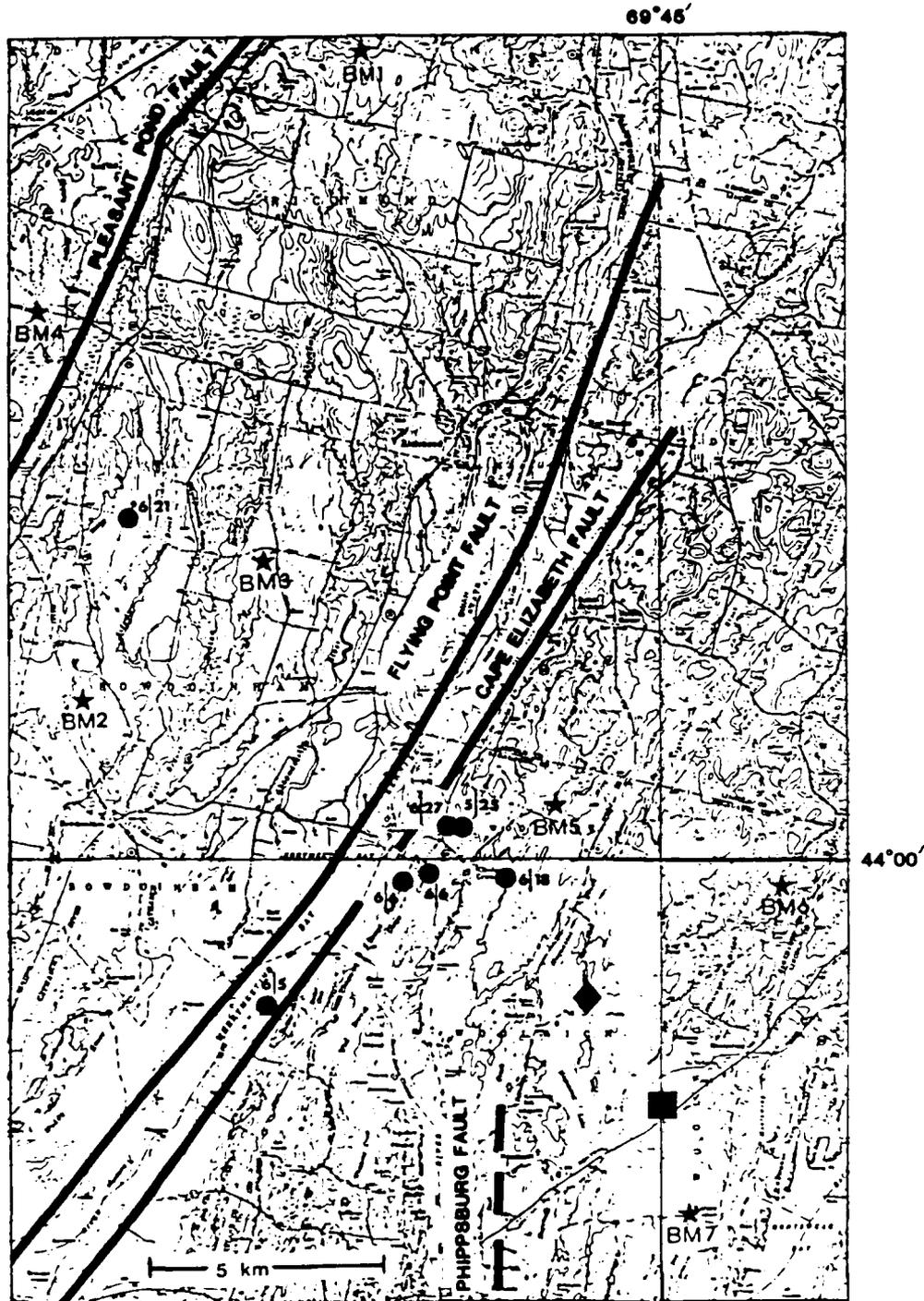
Figure 74. Map of the Merrymeeting Bay area, southwestern Maine, showing location of faults and 1979 earthquakes sequence. Square, mainshock location of Chiburis and Ahner, 1980; Diamond, mainshock location of Ebel, in press a; Dots aftershocks; Stars, portable seismographs (Ebel, in press)

