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SEISMOTECTONIC MAP OF THE EASTERN UNITED STATES*

By Jarvis B. Hadley and James F. Devine

INTRODUCTION

The purpose of a seismotectonic map is to describe the distribution of historic seismic activity in relation to geologic structures and tectonic provinces and to identify structures or regions that are characterized by consistent relations between seismic activity and structural features. An experimental study to test the feasibility of preparing such a map and its usefulness in evaluating earthquake hazards was initiated jointly by the U.S. Geological Survey and the former Seismology Division of the National Oceanic and Atmospheric Administration; financial support was provided by the U.S. Atomic Energy Commission. It was hoped that the study would develop evidence for the existence of tectonic regions or structural provinces that are useful in defining parameters of seismic hazard in relation to engineered structures. As with many experiments, this one did not produce the results envisioned at the outset, but it did result, we think, in assembling seismic and structural data useful in evaluating the distribution and origin of earthquakes in the eastern part of the United States and in delimiting areas of relative seismic activity.

The tectonic provinces of the eastern United States, as outlined and described by King (1951), are as follows:

- I. Central stable region
  - a) Laurentian shield (part of Precambrian Canadian shield)
  - b) Interior lowlands (bordering platforms covered by younger rocks)
- II. Appalachian (orogenic) system or tectonic province
  - a) Fold belt (roughly the Valley and Ridge physiographic province)
  - b) Blue Ridge belt (Blue Ridge physiographic province)
  - c) Piedmont belt (Piedmont physiographic province and part of Blue Ridge physiographic province)
  - d) New England-Maritime belt (extension of belts II a, b, and c north of New Jersey and Pennsylvania)
- III. Coastal Plains (postorogenic deposits overlapping the Appalachian system; of subordinate tectonic significance except in Mississippi embayment).

It is clear from existing structural data, as well as from the distribution of earthquakes, that historic earthquake activity bears no consistent relation to the provinces and subprovinces defined above. Parts of the provinces differ from one another in structure and tectonic history, especially of Mesozoic and Cenozoic age, and in seismic history during the past 200 years. Seismically active zones locally cut across tectonic provinces as, for example, the New Madrid seismic zone and the transverse seismic area in central Virginia. Only the southwestern faulted part of the Appalachian fold belt is identified as a tectonic province with fairly uniform structural characteristics and seismicity.

The basic method of this study has been to compare seismicity, as shown by the location of epicenters of recorded earthquakes (map B), with lithologic, structural, and geophysical features (map A) that may have served to generate or control the release of seismic energy. Results of the study, depicted as areas of varying amounts of seismic activity and varying degrees of inferred structural control, are shown on the seismotectonic map (map C).

The epicenter plot of map B shows computer plotted locations of epicenters of earthquakes of Modified Mercalli (MM) III or greater intensity which occurred during the period 1800-1972. Reasonably complete and accurate data on earthquakes in the eastern United States have been kept since about 1900, and about 75 percent of the earthquakes of MM IV-V and greater used in present study (Coffman and von Hake, 1972, p. 5-57) occurred during this period. For New England and New York this figure is about 66 percent; for the southern Appalachian and coastal areas it is 70 percent; and for the region extending from the Appalachian Mountains to and including the Mississippi River valley it is about 80 percent. Because of the relatively low frequency of occurrence of eastern earthquakes (about 6.2 earthquakes of MM IV-V or greater per year during 1900-1970), it was considered desirable to include data from the earlier period of less complete records in order to obtain as long a record of seismic activity as possible. This means that probably 85 percent of the epicenter locations used are based on macroseismic rather than instrumental observations and are limited correspondingly in accuracy.

The epicenter data used for this study were extracted from the current working file of the National Earthquake Information Service and do not represent a final edited compilation of eastern United States seismicity. However, subsequent refinement and improvement of locations or maximum intensities are expected to result in only minor changes in the seismic-frequency contours developed for this study. The locations and maximum epicentral intensities of earthquakes were recorded on magnetic tape and plotted from the tape. Locations were fixed to the nearest 0.1 degree of latitude and longitude; the symbol used represents the highest epicentral intensity located at the plotted point.