Figure 75. Map of the Dixfield area, western Maine, showing location of
main shock (large diamond) of 1983 earthquake, aftershock
(small diamond), portable seismograph sites (triangles),
geology after Pankiwskyj 1981 and structural lineaments (wavy
lines) (modified after Ebel and McCaffrey, in press)
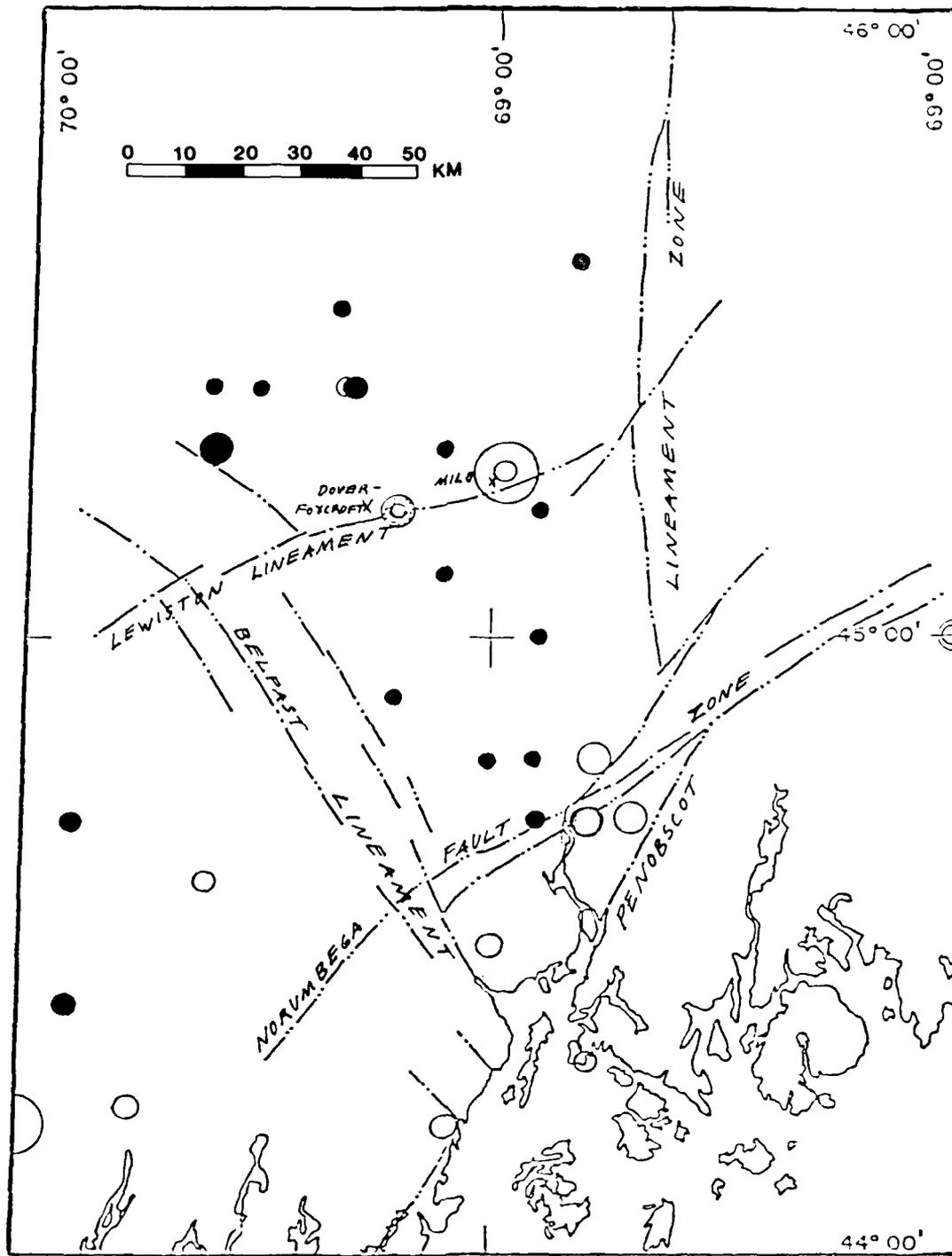
Figure 76. Epicentral map of Penobscot Bay area, Maine, showing selected
LANDSAT Lineaments (epicenters from Boston Edison Co. 1976)
(Barosh, 1978)
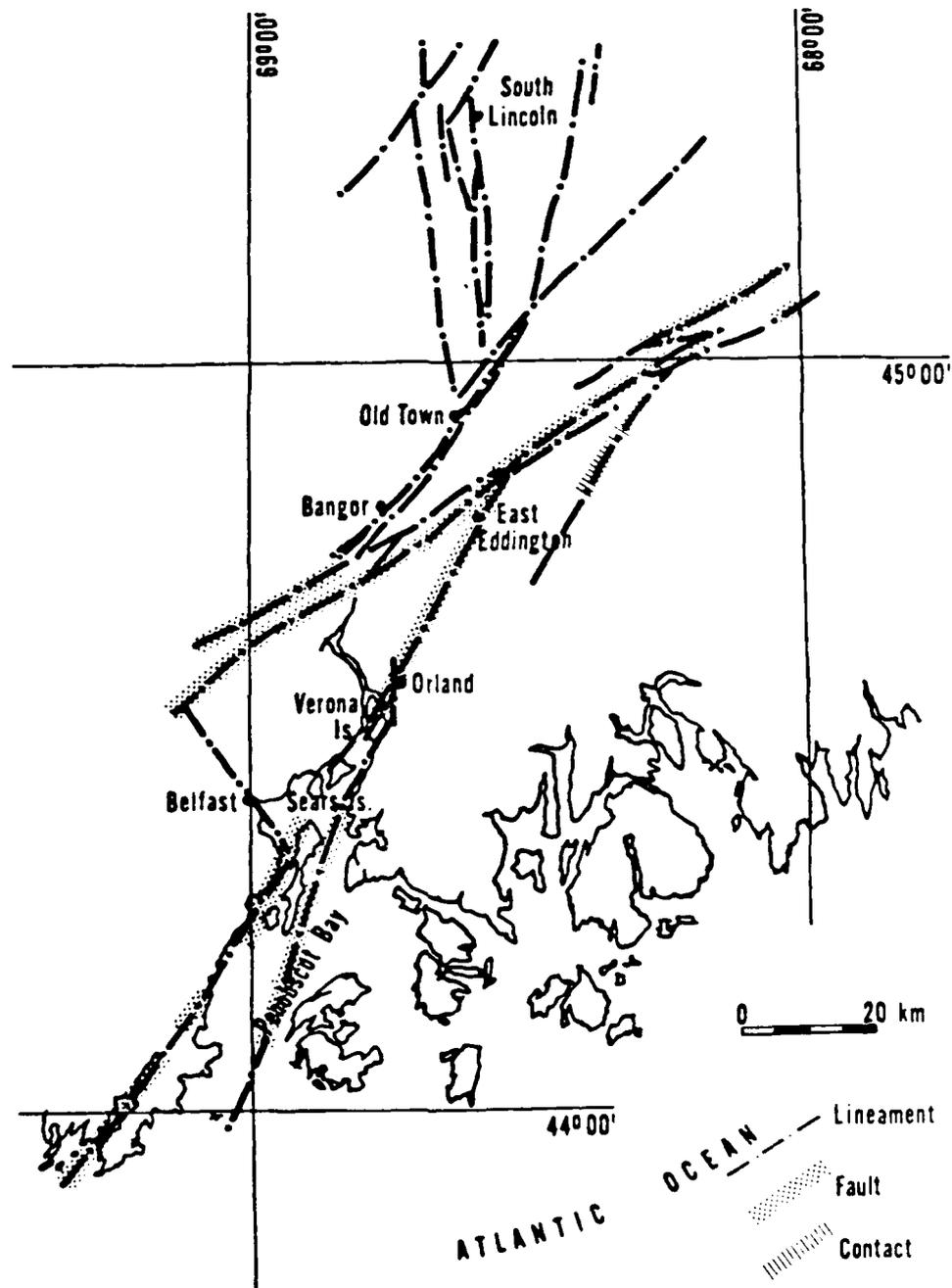
Figure 77. Map showing relation of LANDSAT lineaments with geologic features along the southern portion of the Penobscot lineament zone in the region around Penobscot Bay, Maine (Barosh, 1981b)
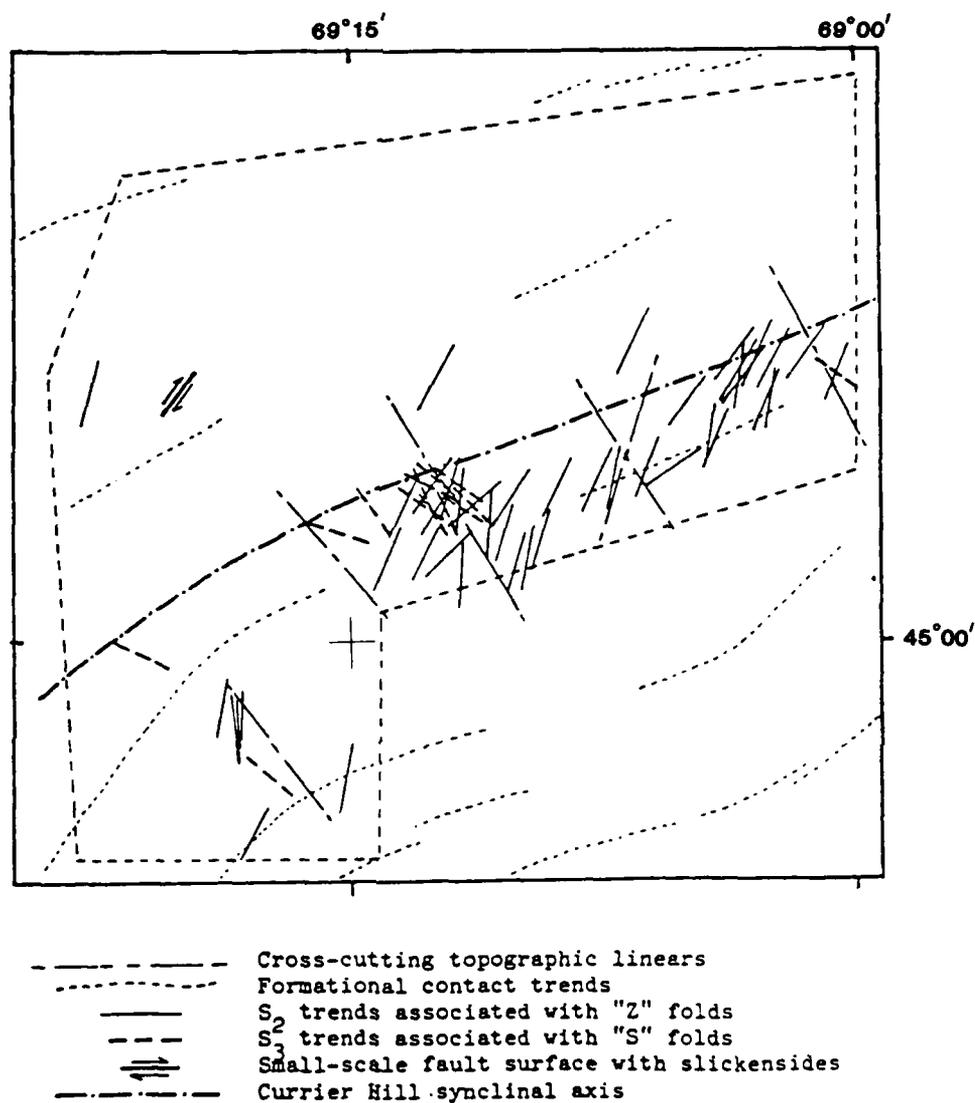
Figure 78. Sketch map of Dover-Foxcroft and Dexter areas, Maine, showing relationship between cross-cutting prominent topographic lineaments and brittle fractures having indications of movement (Westerman, 1981b, Fig. 71)
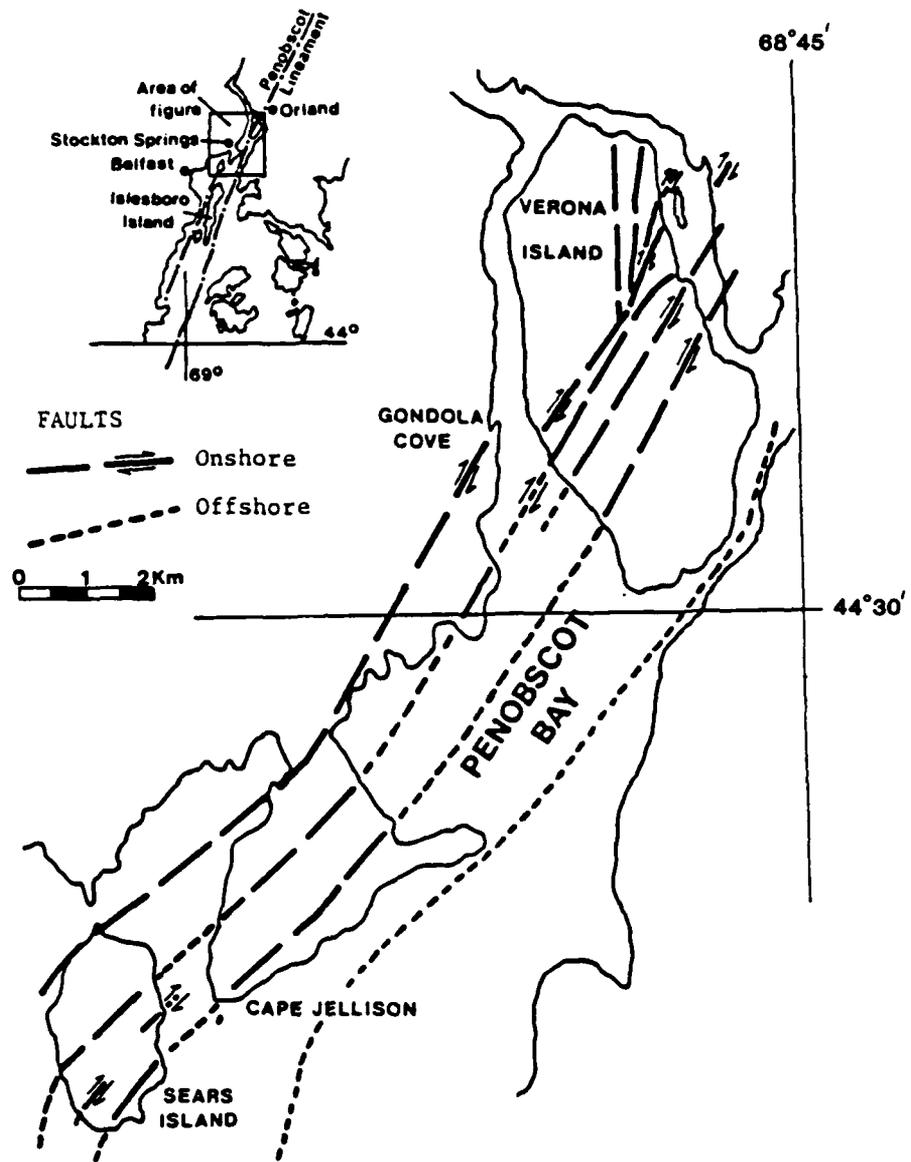
Figure 79. Sketch map of northern Penobscot Bay, Maine, showing approximate position of principal fault zones (Rogers, 1983, Fig. 71)
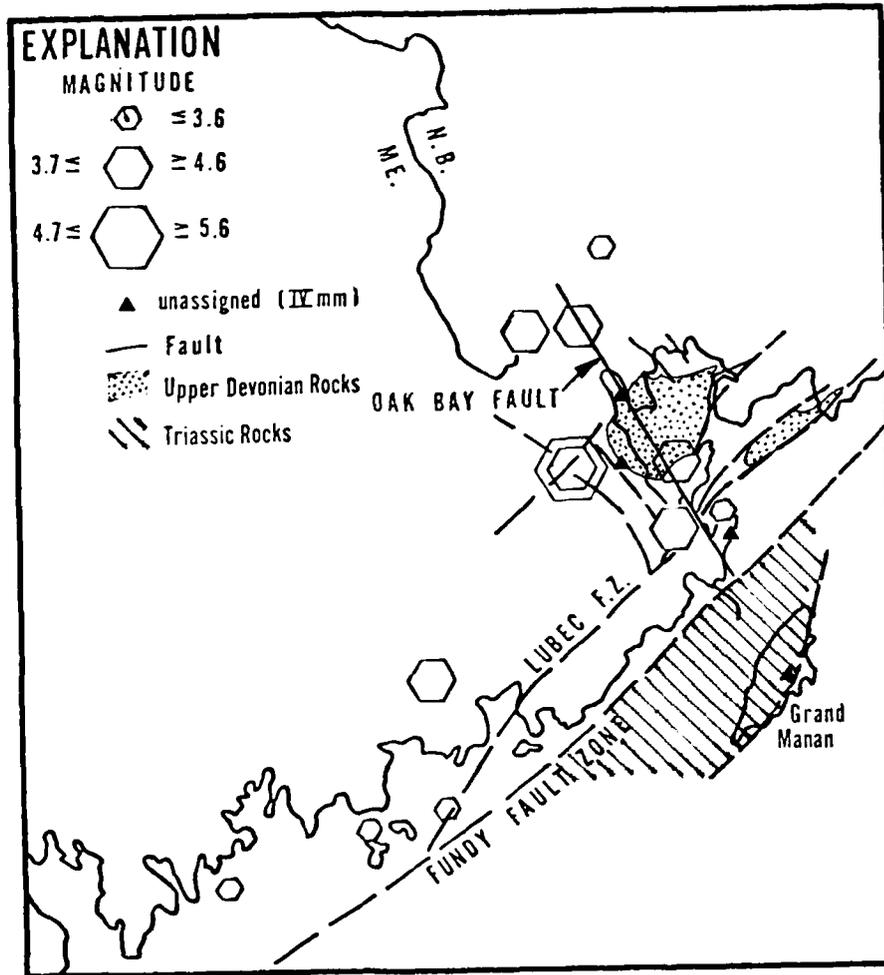
Figure 80. Map of Passamaquoddy Bay area, Maine and New Brunswick, showing relation of earthquake epicenters, selected faults and Devonian Grabens
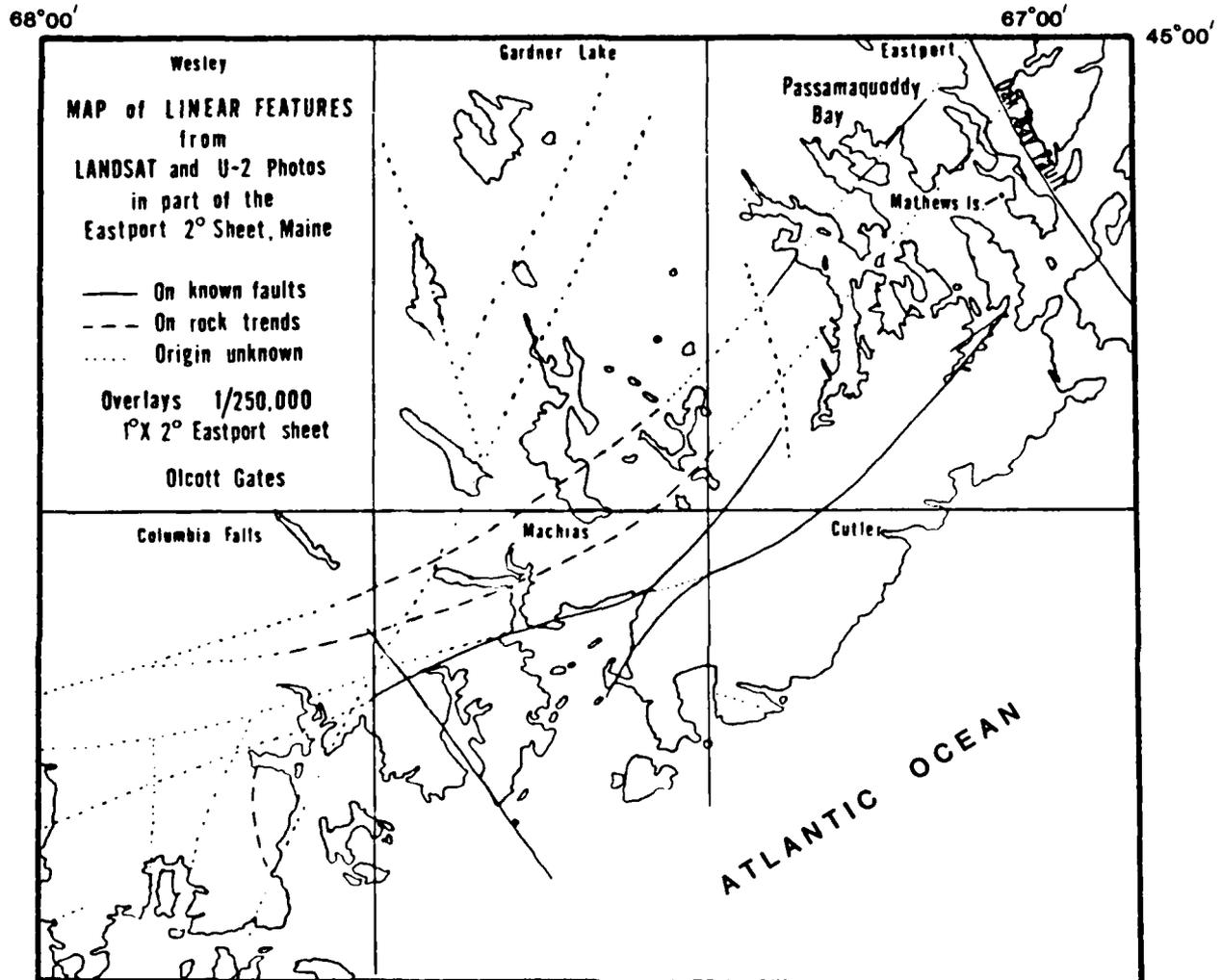
Figure 81. Map of Passamaquoddy Bay and adjacent coastal Maine showing selected faults and lineaments (Gates, 1981, Fig. 46)
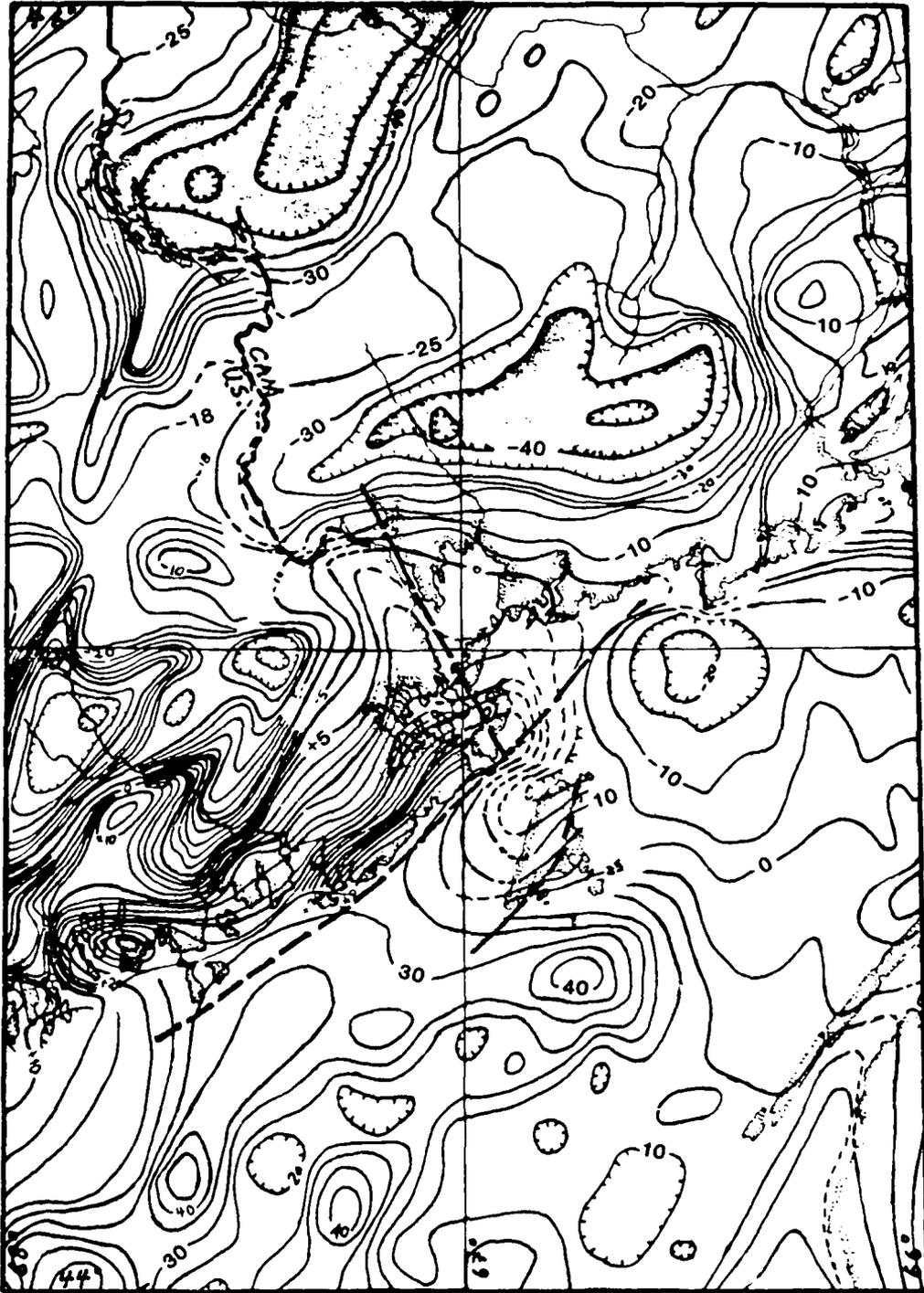
Figure 82. Gravity map of the Passamaquoddy Bay area, Maine and New Brunswick (Hildreth, 1979) and selected faults
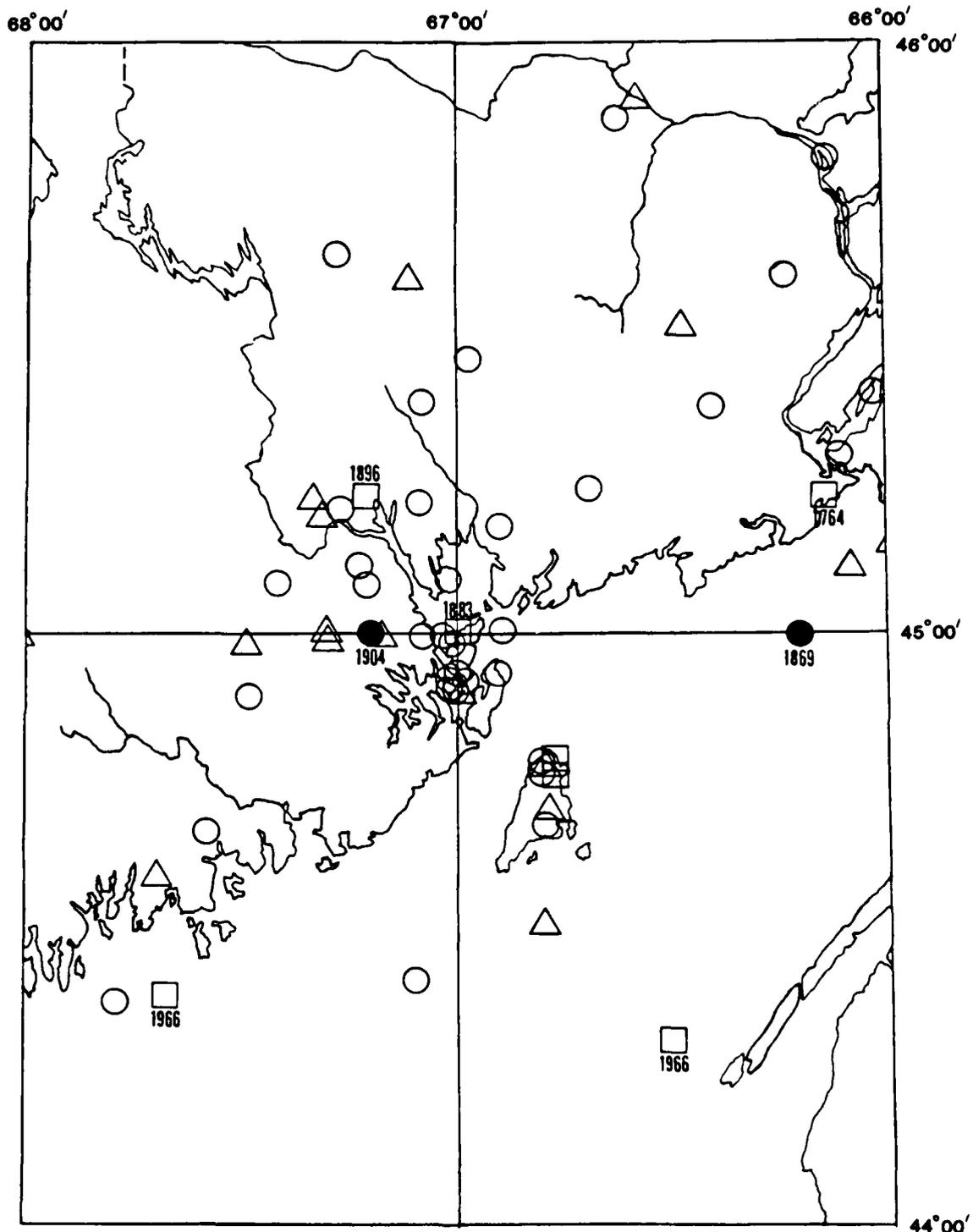
Figure 83. Epicentral map of the Passamaquoddy Bay area, Maine and New Brunswick showing earthquakes through 1980 (from Chiburis and others, 1980 and other data)
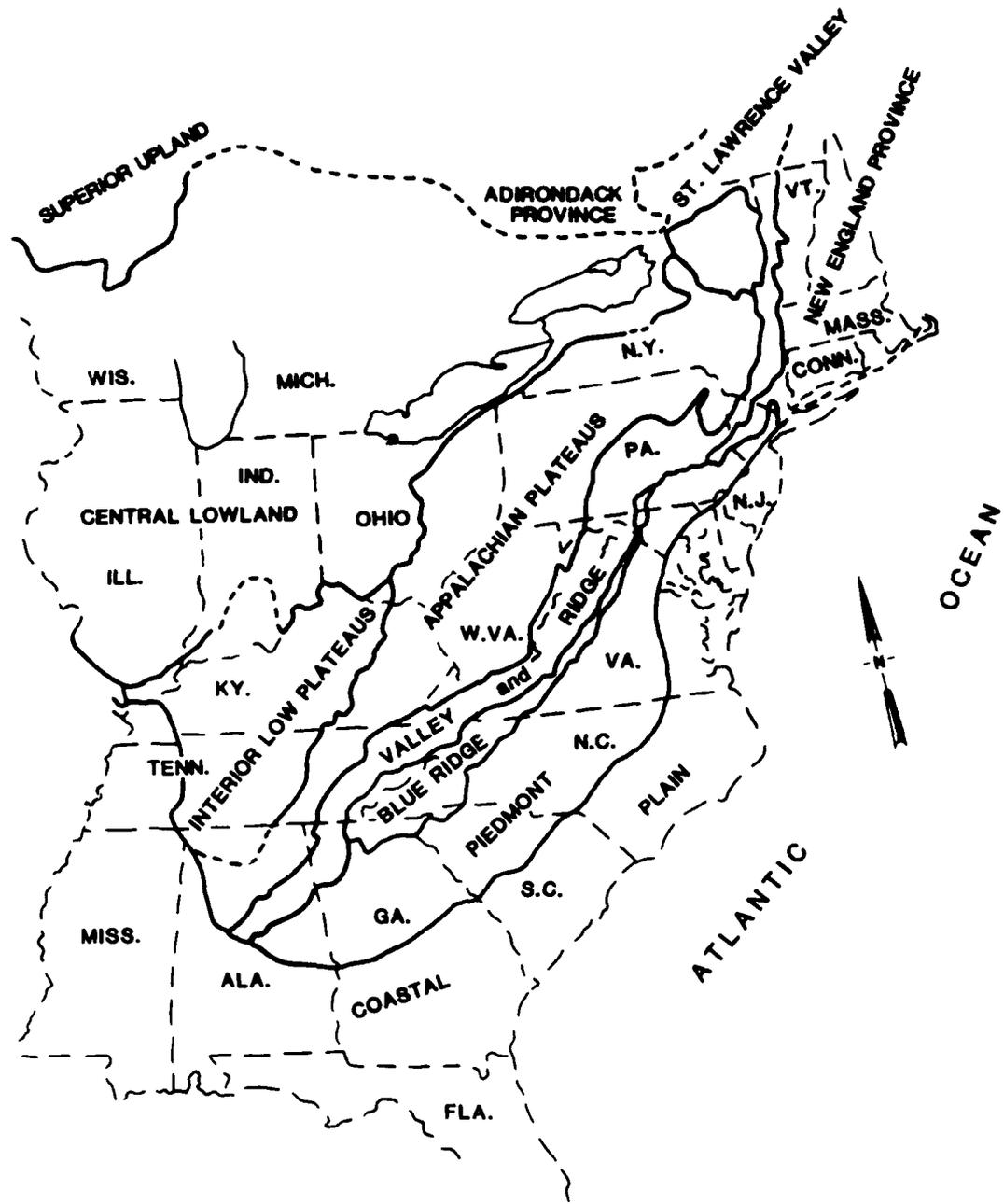
Figure 84. Map showing regional physiographic divisions of the eastern United States (Drumheller and others, 1981, Fig. 3)
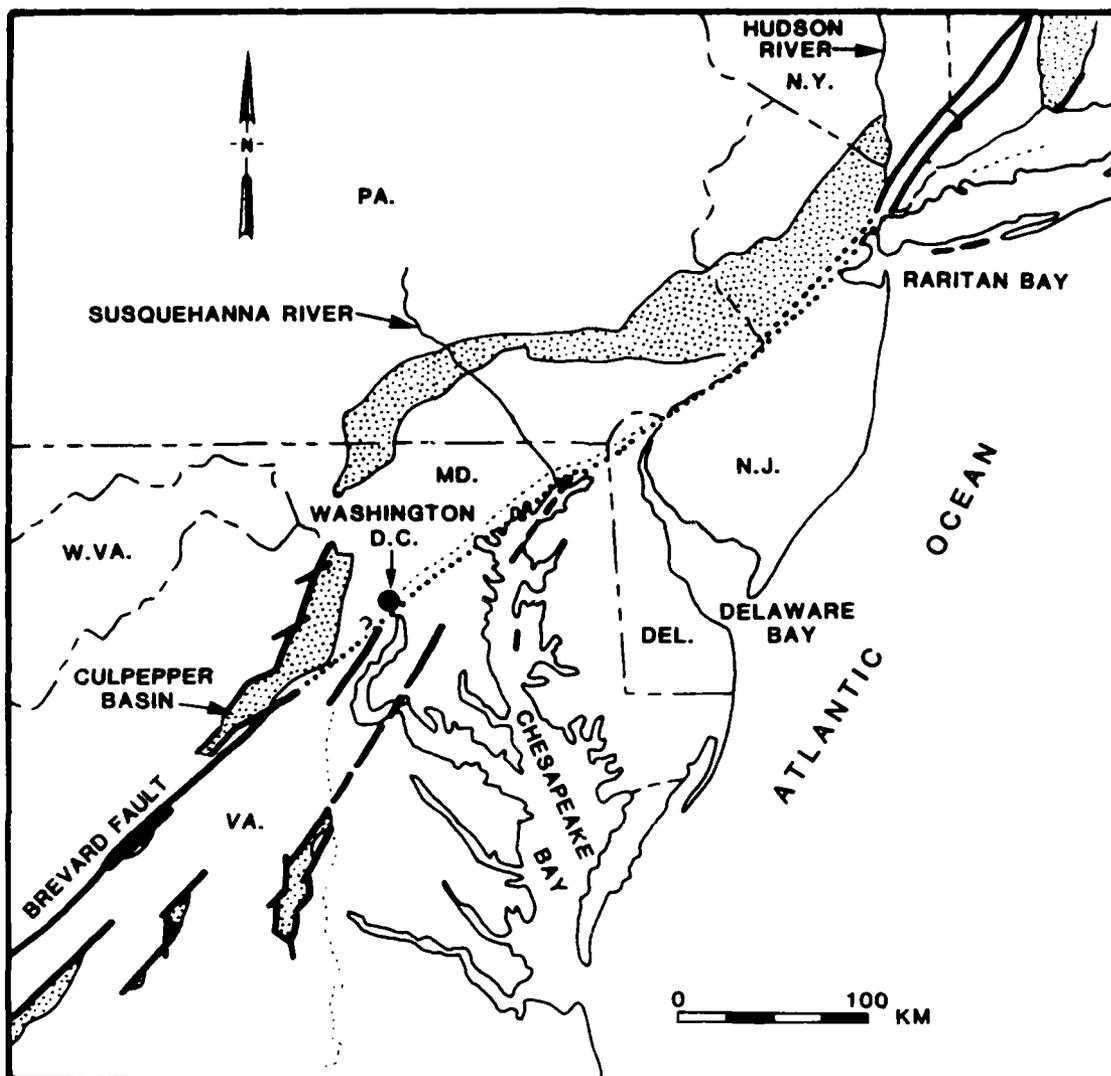
Figure 85. Map of the mid-Atlantic region showing the northern Fall Line Zone (Barosh and Pease, 1979, Fig. 1)

Symbols:

Heavy lines - faults, dashed where interpretive.
Heavy dotted line - probable fault zone, queried where extended by Hobbs (1904) and Moody (1966).
Fine dotted line - contact of Coastal Plain deposits.
Stippled areas - Triassic-Jurassic deposits.
Fault data from Stose, 1928; Jacobeen, 1972; Higgins and others, 1974; Spodjaric, 1973; Lewis and Kummel, 1950; L. Pavlides and R.B. Mixon, personal communication.
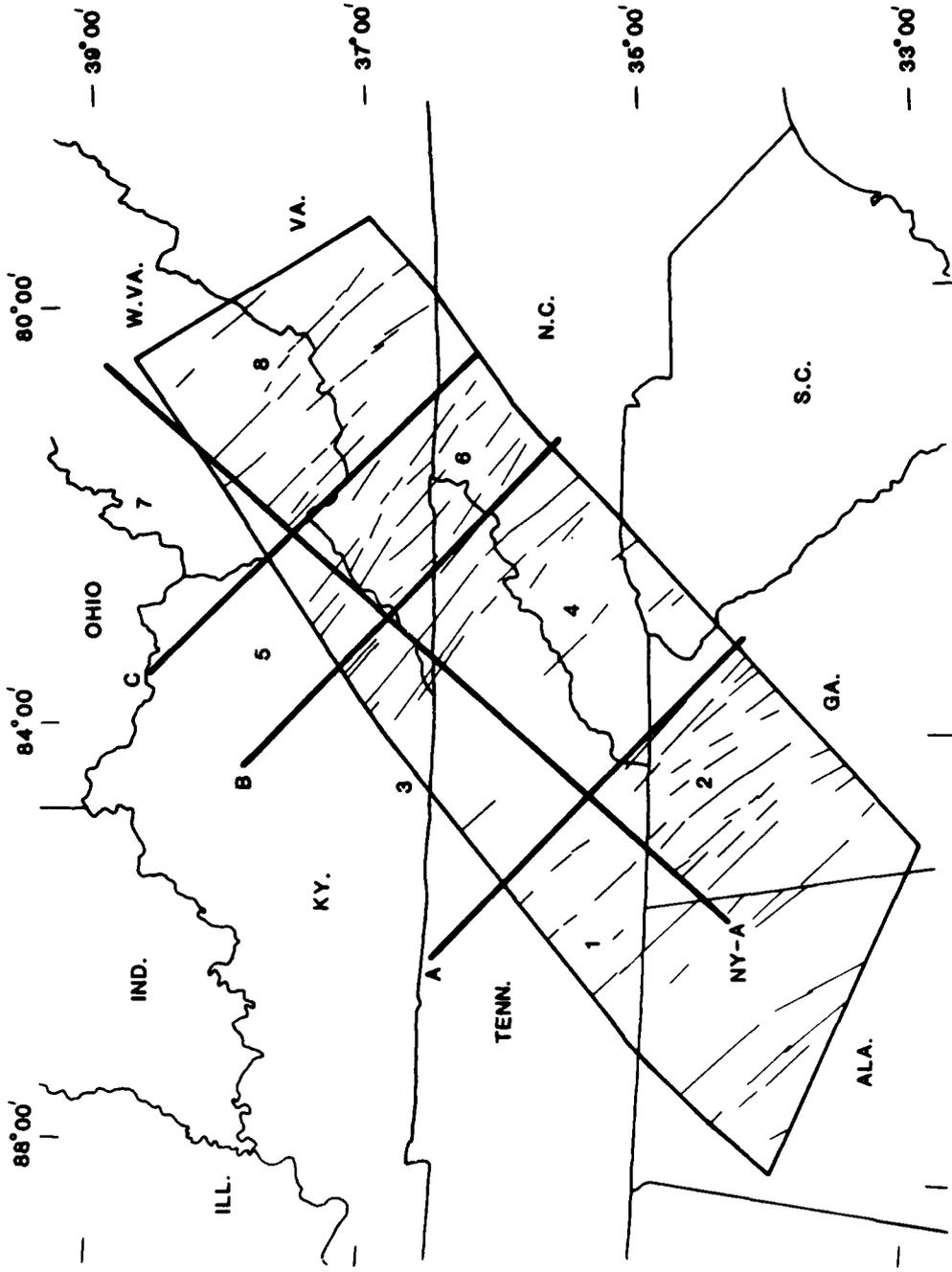
Figure 86. Map of the southern Appalachian Highlands showing tectonic subdivision boundaries and northwest-trending lineaments (Seay and Hopkins, 1981, Fig. 11)
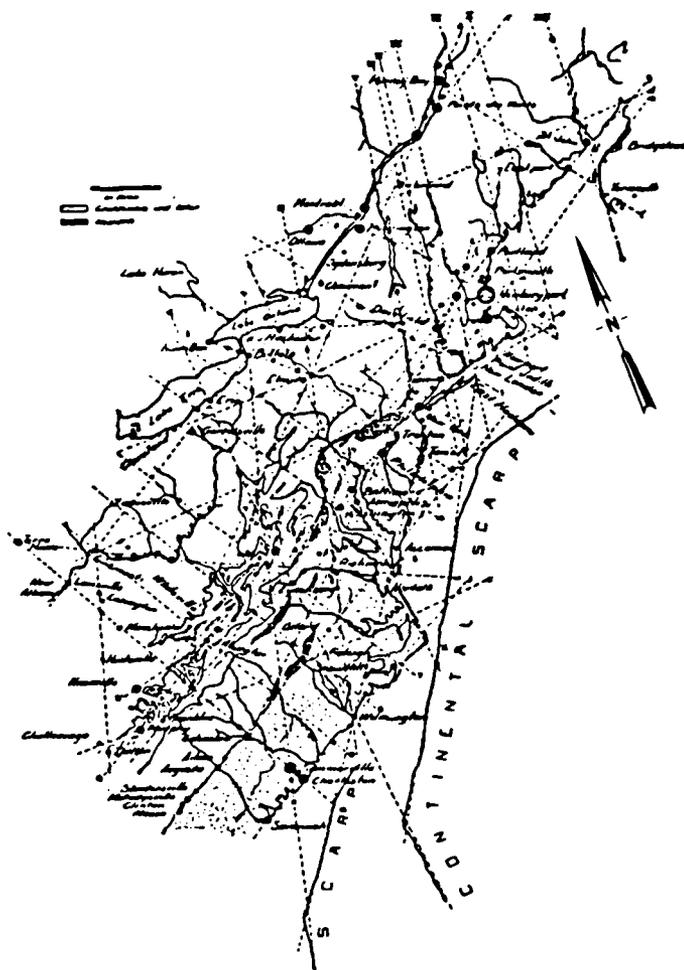
Figure 87. Combined lineament and seismotectonic map of the eastern United States (Hobbs, 1907, Fig. 44)
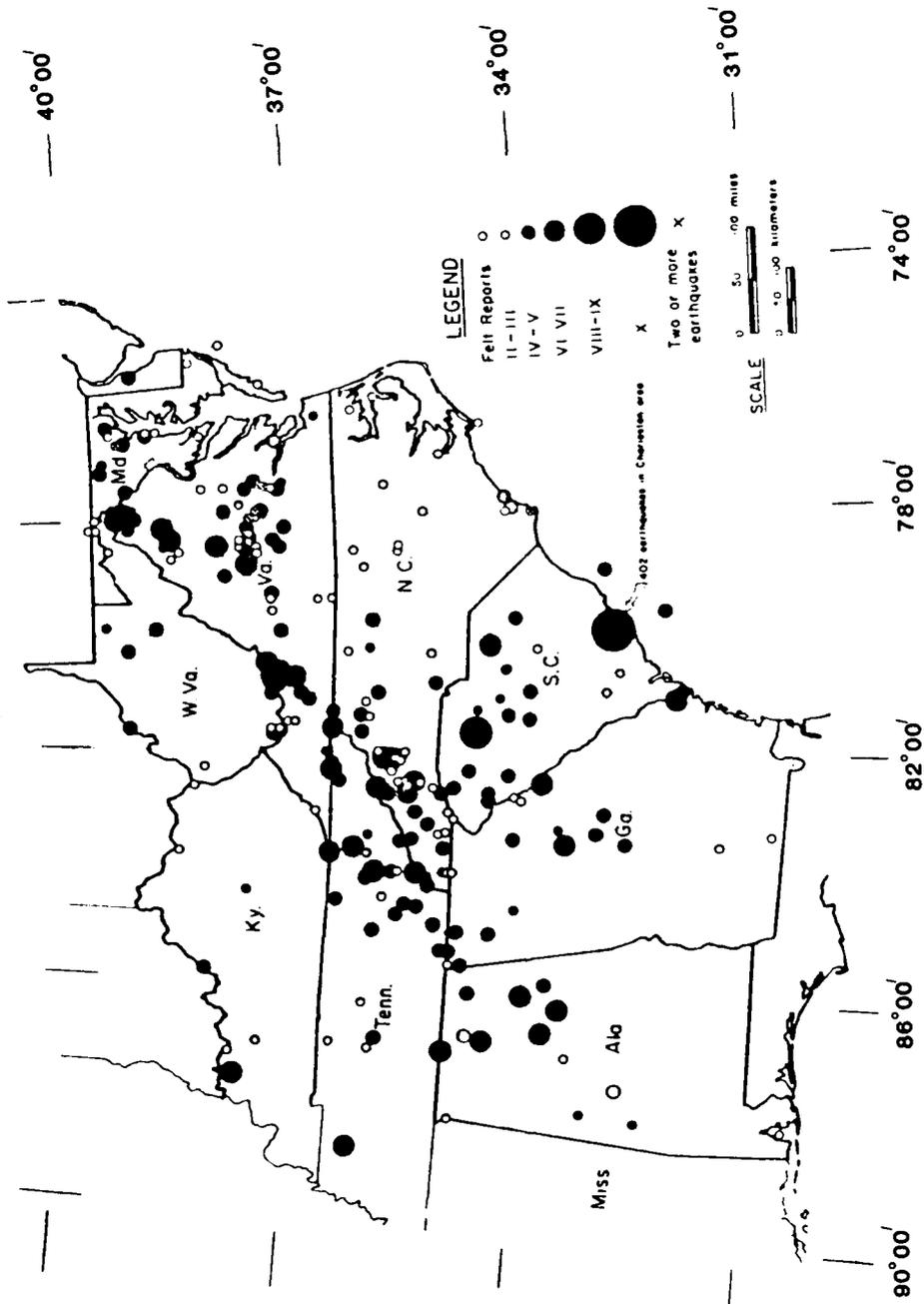
Figure 88. Epicentral map of the southeastern United States showing earthquakes 1754 through 1970 (Bollinger, 1973, Fig. 1)
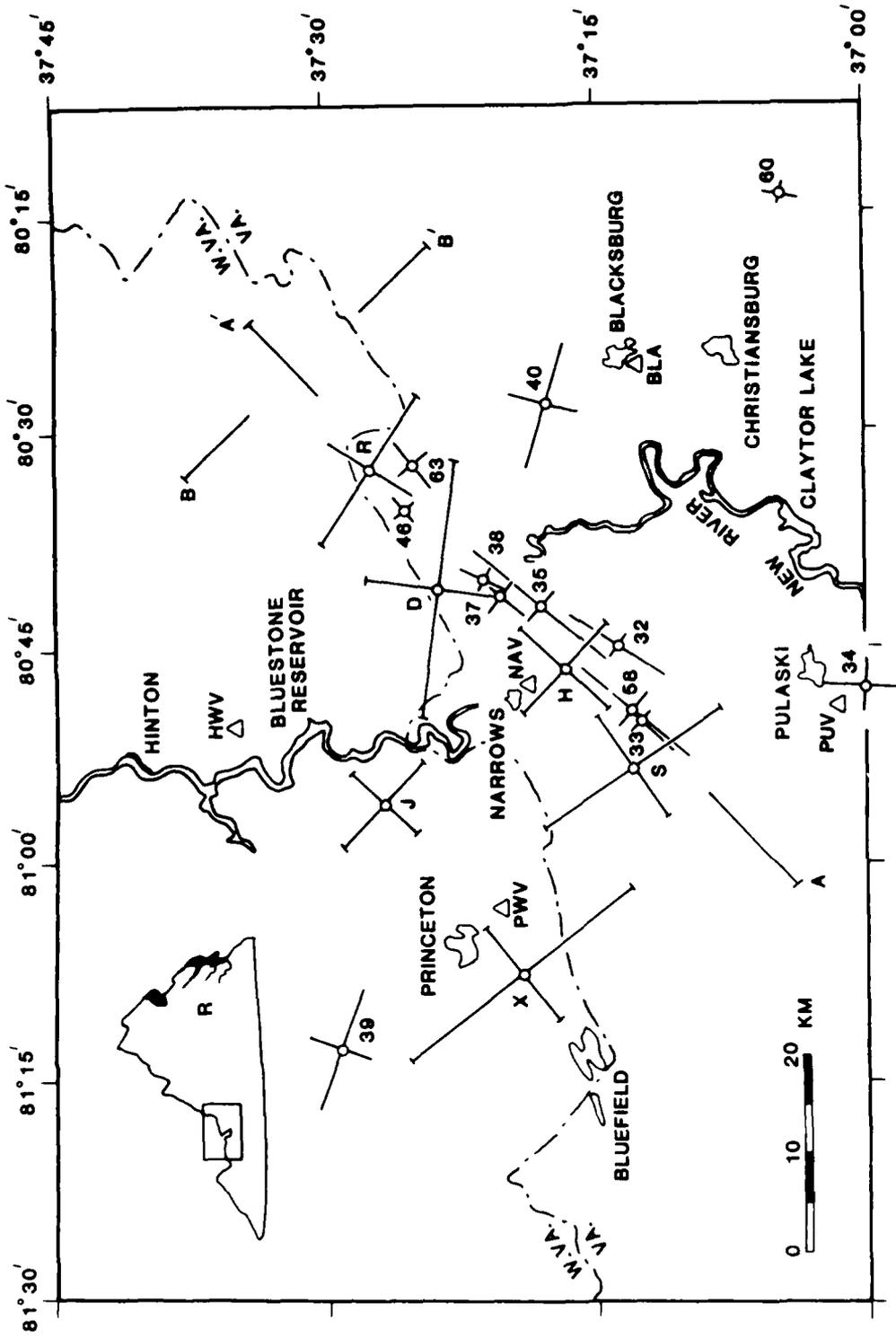
Figure 89. Epicentral map of Giles County, Virginia, and vicinity showing earthquakes 1959 through 1980. Bars indicate area of uncertainty (Bollinger, 1983, Fig. 2)
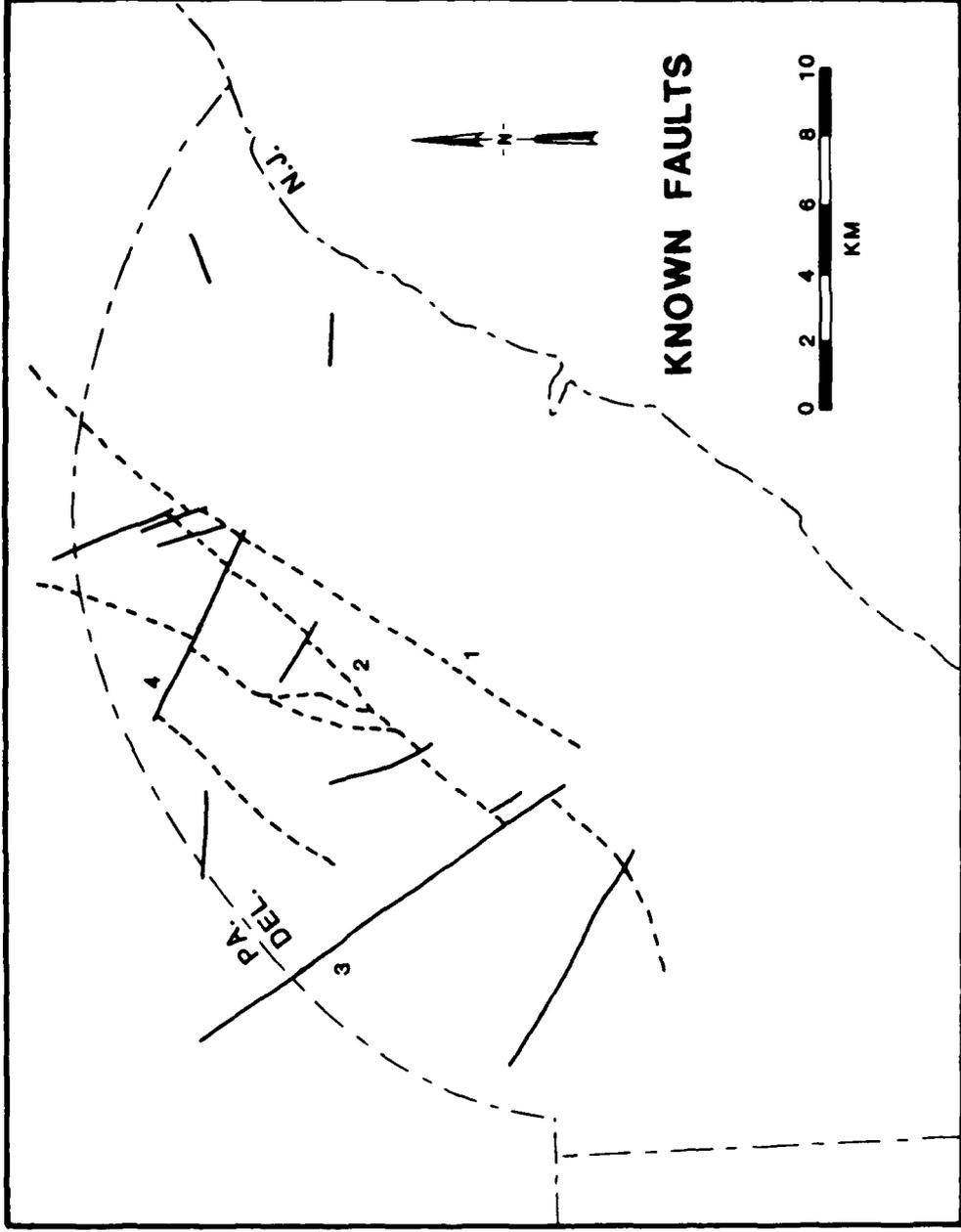
Figure 90. Map showing mapped orogenic (dashed lines) and post-orogenic (solid lines) faults in northern Delaware and adjacent Pennsylvania (Thompson and Hager, 1979, Fig. 2)
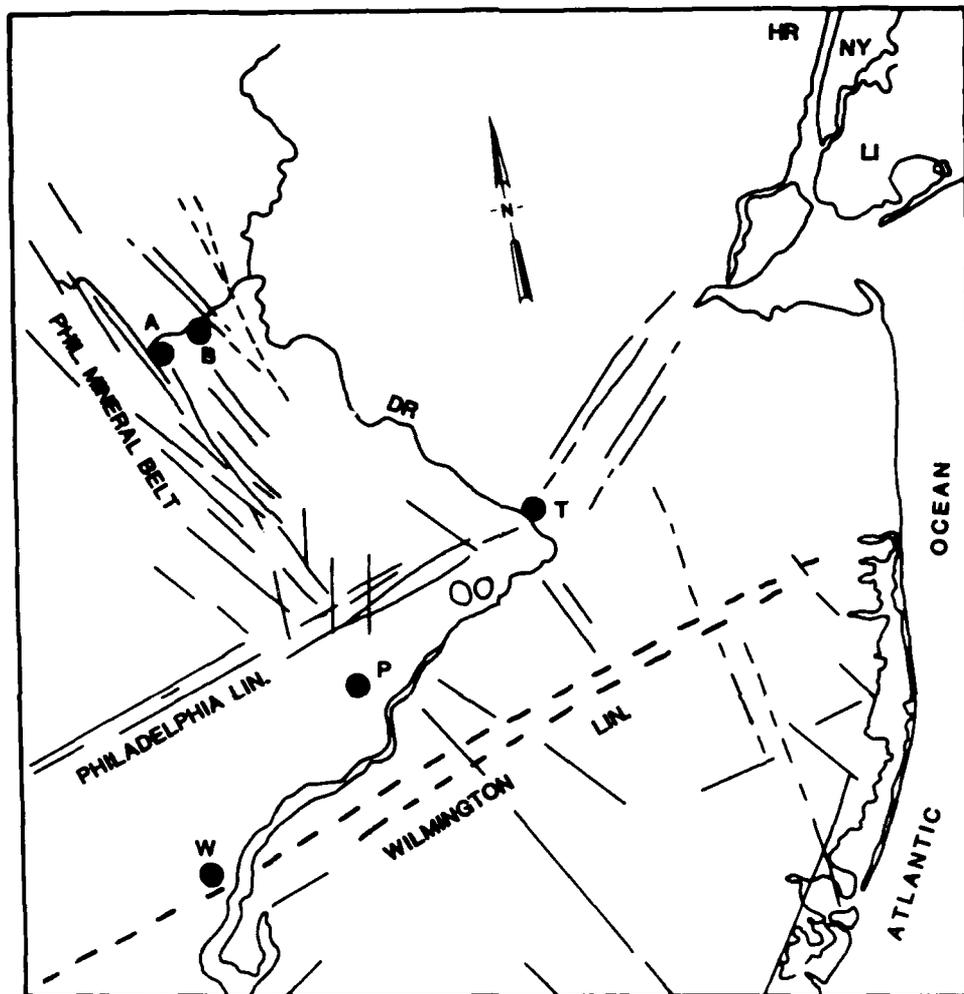
Figure 91. Map of northern New Jersey and adjacent Pennsylvania showing selected LANDSAT lineaments (Gableman, 1979, Fig. 5) HR = Hudson River; NY = New York; LI = Long Island; DR = Delaware River; T = Trenton; B = Bethlehem; A = Allentown; P = Philadelphia; W = Wilmington. Distance across bottom of image is approximately 185 km
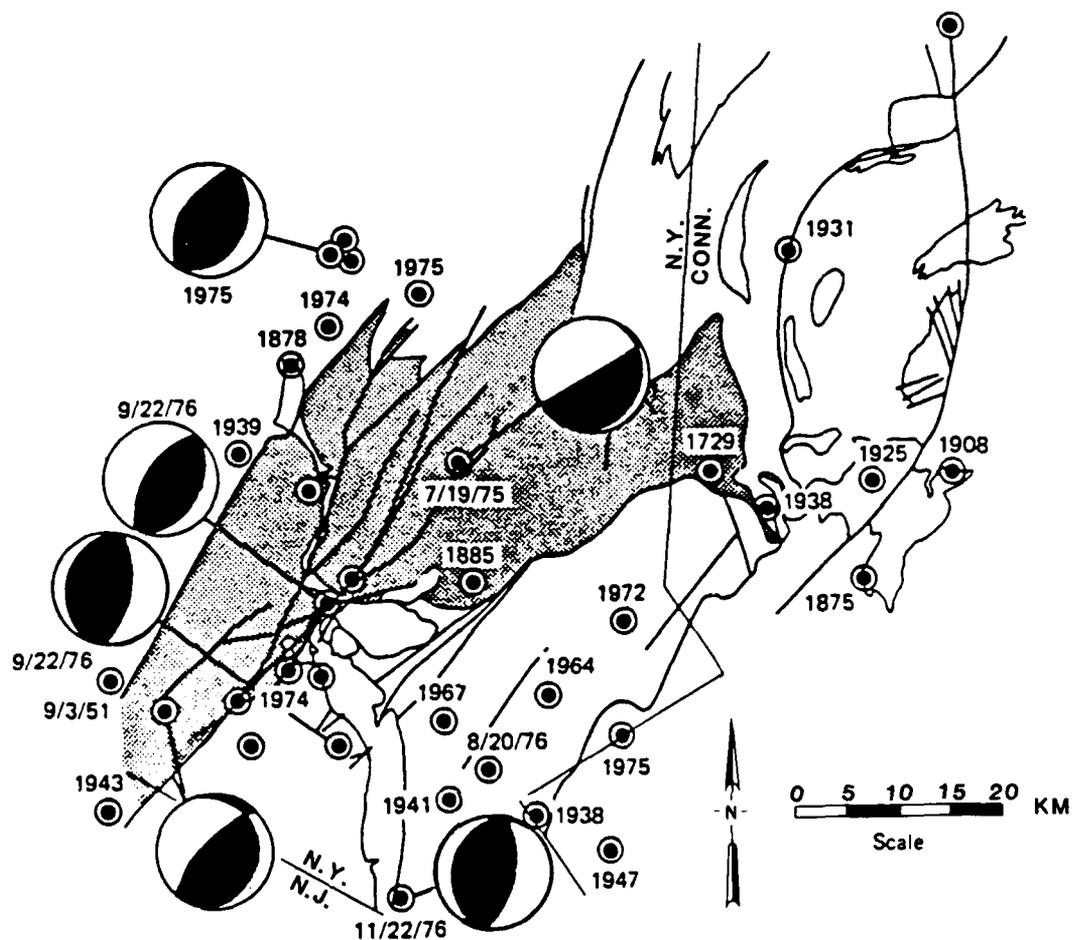
LEGEND

◉ Epicentral locations for historic and measured earthquakes

1729 Year of seismic event

◑ Focal mechanism solution for measured events

▣ Hudson Highlands

Data from Aggarwaal and Sykes (1978); Chandra (1977)

Figure 92. Epicentral map of southwest Connecticut and adjacent New York showing earthquakes through 1976 and selected faults with Mesozoic movement (Tillman, 1982, Fig. 17)

MAJOR FAULT ZONES

OF WESTERN CONNECTICUT,
SOUTHEASTERN NEW YORK,
AND NORTHERN NEW JERSEY
TARGET EXPLORATION INC.