The contours shown on the epicenter plot were developed by counting the number of epicenters, including repeated events at the same plotted location, within a circle representing 10^4km^2 at the original map compilation scale of 1:2,500,000. They were generated by tracing in a somewhat generalized fashion the position of the center of the counting circle while moving it so as to include up to but not more than the 4, 8, 16, 32, and 64 epicenters selected as contour intervals. Because we wished to emphasize the areal distribution of seismic activity which might be related to geologic structure, rather than the total seismic energy released, the intensities of individual earthquakes were ignored in contouring seismic frequency.

We wish to acknowledge here the indispensible contribution of D. W. Gordon and Carl Stover, then of the Seismology Division of the National Oceanic and Atmospheric Administration and now of the Geological Survey, in extracting the earthquake data and preparing the computer plot. We also acknowledge the work of D. W. Gordon in preparing an earlier version of the seismic-frequency contours.

The accompanying tectonic map shows the major structural features and tectonic provinces of the eastern United States as currently understood, with emphasis on features that seem to be involved in historic seismic activity (map A). The features shown are abstracted primarily from the Tectonic map of the United States (Cohee, 1962) and the Basement rock map of the United States (Bayley and Muehlberger, 1968), with minor revisions and additions based on information published or obtained since those maps were compiled. For New England and adjacent parts of Canada, additional geologic information was obtained from a recently published tectonic map of the Appalachian region (Rodgers, 1970, pl. 1), from geologic maps of New Hampshire and Vermont (Billings, 1956; Doll, 1961), and from unpublished data for southern New England (Barosh, P. J., written commun., 1973). Aeromagnetic maps of many areas and from various sources were used, especially for information on the structure of the Precambrian basement west of the Appalachian tectonic system, on the pre-Cretaceous basement beneath the Atlantic Coastal Plain, and on the rocks beneath the Gulf of Maine. For similar purposes, as well as to obtain information about the deeper structure in the Appalachian tectonic belts, the Bouger gravity anomaly map of the United States (Woollard and Joesting, 1964)

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was used.

In addition to assembling and analyzing the seismic and structural data, descriptions of about 60 earthquakes that had been investigated in the field or otherwise described in more than cursory fashion were examined for evidence indicating structural control. With the exception of the New Madrid earthquakes of 1811-12, none of these reports contained evidence of surface-breaking faults. For only a few earthquakes which have occurred since 1960 did such descriptions indicate evidence of structural control.

To review the tectonic and structural features of the eastern United States and to provide background for their relation to historic seismic activity, as presented in the accompanying seismotectonic map, a brief structural summary follows.

SUMMARY OF STRUCTURAL FEATURES

Precambrian structures

Precambrian rocks underlying a relatively thin cover of Paleozoic rocks are the dominant supracrustal rocks throughout the interior of the United States and Canada. Although largely concealed within the United States, the lithology, age, and distribution of the Precambrian rocks are at least partly known from exposures, drill records, and aeromagnetic and gravity surveys. Structures thus recognized include both local trends and regional features, such as the Lake Superior syncline in Wisconsin, Minnesota, and Iowa and the so-called Grenville metamorphic or orogenic front that has been traced from eastern Canada southward into Michigan and Ohio.

For the most part, major Precambrian structures of the continental interior are not specifically associated with current or historic seismicity. It is not known whether earthquakes have resulted from movements on Precambrian faults or other structural discontinuities, especially in the central United States where scattered smaller earthquakes commonly cannot be related to known features in the Paleozoic or younger rocks. It is notable, however, that such structures seem to have played only a minor or local role in determining the release of seismic energy (Kisslinger and Nuttli, 1965).

Rocks similar in age and Precambrian structural history to those in the eastern part of the continental interior occur also within the Appalachian system from New England to Georgia. Their structural trends have, however, been so strongly overprinted in most places by Paleozoic deformation that no distinctively Precambrian structures can be related to regional seismic activity.

Paleozoic structures

Appalachian system

Paleozoic structures include structures that have affected the Paleozoic rocks and can reasonably be regarded as having formed before Mesozoic time. They are concentrated in the Appalachian and Ouachita tectonic systems, where their characteristic features are those of supracrustal compression and shortening across the belts, resulting in folds and faults, regional dynamothermal metamorphism, and emplacement of various intrusive rocks. Since most of the Ouachita system lies outside the area of the present study, the subsequent discussion deals mainly with the Appalachian system.