HUDSON RIVER

CORNWELL FAULT

N.Y.
CONN.

GEORGETOWN FAULT

SAW MILL RIVER
FAULT

N.Y.
N.J.

POMPTON
LAKES
FAULTS

RAMAPO

STATEN ISLAND

RARITAN
BAY

MILES

0   5   10   15   20
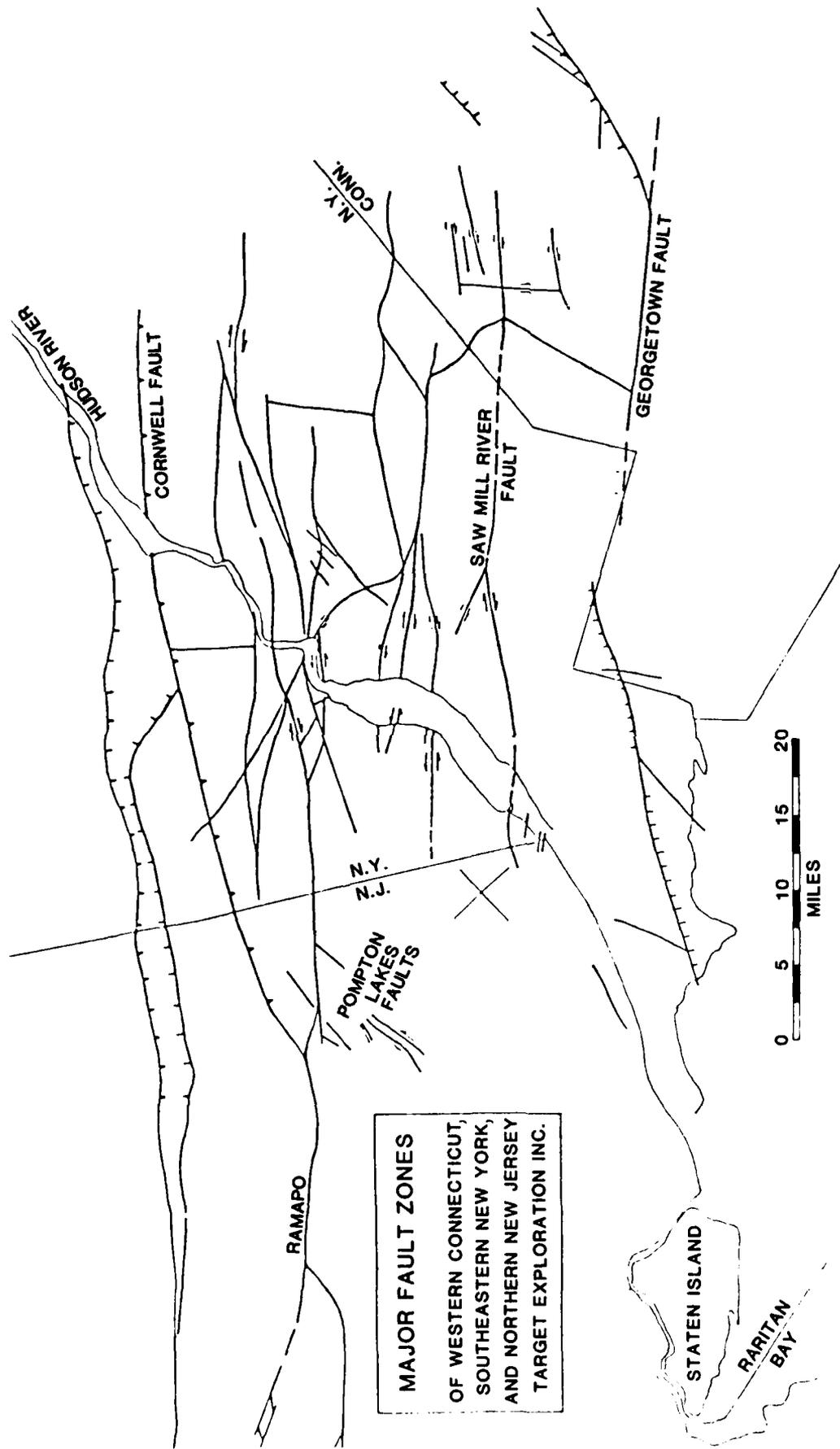
Figure 93. Map showing major fault zones of southwestern Connecticut and southeastern New York (Tillman, 1983)
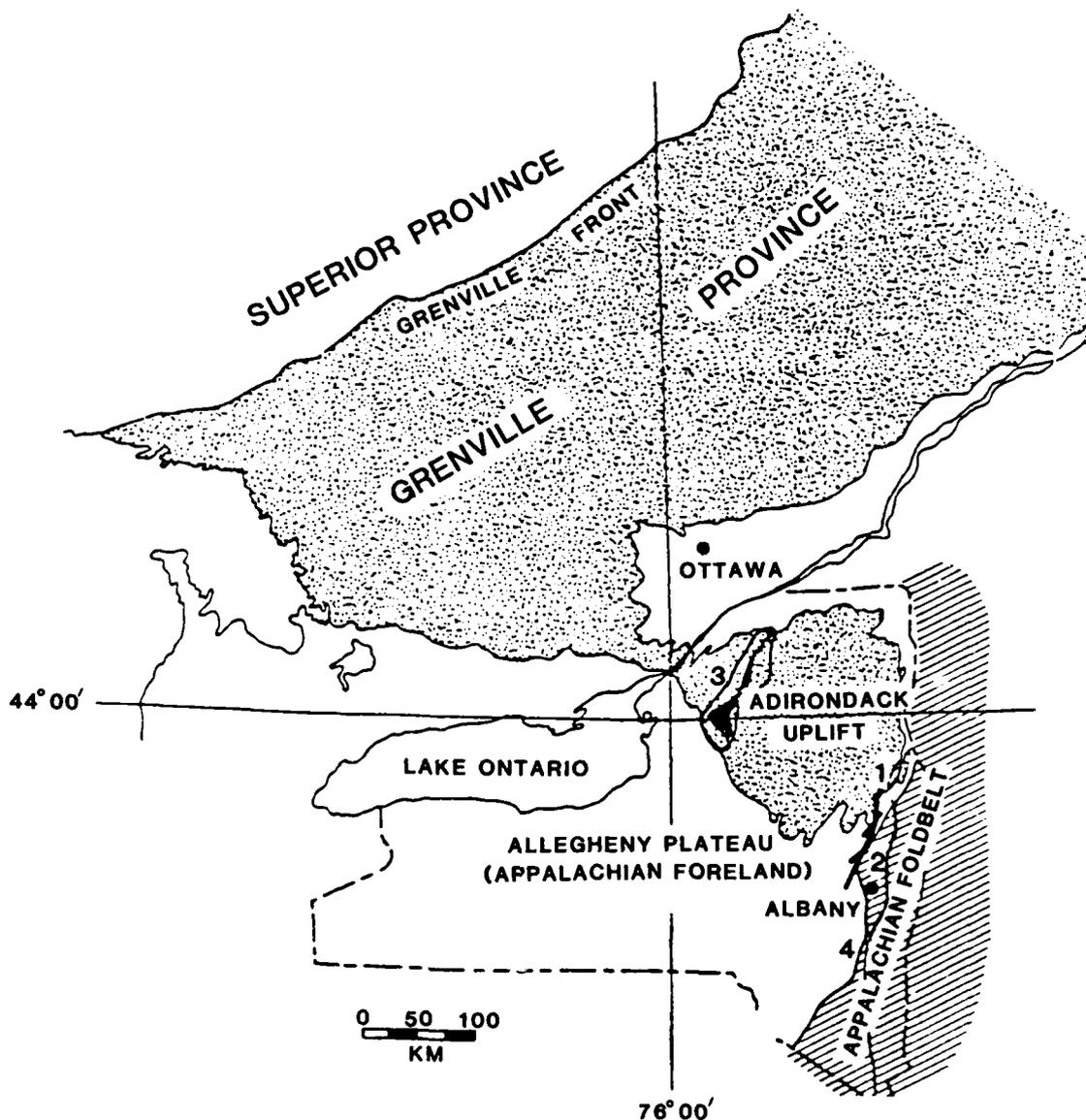
Figure 94. Geologic province map of New York and adjacent Canada (Isachsen, 1982, Fig. 20)
1. Lake George; 2. McGregor-Saratoga-Ballston Lake Fault system (heavy lines); 3. Areas studied (inside black lines) connection with investigation of the Carthage-Colton zone of mylonitization (solid black); 4. Central segment of the Hudson River. Location of studies of exposed faults in northern area of Adirondack Uplift.
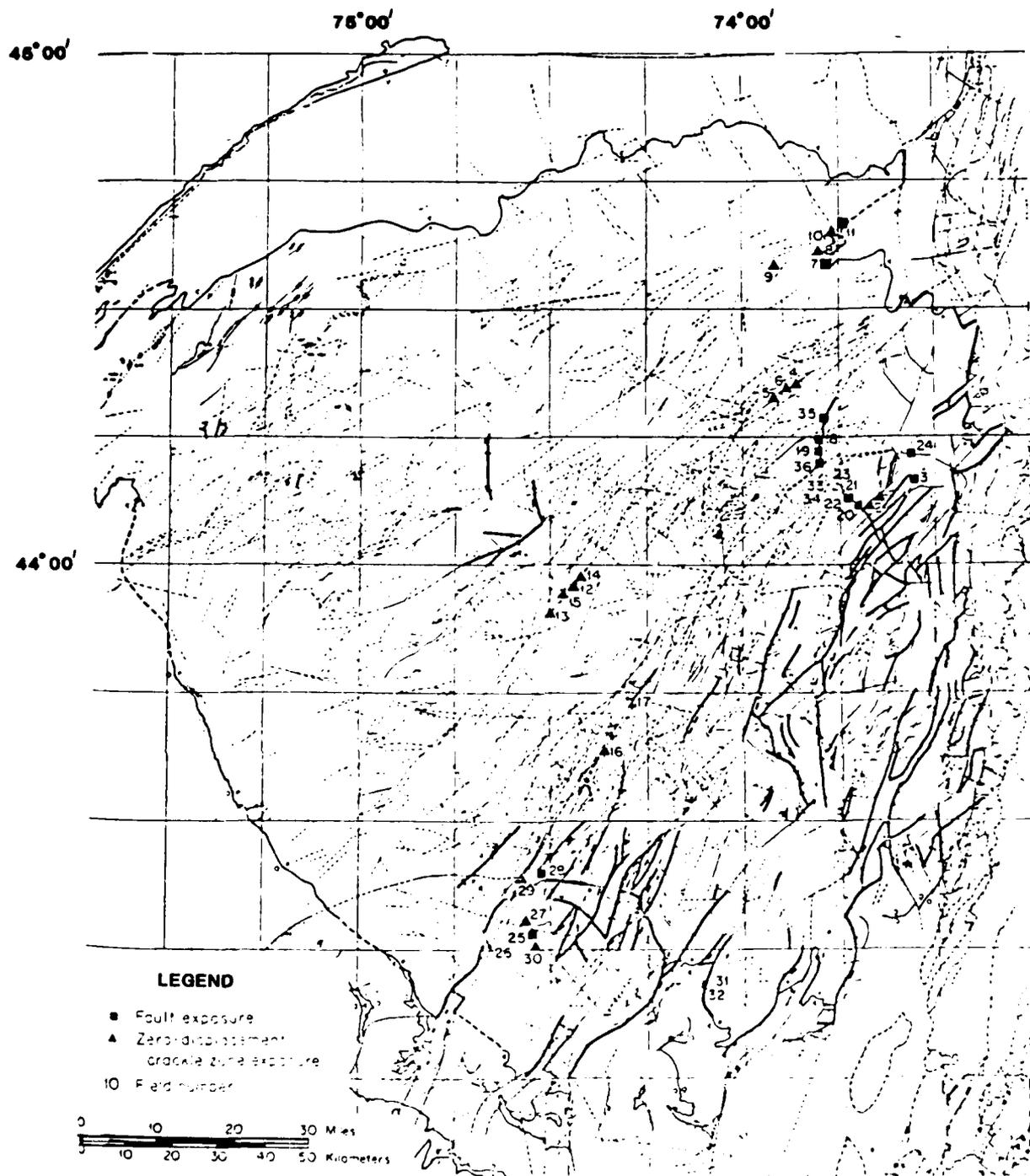
Figure 95. Brittle structures map of the Adirondack Mountains and vicinity
(from Isachsen and McKendree, 1977) (Wiener and Isachsen, 1983,
Fig. 3)
Medium weight line shows unconformity around Adirondack border.
Heavy lines represent high angle faults, within and bordering
the Adirondacks, for which displacement, movement sense or
breccia localities are known. Dashed lines are linear valleys
(lineaments) whose bedrock control is not yet studied. Study
areas are identified as faults (solid square) or zero
displacement crackle zones (triangle).
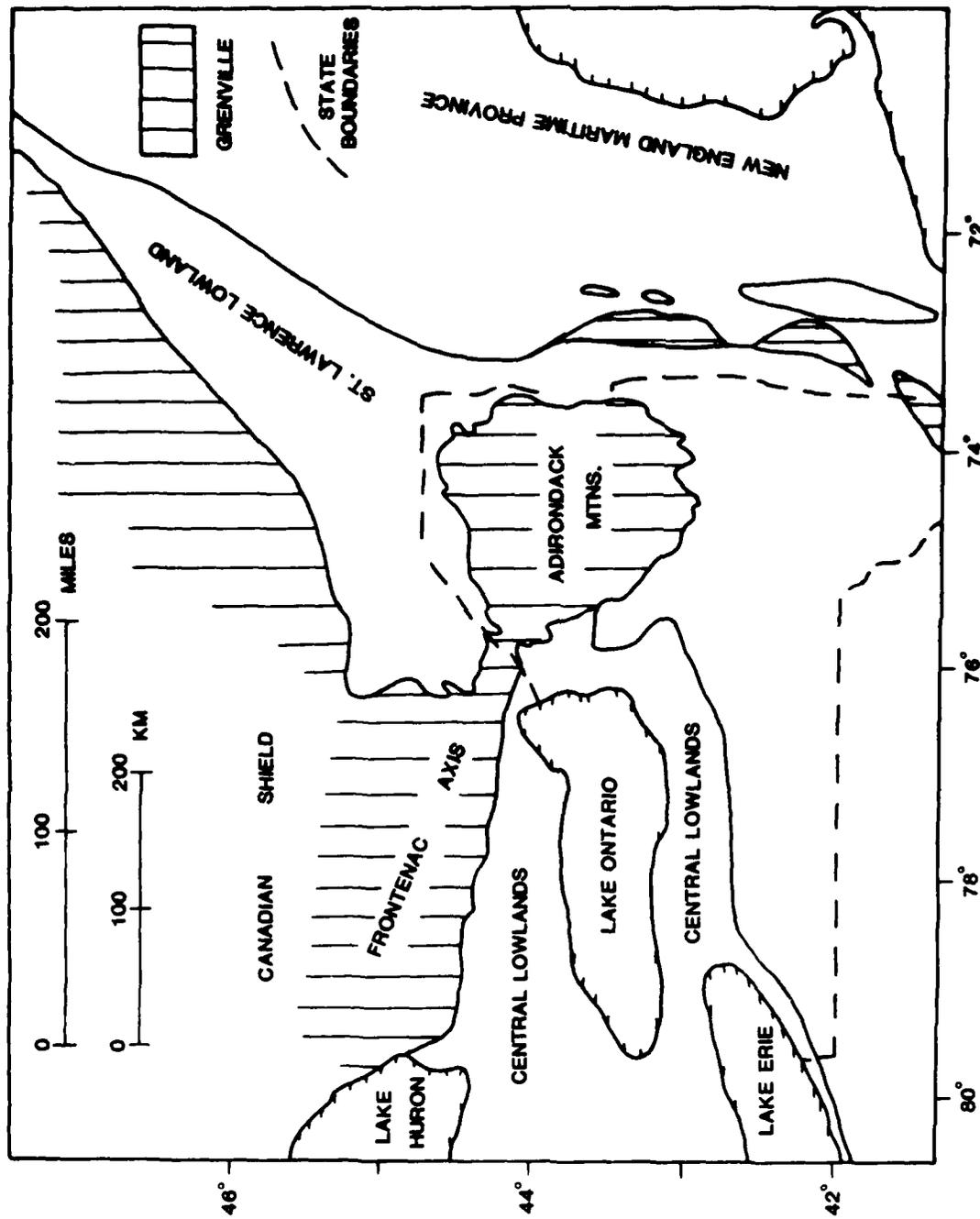
Figure 96. Physiographic map of the St. Lawrence Valley and surrounding region
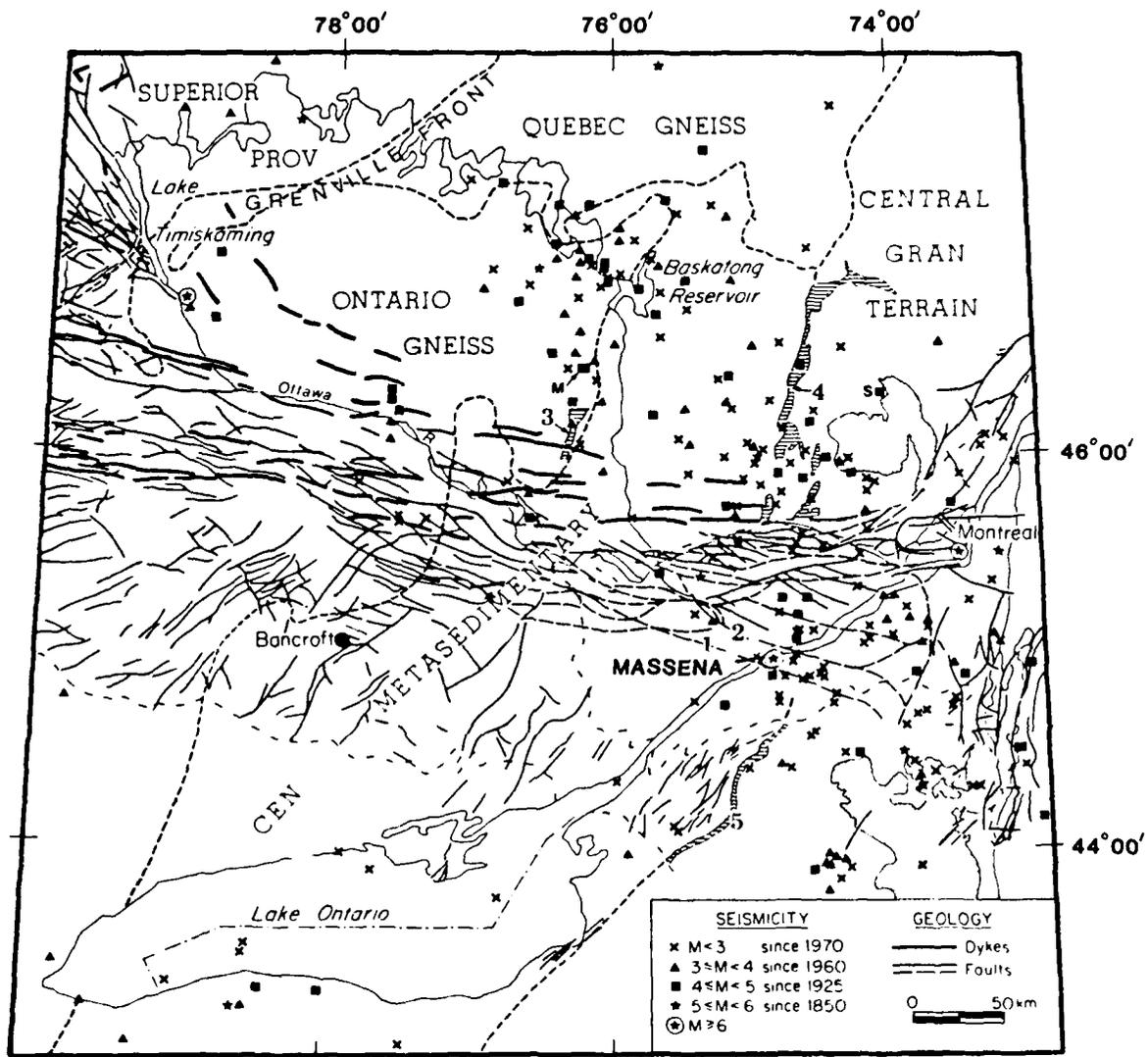
Figure 97. Epicentral map of the Ottawa Valley and surrounding region showing faults and dikes (Forsyth, 1981, Fig. 1)
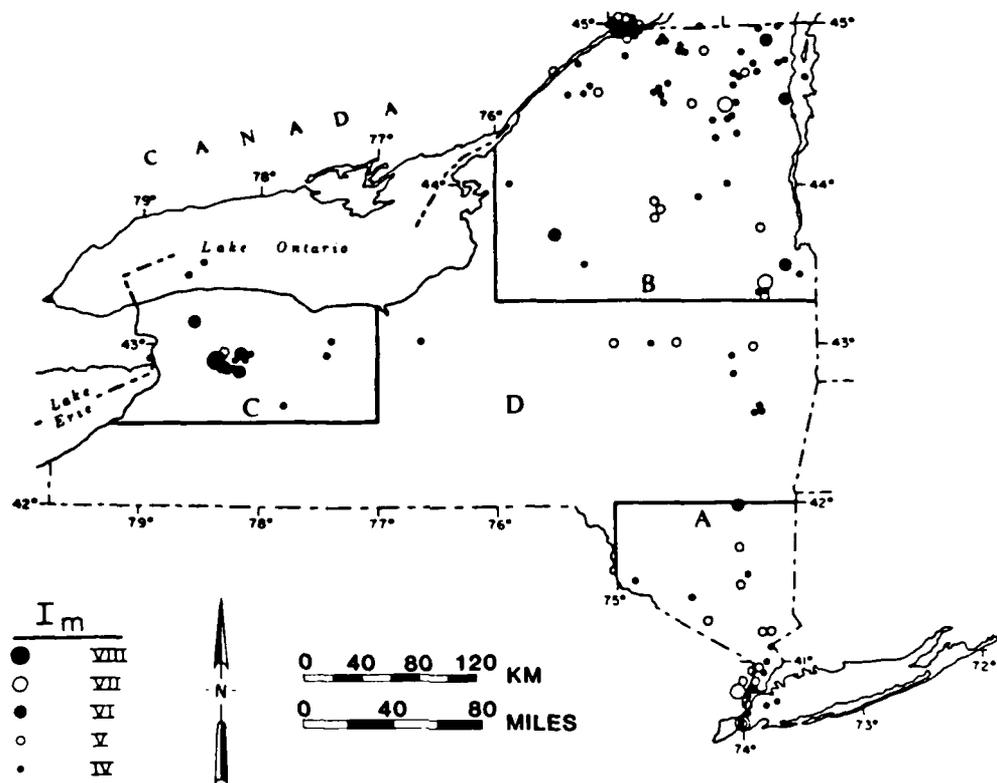
Figure 98. Epicentral map of New York showing areas of relatively greater seismicity: A. Hudson Highlands-Raritan Bay; B. Adirondack Mountains; and C. Attica
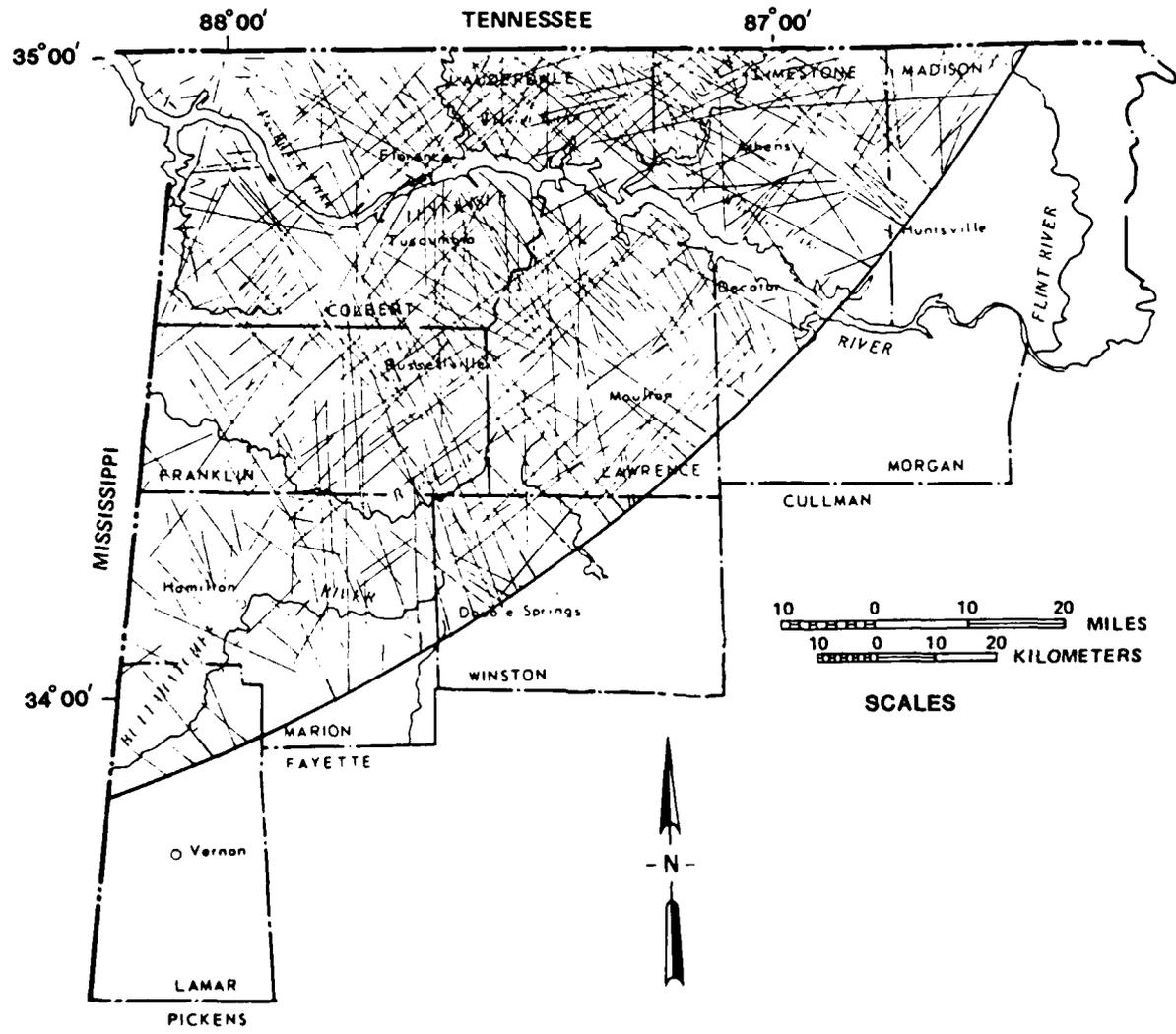
Figure 99. LANDSAT lineament map of northwest Alabama (compiled by Kidd, 1980, Fig. 12)
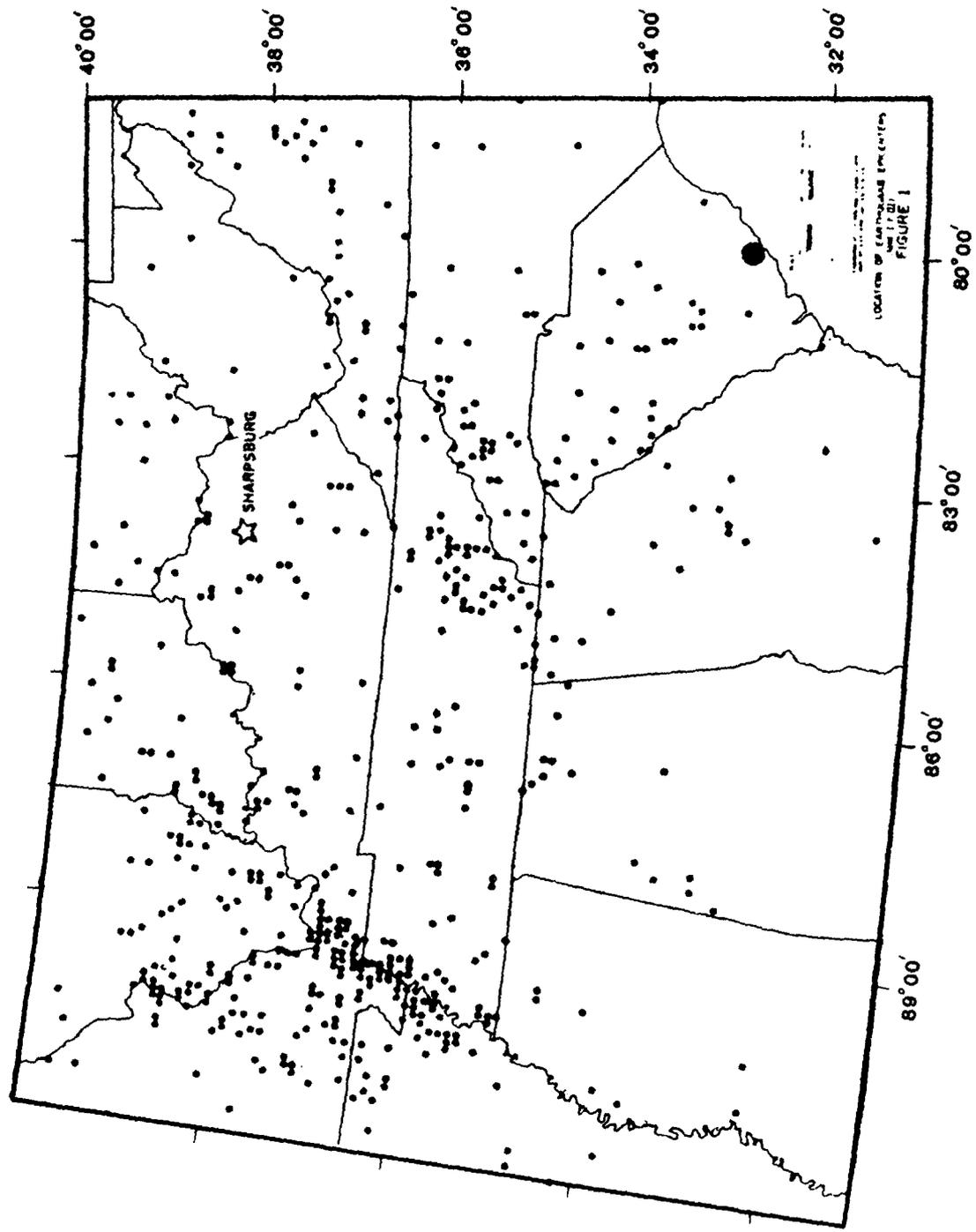
Figure 100. Epicentral map of the east-central United States (Reinbold and others, 1981, Fig. 1)