The Appalachian system dominates the eastern part of the continent, especially within the United States. It consists of several structural belts which can be traced throughout the length of the system. The northwestern or Appalachian fold belt consists of strongly folded Paleozoic sedimentary rocks 3 to 10 km thick, broken by low-angle thrust faults that are largely confined to the sedimentary rocks above a deeply buried but little deformed basement surface. The fold belt is 32 to 96 km wide in most places but increases to nearly 200 km in Pennsylvania and West Virginia. Repeated low-angle faults duplicating parts of the sedimentary sequence are prominent throughout the southwestern part of this belt, but they are much less abundant in the central part. The equivalent structural belt north of New York City is much narrower and is characterized by both folds and low-angle faults. The western boundary of the fold belt southwest of New York City is a transitional zone in which the characteristic Appalachian folds die out in the relatively undeformed rocks of the Appalachian Plateaus of the continental interior. North of New York the structural province adjoining the fold belt on the west is the Canadian shield and its outlier in the Adirondack Mountains of New York state. This part of the boundary is not only sharper but presents a much stronger tectonic contrast than that farther southwest.

The Blue Ridge belt, bordering the fold belt on the southeast, is characterized by many folds and faults and, more importantly, by major deformation involving not only the Precambrian basement rocks but also several kilometers of upper Precambrian sedimentary and volcanic rocks, which are mostly absent from the fold belt. The Blue Ridge belt is also characterized by higher grades of regional metamorphism; hence its rocks are relatively strong crystalline rocks, as distinguished from the less crystalline and generally weaker sedimentary rocks of the fold belt. In many places the northwestern boundary of the Blue Ridge belt and its northern extension, the Green Mountains of western New England, is marked by low-angle faults on which the rocks of the Blue Ridge and Green Mountain belts have been carried westward over the rocks of the fold belt. This is especially true in the southwestern part of the system, where the Blue Ridge rocks have been thrust many kilometers over the rocks of the fold belt.

The Appalachian Piedmont belt adjoining the Blue Ridge belt on the southeast is as wide as or wider than the Blue Ridge and is at least as complex structurally. It is less well known geologically and consists largely of unfossiliferous sedimentary and volcanic rocks moderately to strongly recrystallized and invaded by many different kinds of felsic and mafic intrusive rocks. The metamorphosed volcanic and sedimentary rocks are probably mostly of early Paleozoic age but may include rocks of late Precambrian age, as well as isolated bodies of older Precambrian rocks. Deformation, metamorphism, and intrusion are believed to have taken place at several times during the Paleozoic. Rocks in the northeastward continuation of the Piedmont belt in New England range in age from early through middle and late Paleozoic, and their structural history has been at least as complex as that of the Piedmont rocks.

The geologic boundary between the Blue Ridge and Green Mountain belts and the Piedmont belt to the southeast is not well defined. In the southern Appalachians the geologic boundary has been interpreted as a zone of combined recumbent folding, overthrust faulting, and pre-Triassic high-angle or strike-slip shearing known as the Brevard fault zone (Reed and Bryant, 1964; Burchfiel and Livingston, 1967; Hatcher, 1971; Roper and Justus, 1973). Farther northeast in the central and northern Appalachians, the boundary is even less well defined. For the purposes of this report, the Blue Ridge and Piedmont belts and their equivalents in New England are considered as a single tectonic province.

Rock types and structures characteristic of the Appalachian Piedmont disappear eastward beneath the deposits of the Atlantic Coastal Plain so that no structurally significant eastern boundary is known in the eastern United States. Evidence from drill holes and aeromagnetic surveys, however, suggests that the rocks of the Florida platform buried beneath the Coastal Plain are significantly different from those of the Piedmont in central Georgia, and therefore a major tectonic province boundary may exist between them.

Central stable region

Distinctive Paleozoic structures of the continental interior are principally broad upwarps such as the Cincinnati, Findlay, and Kankakee arches, the Ozark uplift, and the Nashville dome, bordered by moderately deep depositional basins of Paleozoic rocks, including the Michigan and Illinois basins. Exceptionally deep depositional troughs border the Appalachian and Ouachita tectonic systems. These uplifts and the complementary basins developed gradually throughout most of Paleozoic time. In places the uplift or subsidence was accompanied by high-angle faulting and mild folding, the effects of which are included among the Paleozoic structures on the accompanying tectonic map.