LEGEND

〜 APPROXIMATE OUTCROP AREA OF ORDOVICIAN ROCKS ON JESSAMINE DOME

— FAULT

⊗ CRYPTOEXPLOSIVE STRUCTURE

NE LETTERS DESIGNATE MAJOR BLOCKS
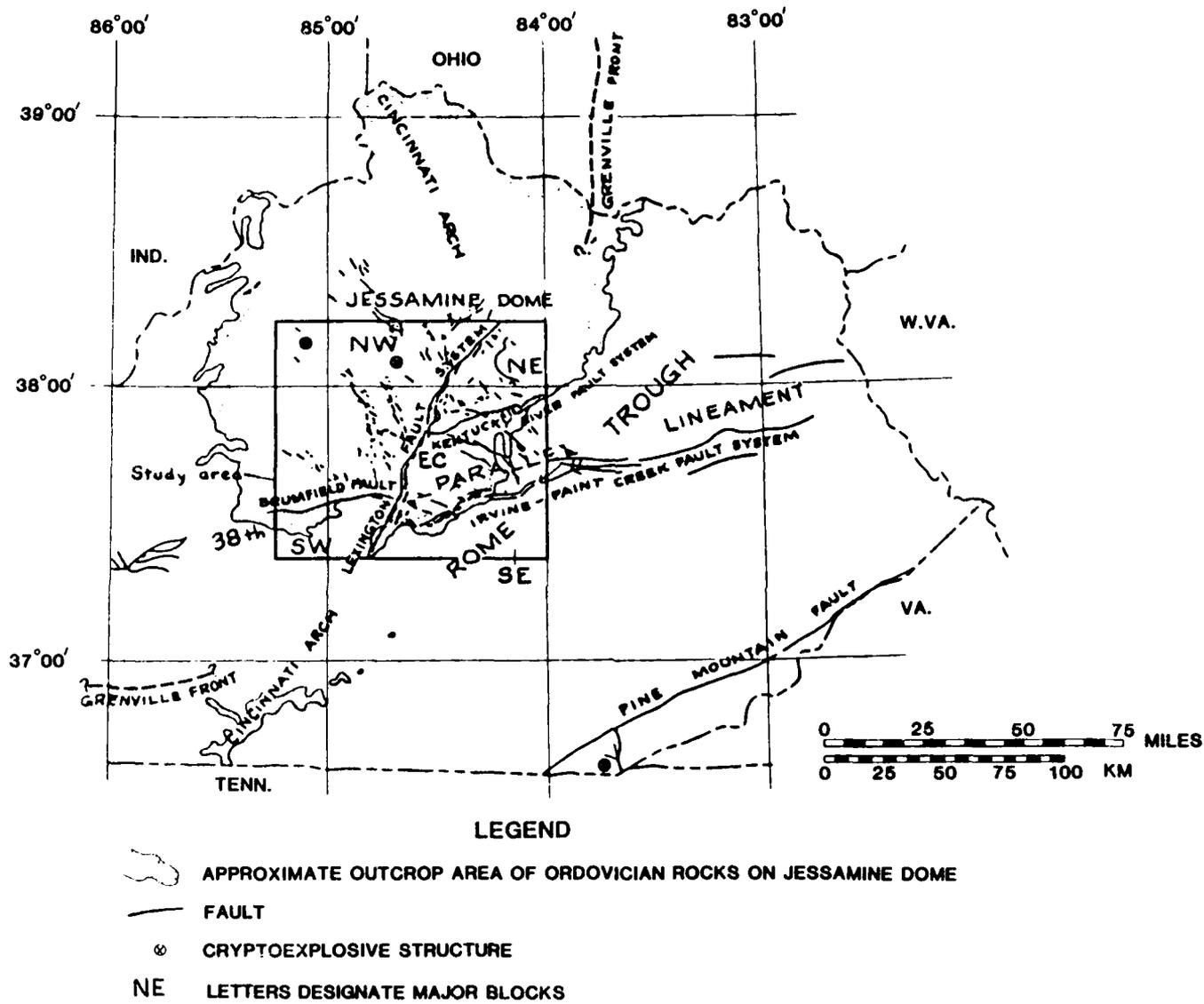
Figure 101. Map showing major structural features in central and eastern
Kentucky (Black and others, 1981, Fig. 1)
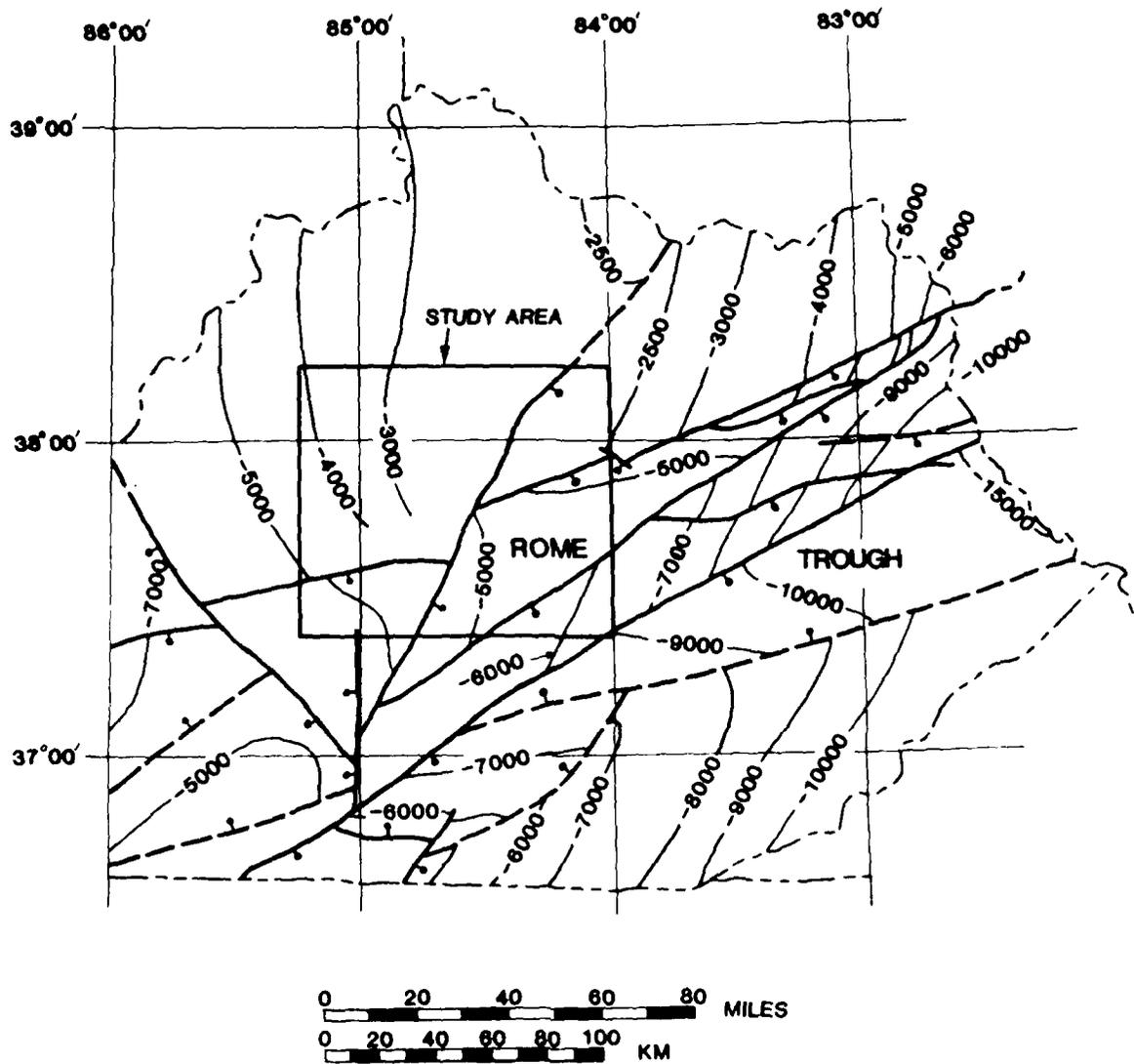
Figure 102. Structural map of the basement of central and eastern Kentucky according to Harris (1975). Bar and ball on downthrown side of fault. Contour interval, 1,000 feet. Intermediate contours shown locally where there was sufficient data. Not contoured locally. Sea level datum (Black and others, 1981, Fig. 4)
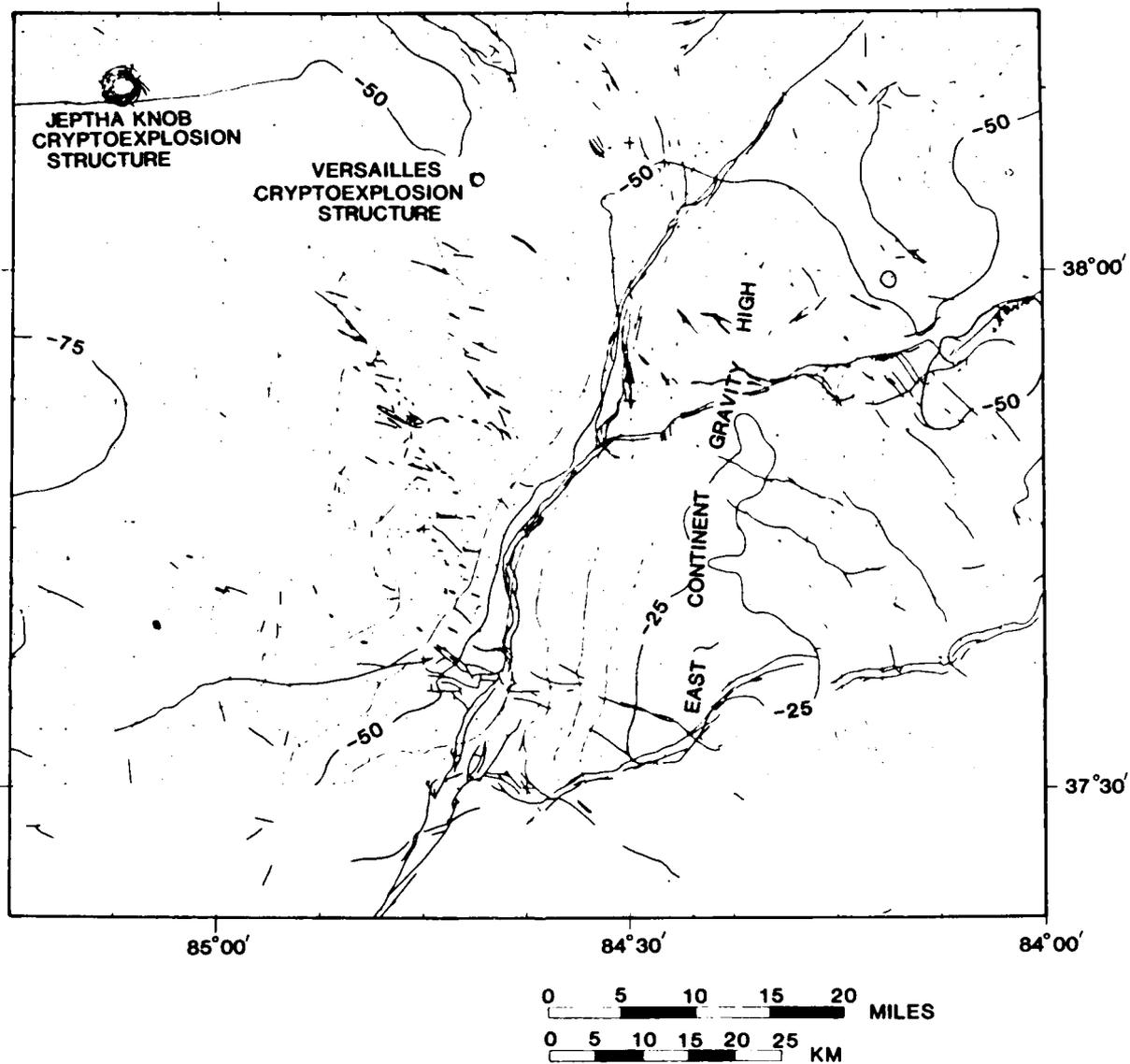
Figure 103. Bouguer gravity data shown in relation to surface faults in central Kentucky. Contour interval, 5 milligals. Dots show field station locations. Crosses indicate corners of 7½-minute quadrangles. Dark areas adjacent to faults are local occurrences of fault-associated dolomite (Black and others, 1981, Fig. 6)

DEVONIAN

Symbol	Formation
	"CANADAWAY GROUP" Shale, siltstone
	JAVA FORMATION Shale
	WEST FALLS FORMATION Shale, siltstone
	SONYEA FORMATION Shale
	GENESEE FORMATION Shale, limestone
	HAMILTON GROUP Shale, limestone
	ONONDAGA LIMESTONE AND BOIS BLANC FORMATION Dolostone, limestone, sandstone