Structures of late Paleozoic(?) and Mesozoic age

Appalachian system

The orogenic events that formed the Appalachian system in the eastern United States were succeeded by a new tectonic regime both in the Appalachian orogenic belt and in the central stable region to the west. This regime was at least partly extensional and, caused high-angle faulting of near-surface rocks, in contrast to the compressional and deep-seated effects of the earlier tectonism. These later features are divided into two broad groups, a series of fault-block basins of Triassic age entirely within the Appalachian orogenic system, and a group of riftlike fault zones within the central stable region.

Structures known to be Triassic are confined essentially to the Appalachian Piedmont and its New England counterpart. They consist of block-faulted basins in which continental sediments were deposited, in part contemporaneously with the faulting. The deformation was primarily extensional, resulting in tilted and down-dropped basins bordered by high-angle faults. Invasions by basaltic rocks are associated with these movements throughout the Piedmont belt. Neither the sedimentary nor the igneous rocks have been extensively deformed or metamorphosed, as they were formed after Paleozoic deformation and regional metamorphism had ceased completely.

In New England a chain of shallow intrusive and associated volcanic rocks known as the White Mountain plutonic-volcanic series extends northward from central to northern New Hampshire. They include stocks, arcuate dikes, and related intrusive rocks of Jurassic age which, like the Triassic rocks, are younger than the Appalachian tectonic belts. They are well represented in the seismically active part of central New Hampshire, and a recent earthquake has been ascribed to displacement along the border of one of the intrusives (Leet and Linehan, 1942).

Central stable region

The fault zones in the central stable region include a series of east-trending fault zones extending through Kentucky, across southern Illinois, and into southeastern Missouri. They are known in eastern Kentucky as the Kentucky River fault zone and in western Kentucky as the Rough Creek fault zone, which continues into southern Illinois as the Shawneetown and Cottage Grove fault zones. A western continuation includes the well known St. Genevieve fault zone and several other faults in southeastern Missouri. The faults are recorded in nearly horizontal Paleozoic rocks so that, although they may have significant strike-slip displacements, such displacements are difficult to detect. The presence of high-angle reverse faults and evidence of right-lateral movement in parts of the Rough Creek, Shawneetown, and Cottage Grove fault systems have been interpreted as evidence of wrench faulting (Heyl and Brock, 1961).

The fault zones include many normal or gravity faults, some with displacements of several hundred meters, as well as reverse faults. Large displacements are found in rocks as young as Pennsylvanian, and associated doming and mafic and alkalic intrusive rocks of Hicks Dome in southeastern Illinois are regarded as Permian (Heyl and others, 1965, p. B11). At the northern edge of the Mississippi embayment, some of these faults are covered by Upper Cretaceous sedimentary rocks of the Coastal Plain showing that most of the displacements had taken place before they were laid down. The fault zones seem to bear little tectonic relationship to the earlier Paleozoic arches and domes previously described; for example, the Kentucky River, Rough Creek, and Cottage Grove zones trend perpendicularly across the Cincinnati-Nashville arch and extend well into the sedimentary basins on either side. They represent quite a different tectonic process which probably occurred in the interval between Pennsylvanian and Late Cretaceous time.

A third set of high-angle faults crosses the trend of the Rough Creek and Cottage Grove fault zones in southeastern Illinois and extreme western Kentucky. These northeast-trending faults, known in southeastern Illinois as the Wabash Valley faults, form a zone about 60 km wide which is parallel to the axis of the Mississippi embayment syncline and is sharply defined against the east-west faults in western Kentucky and against the faults of the St. Genevieve system in southern Illinois (Heyl and others, 1965, pl. 2). Although some of the northeast-trending faults have been described as continuous and contemporary with the east-west faults, detailed investigation has shown that they are somewhat younger and more continuous than the east-west faults where the two sets intersect in the Illinois fluorite district (Weller, Grogan, and Tippie, 1952). From evidence (partly seismic) that some of the northeast-trending faults continue southwestward beneath the seismically active part of the embayment, Heyl and others (1965, p. B-5) have applied the name New Madrid fault zone to them.