PRECAMBRIAN

PRECAMBRIAN ROCKS
undifferentiated

SILURIAN

AKRON DOLOMITE AND
   SALINA GROUP
   Shale, dolostone

AKRON DOLOMITE AND
   SALINA GROUP
   Dolostone, shale, salt, gypsum

LOCKPORT FORMATION
   Dolostone, limestone

CLINTON GROUP
   Dolostone, shale, limestone

ORDOVICIAN

"MEDINA GROUP" AND
   QUEENSTON SHALE
   Sandstone, shale

MEAFORD-DUNDAS FORMATION
   Siltstone, limestone

BLUE MOUNTAIN AND
   COLLINGWOOD FORMATIONS
   Limestone

TRENTON AND BLACK RIVER
   GROUPS
   Limestone, dolostone

Figure 104. Geologic map of northwestern New York and adjacent Ontario showing location of Clarendon-Linden Fault (Hutchinson and others, 1979, Fig. 1)
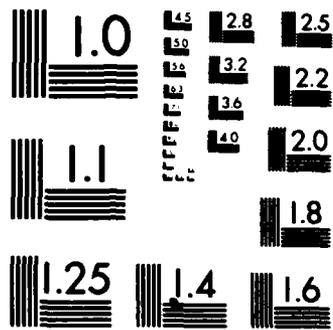
MICROCOPY RESOLUTION TEST CHART

Figure 105. Magnetic map of Attica, New York and vicinity showing interpreted lineaments (modified from Thompson and Kujawski, 1975)
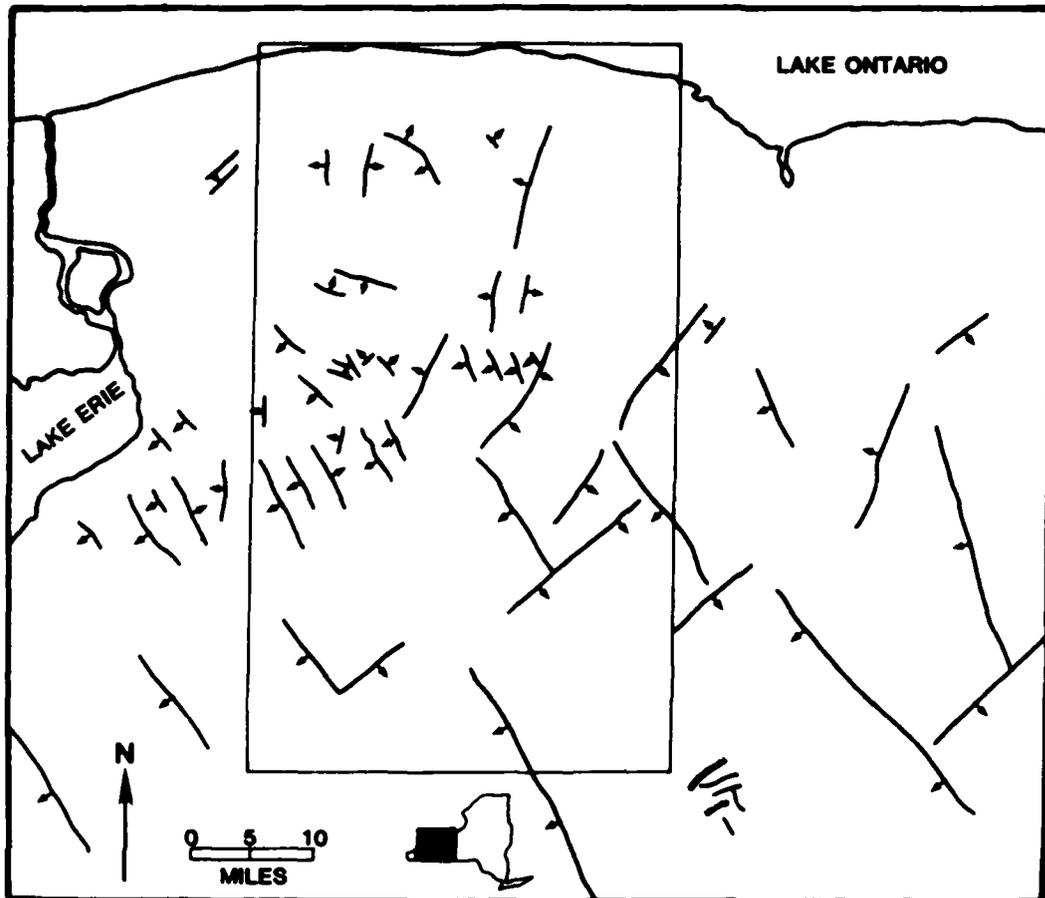
Figure 106. Map of western New York showing structural interpretation of outcrop patterns and linear topographic features based upon geologic data and aerial photographs (Fakundiny and others, 1978, Fig. 6). Each line represents the trend of a suspected monocline. Arrows show down-folded limb. These folds may represent surface expression of basement faults. Fault symbols with barbs on down thrown block are from Fisher et al., 1970. Rectangle defines study area of Fig. 1.
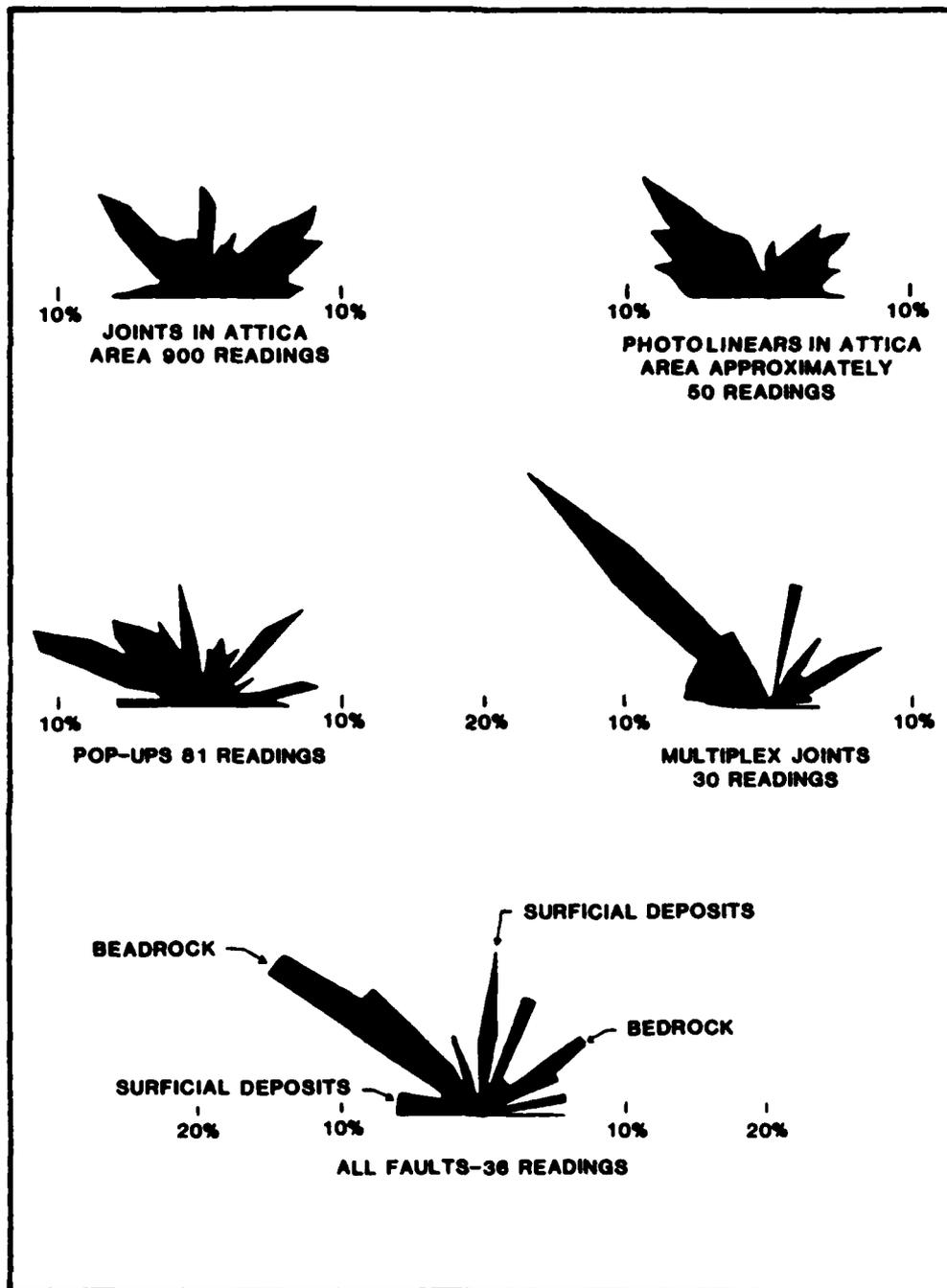
Figure 107. Rose diagrams of small-scale structural features of western New York (Fakundiny and others, 1978, Fig. 7). Diagrams show only the northern half of rose. All data normalized to 100 percent fro purposes of comparison. Note the predominant NW trend and subordinant NE trend. Pop-ups, multiplex joints, and faults are measured.
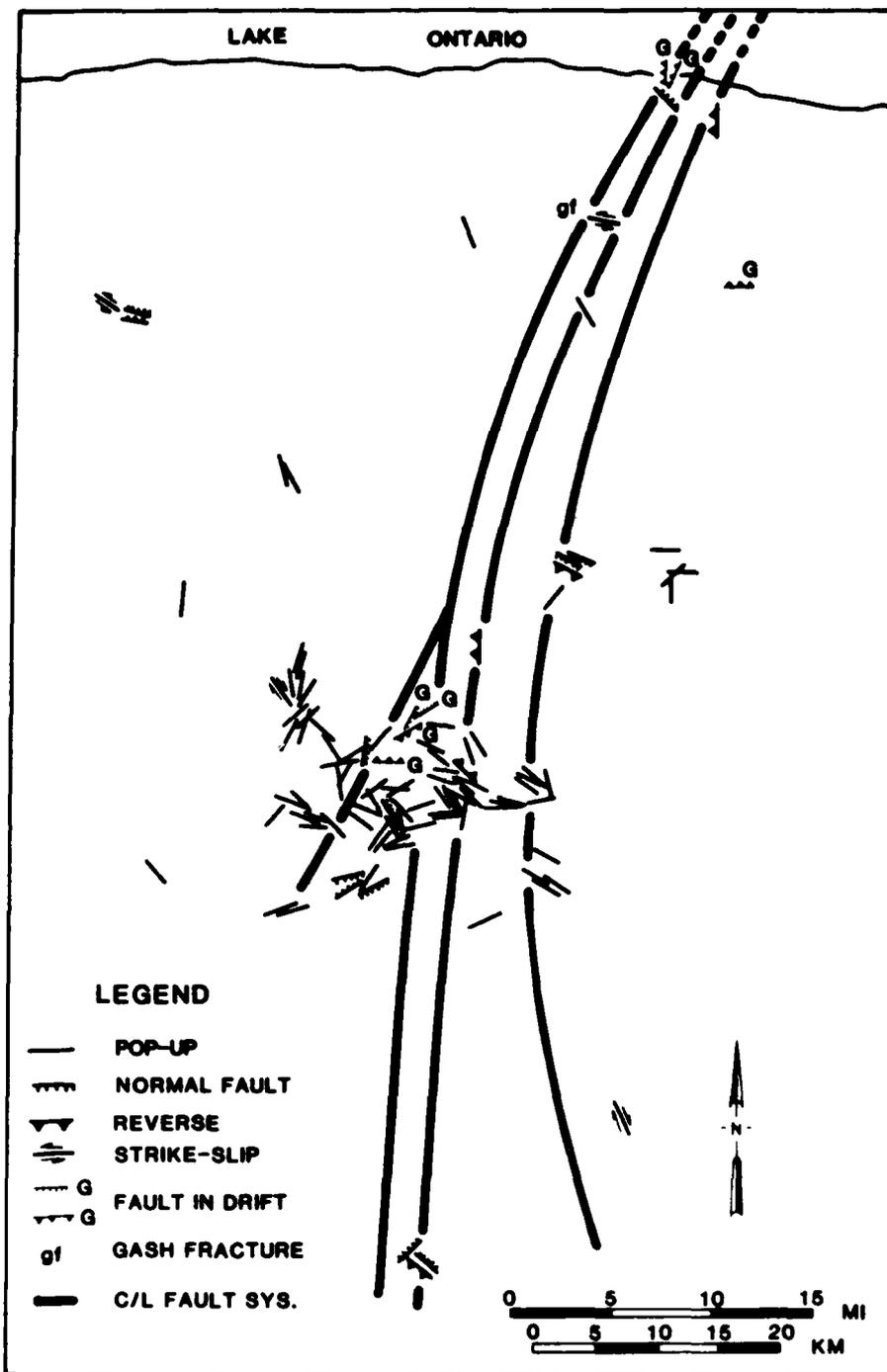
Figure 108. Map of the Attica region, western New York, showing instability structures and traces of the Clarendon-Lindeon Fault system (Fakunainy and others, 1978, Fig. 21)
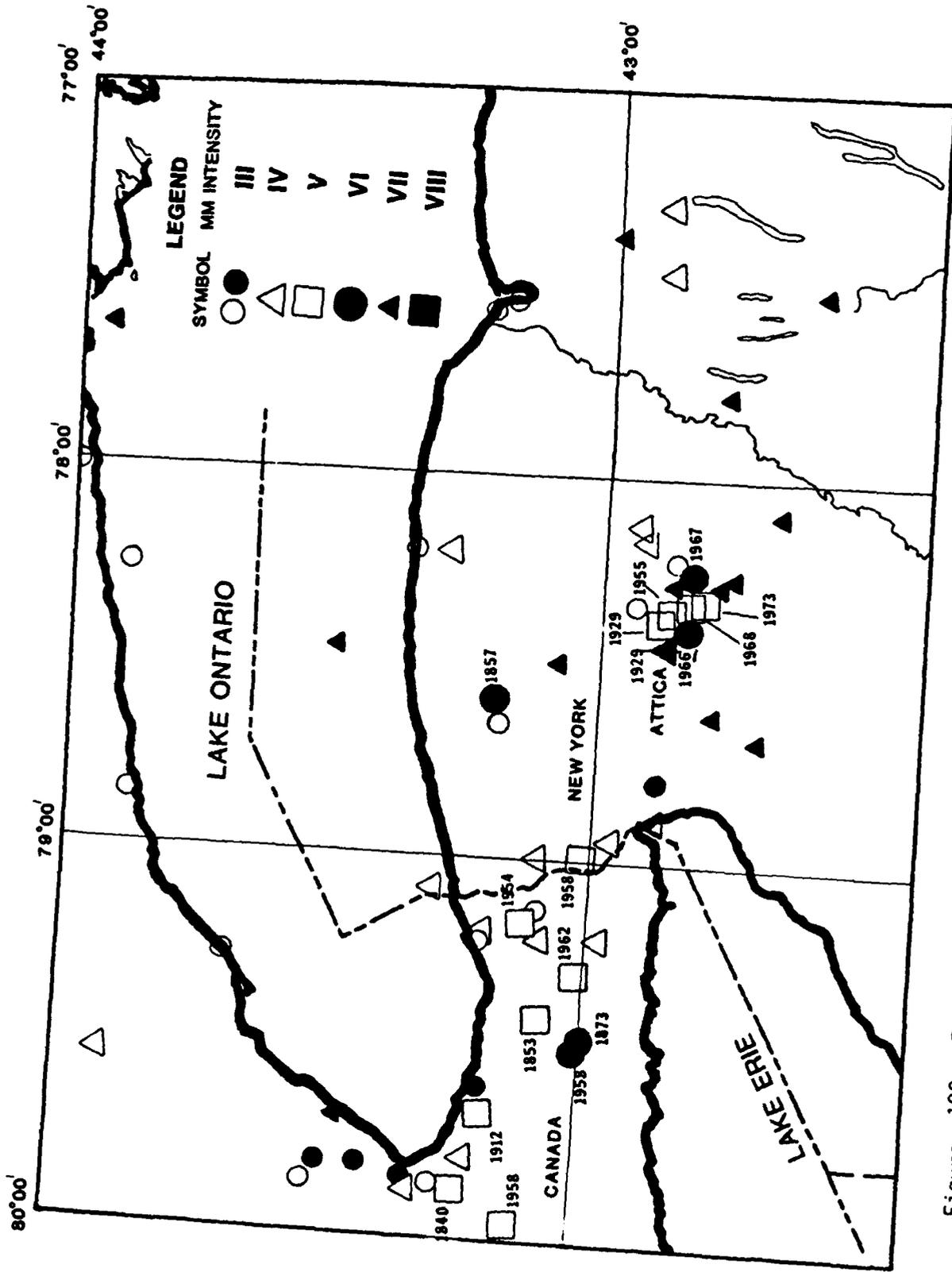
Figure 109. Epicentral map of northwestern New York and adjacent Ontario showing earthquakes through 1980 (Nottis and Mitronouas, 1983) Attica located in cluster of events.
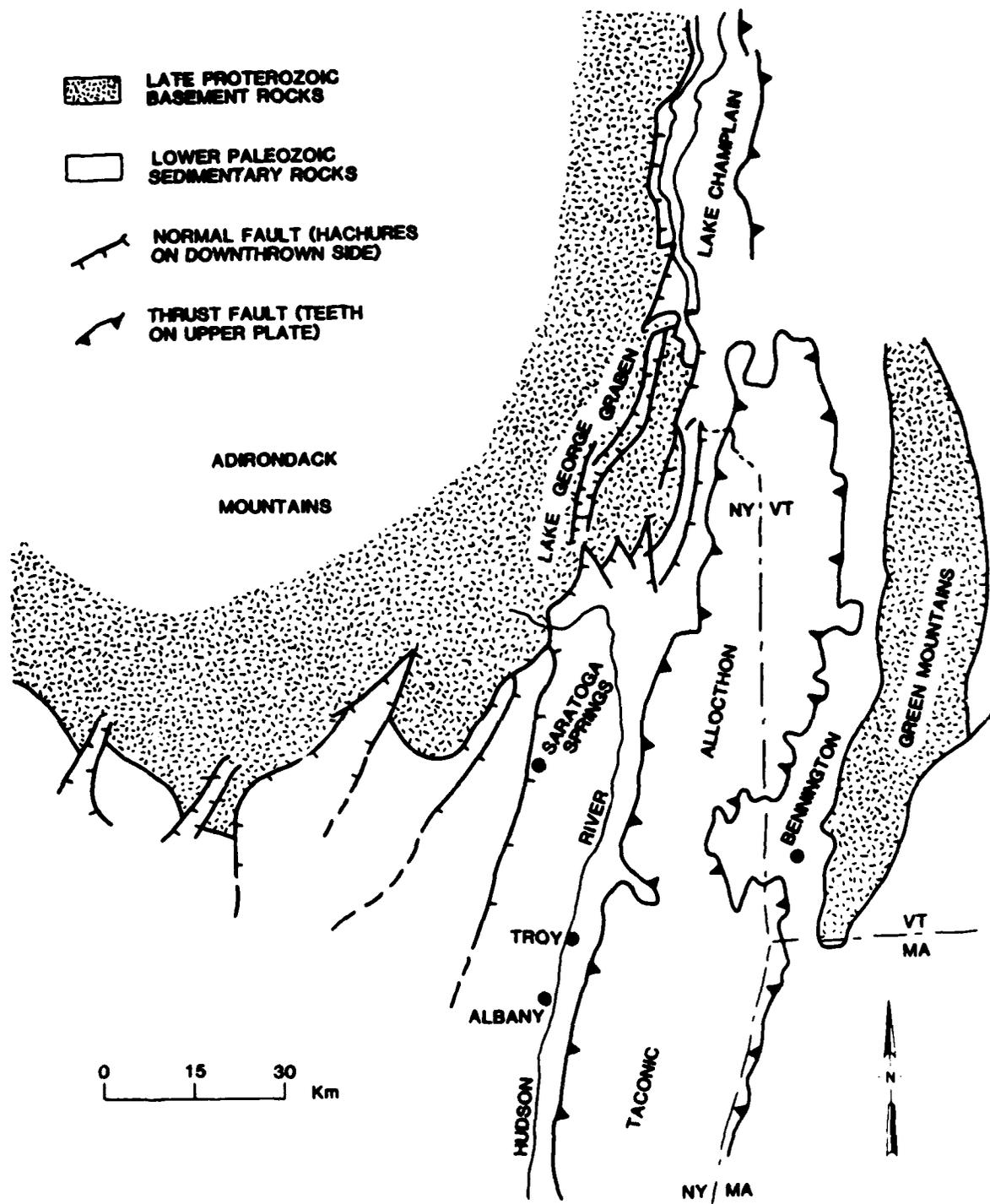
Figure 110. Generalized geologic map of the Lake Champlain-Hudson River region (Putnam and others, 1983, Fig. 17)
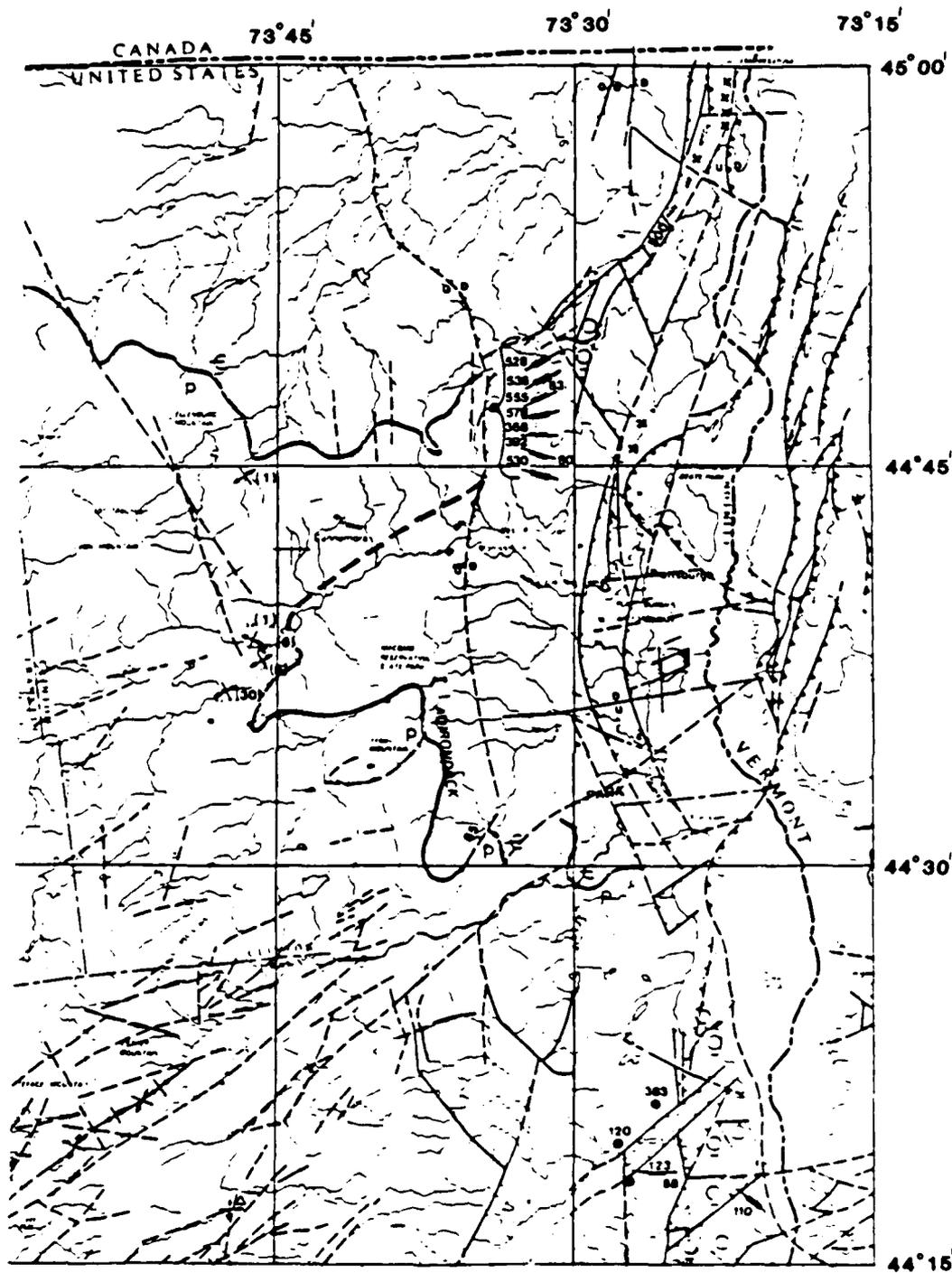
Figure 111. Northeastern corner of the Brittle Structures Map of New York, showing the addition of attitudes and K/Ar ages of dikes (modified from Isachsen and McKendree, 1977) (Isachsen, 1983, Fig. 35)
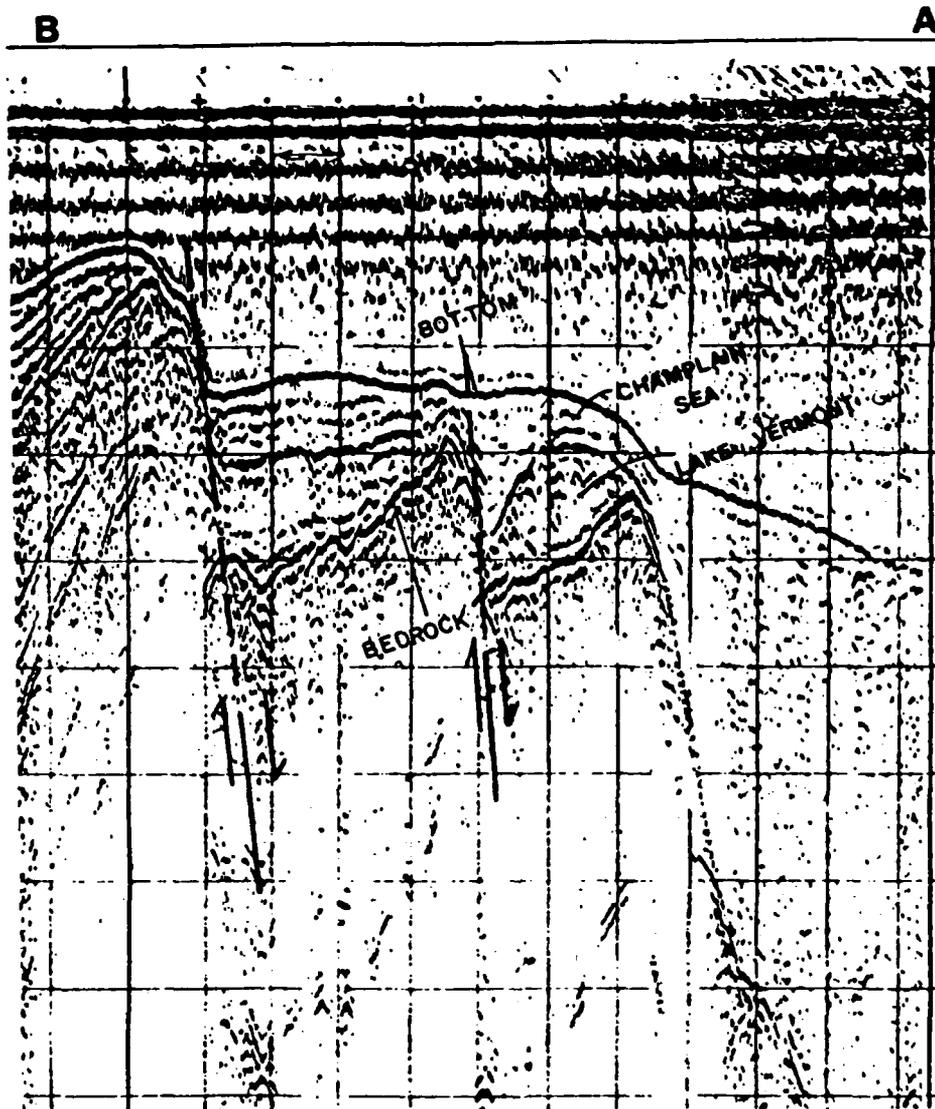
Figure 112. Seismic profile on Lake Champlain showing features interpreted as bedrock faults. Suspected faults depicted by dashed lines. Horizontal lines spaced 60' apart, vertical lines 475'. Vertical exaggeration about 13.1. Data from air-gun survey. (data from A.S. Hunt and J.J. Dowling: in Barosh, 1981, Fig. 50)
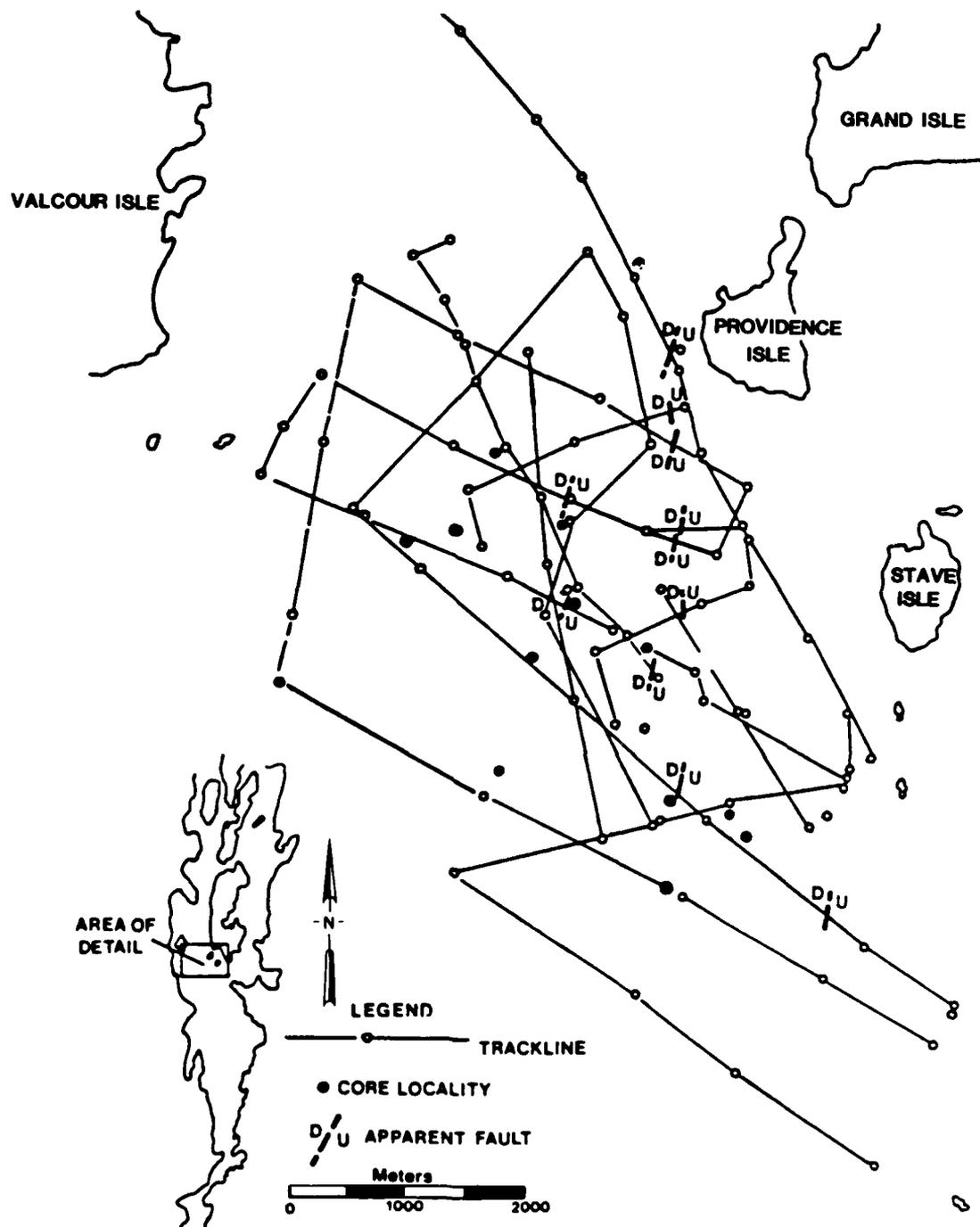
Figure 113. Map of central Lake Champlain showing ship track, core locations and apparent Holocene faults (Hunt, 1982)
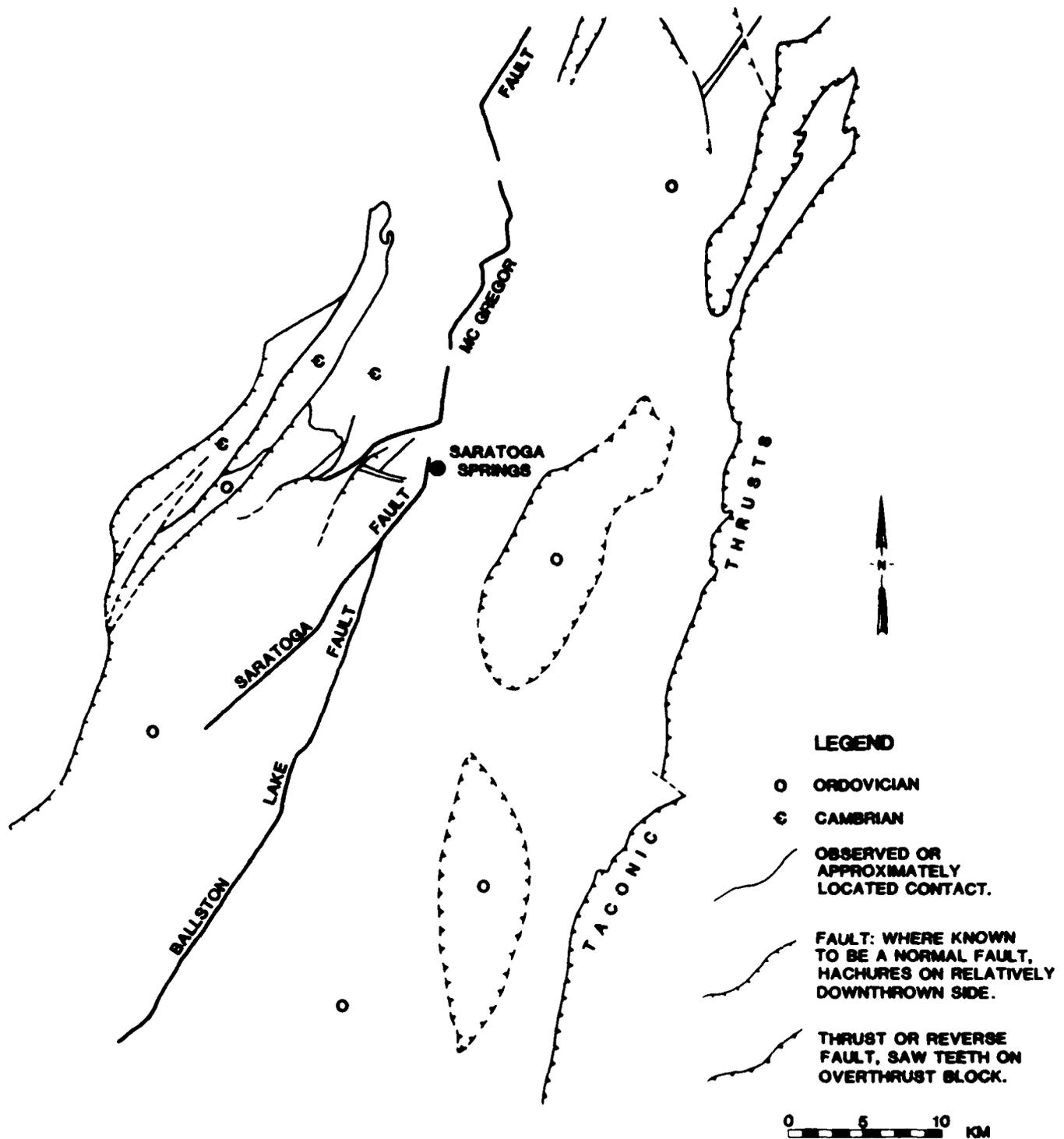
Figure 114. Geologic map of Saratoga Springs and vicinity, New York, showing the McGregor-Ballston Lake fault system that is expressed as scarps (Isachsen, 1981a, Fig. 24)
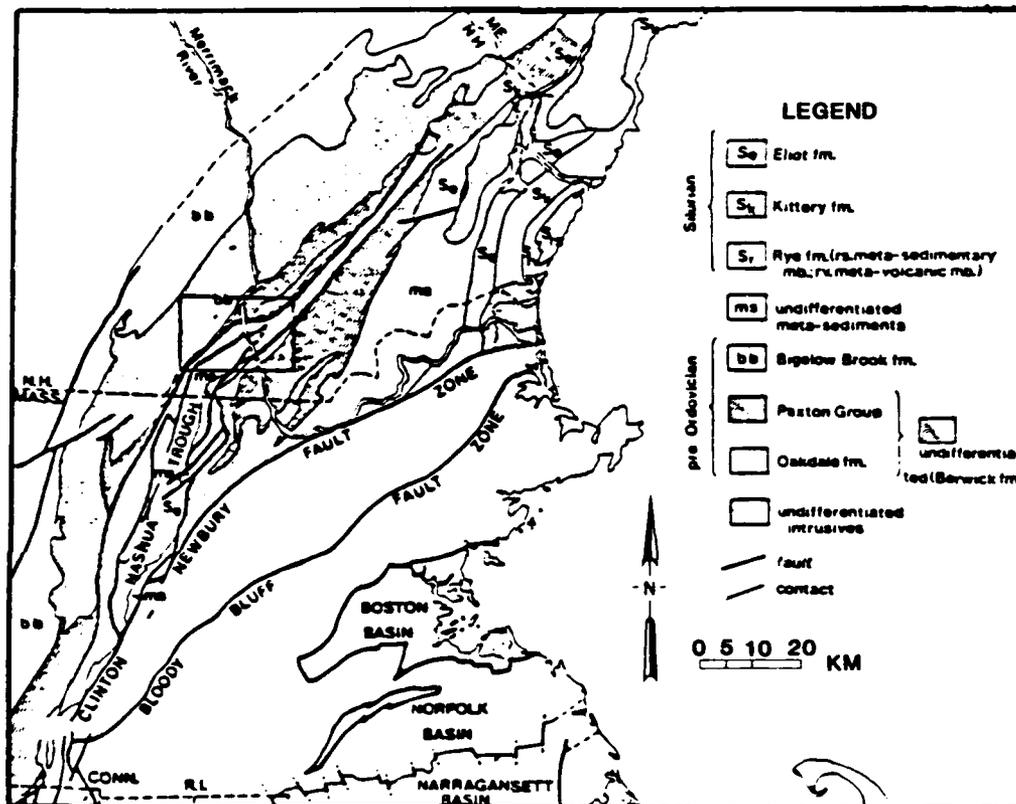
Figure 115. Sketch map showing generalized geologic structure of southeastern New Hampshire and adjacent Maine and Massachusetts, and location of Fig. 116 (Barosh and Moore, in prep.) (Smith and Barosh, 1983, Fig. 51)
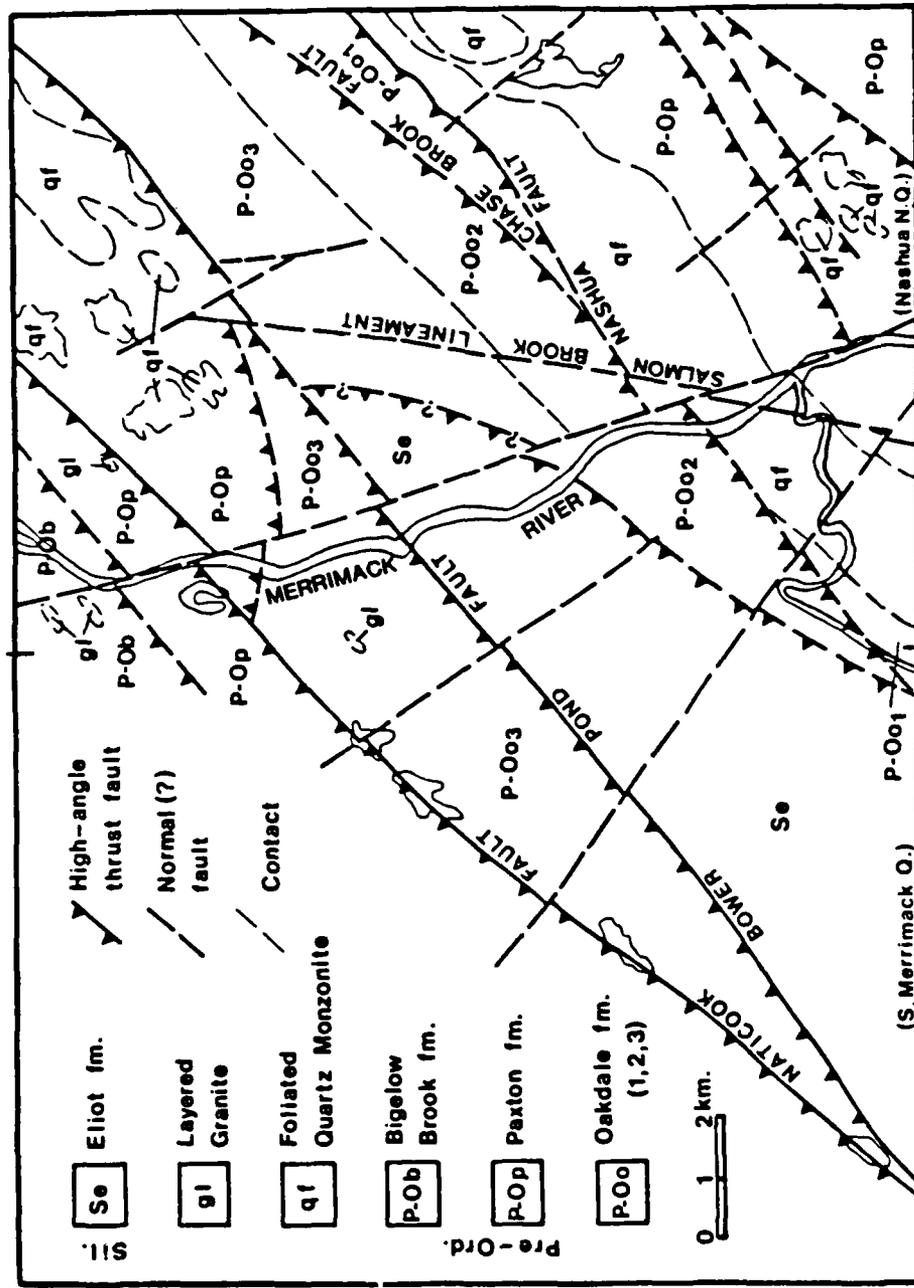
Figure 116. Geologic sketch map of the Nashua North and South Merrimack 7½-minute quadrangles, New Hampshire (Smith and Barosh, 1983, Fig. 52)
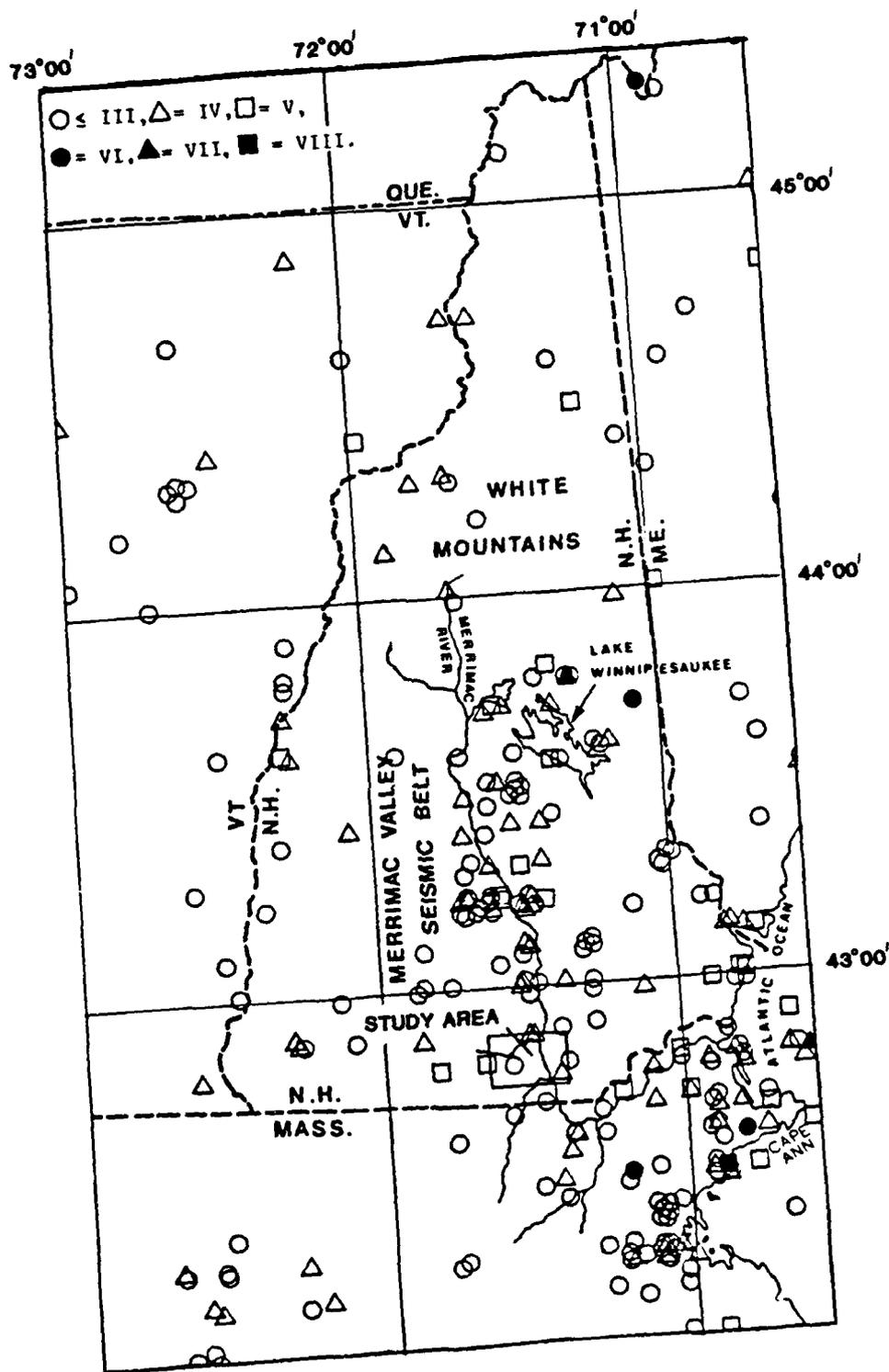
Figure 117. Epicentral map of New Hampshire and surrounding region showing earthquakes through 1980 (data from Chiburis and others, 1980) (Smith and Barosh, 1983, Fig. 50)
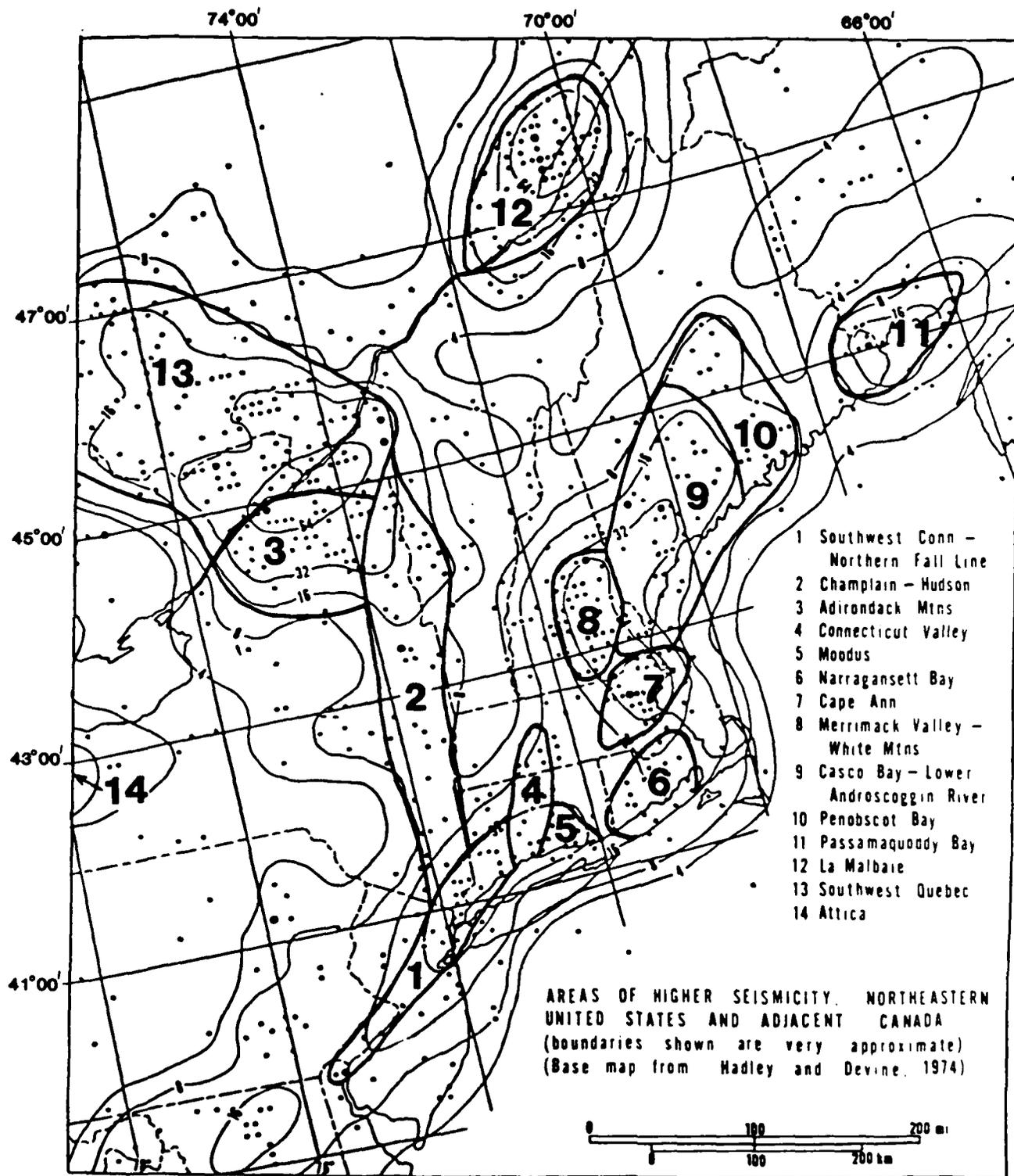
Figure 118. Map of the northeastern United States and adjacent Canada showing relatively higher areas of seismicity (Barosh, 1981, Fig. 2)
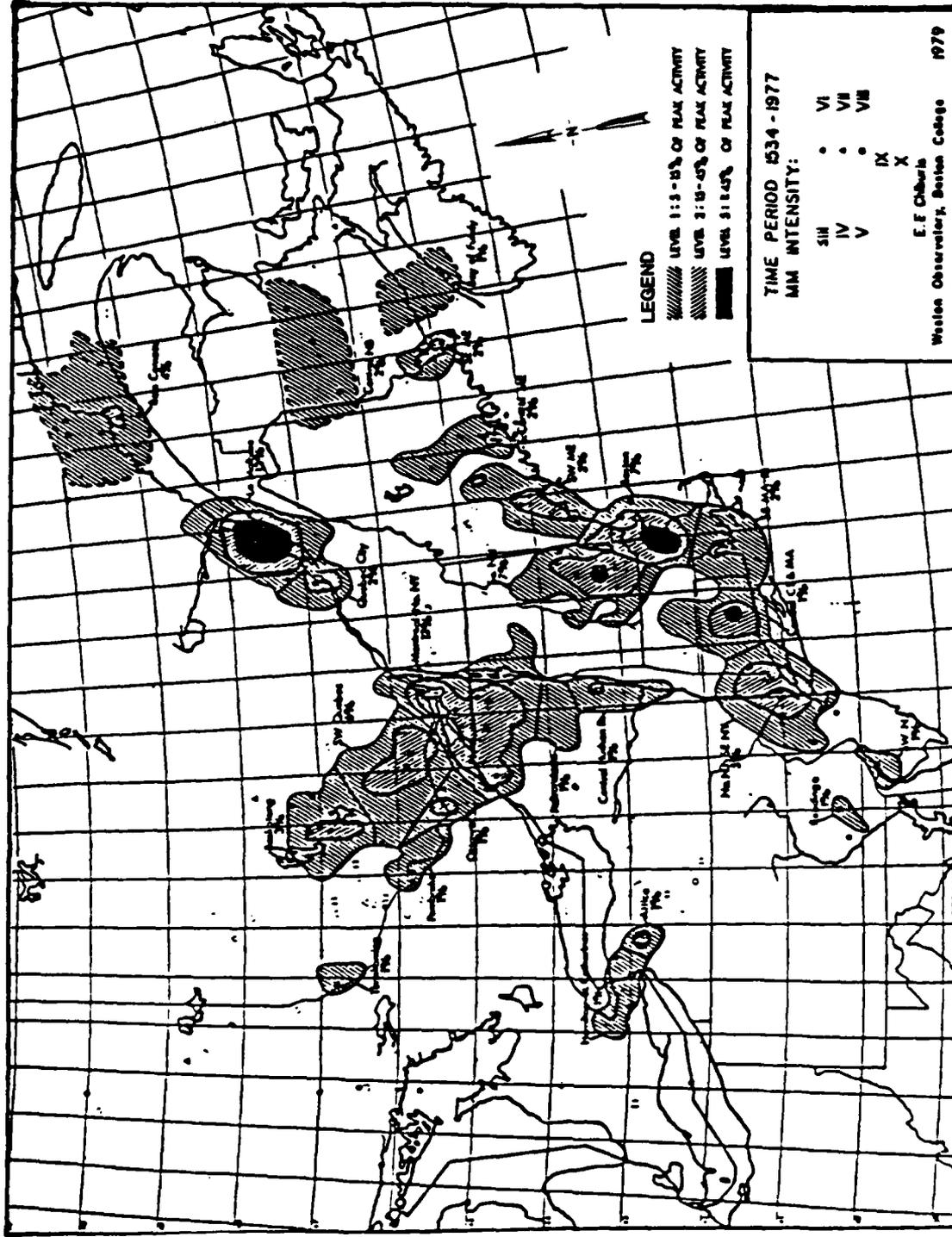
Figure 119. Map of the northeastern United States and adjacent southeastern
Canada showing seismic subregions contoured at three levels of
relative activity for the time period 1534-1977 and their
percentage of the overall activity (Chiburis. 1981)
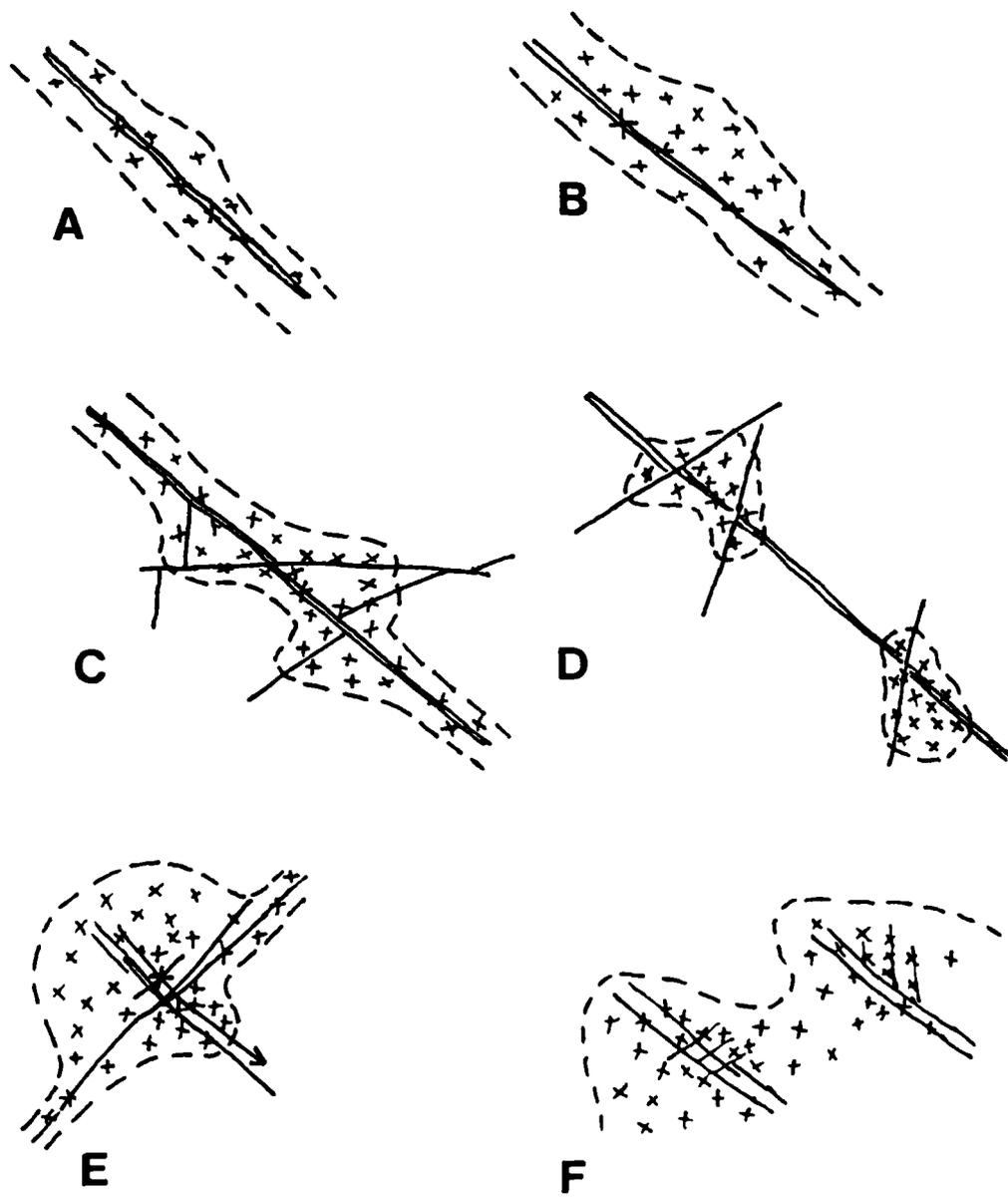
Figure 120. Sketch maps showing possible relation of seismic zones with caustive structures
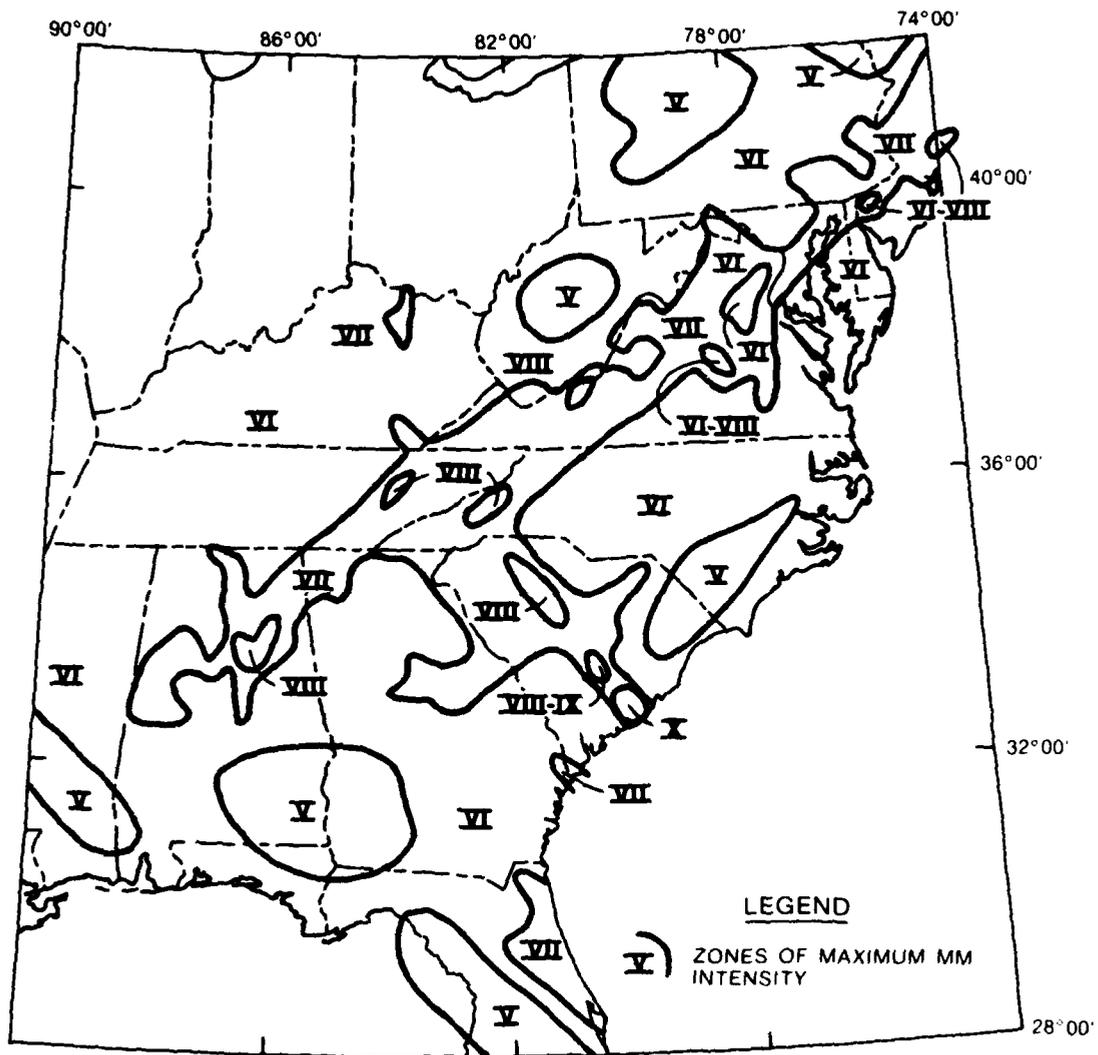
Figure 121. Seismic zonation map of southeastern United States
(see text for explanation)
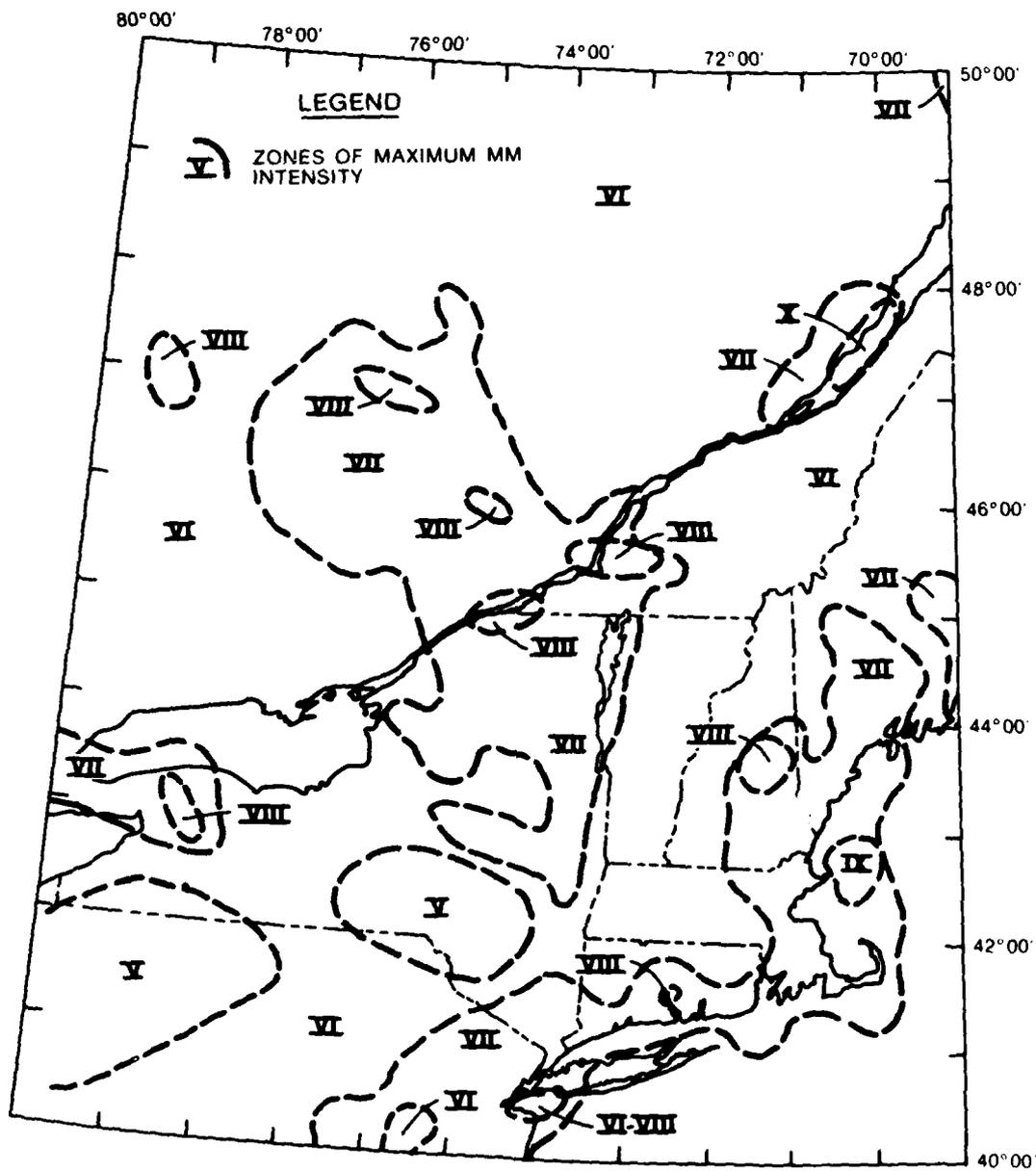
Figure 122.   Seismic zonation map of northeastern United States
(see text for explanation)

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