Data on the pre-Cretaceous subsurface geology of the Mississippi embayment show that the Paleozoic rocks are arched over and eroded from a northwest-trending uplift (Pascola arch) between the Nashville dome and the Ozark uplift and that they may be considerably faulted as well (Freeman, 1945; 1953, pl. 5). The locations, trends, and displacements on these faults are not well known, but it is possible that both northeast-trending faults of the New Madrid system and northwest-trending faults related to the St. Genevieve fault and to the faults exposed in the Ozark uplift are present.

The northeast-trending faults, like some of the east-west faults, are overlapped by the Coastal Plain deposits of the embayment, but recent work has shown that movements on them continued well into Late Cretaceous and probably into Tertiary time (Ross, 1963). Thus, although the faults were well established by Late Cretaceous time, they appear to have been tectonically active during subsidence of the embayment syncline which began in the Late Cretaceous and continued into Eocene time. Topographic features and local structural evidence within the embayment suggest post-Pliocene movement on faults of this trend possibly as far south as the vicinity of Memphis, Tenn.

The other major region of high-angle faulting lies along the southeastern edge of the Canadian shield in the vicinity of the Adirondack Mountains and the Ottawa and St. Lawrence lowlands. Lower Paleozoic rocks on the margin of the shield appear in tilted fault blocks on the east flank of the Adirondacks and in the St. Lawrence lowland east of Montreal. They also occur in a west-trending fault and graben zone 96 km wide that penetrates the shield and separates the Adirondack uplift from the main part of the shield to the north. Middle Ordovician rocks are preserved on the down-dropped blocks, but the age of the subsidence is not well established. Dikes and stocks in and near the faulted areas range in age from early Paleozoic to Early Cretaceous, based on K-Ar dates (Doig and Barton, 1968). The development of topographic lowlands on the down-dropped blocks has resulted from differential erosion of the less resistant Paleozoic rocks and not from recent fault movements. The faults trend east-west or west-northwest and in places north-south, but they are generally markedly discordant to structural trends in the adjacent Precambrian rocks. King (1951, p. 22-23) concludes that they were formed subsequent to Paleozoic time and correlates them with post-Early Cretaceous faulting in the shield.

Coastal Plain subsidence

A group of Mesozoic and Tertiary structures is associated with subsidence of the continental borders along the Atlantic and Gulf coasts. For the most part, subsidence began in Cretaceous time and persisted throughout most of the Tertiary although the rate and amount of subsidence varied considerably in location and time during this period. The basement surface is relatively high and the Coastal Plain deposits thinner along the northwest-trending Cape Fear arch than in the bordering Coastal Plain embayments to the north and south. The sedimentary record indicates that the Cape Fear arch was tectonically positive during the Early Cretaceous and again during the Eocene, but that it has been essentially neutral relative to adjoining areas during the late Cenozoic. A similar uplift probably exists in the Norfolk, Va., region bordering the deep Salisbury embayment in Maryland and Delaware. Although this structure is smaller and less well known than the Cape Fear arch, Eocene faulting along it has been suggested on meager evidence (Cederstrom, 1945). Aside from this, the Atlantic Coastal Plain arches and embayments are not known to be separated from one another

by sharp structural or stratigraphic discontinuities that would indicate faulting.

High-angle faults that displace Coastal Plain strata have been observed in the Gulf Coastal Plain, where they are largely strike faults bounding depressed blocks. Subsurface evidence commonly indicates that these faults flatten with depth and have resulted from basinward sliding within the sedimentary pile. Because few if any earthquakes are associated with these faults, most are not shown on the accompanying tectonic map. Linear topographic depressions, displaced flood plains, and even fault scarps have been reported along the Baton Rouge fault zone (Durham and Peeples, 1956).

Known faults are much less common in the Atlantic Coastal Plain. Those that have been reported were observed very locally in artificial excavations or interpreted from local subsurface data. They are known to displace Coastal Plain strata from Late Cretaceous to Eocene or younger age, and some displace the basement rocks as well (Jacobeen, 1972; Glaser, 1971; Bruce O'Connor, oral commun., 1974). Linear surface features seen on aerial photographs or satellite imagery have been interpreted as representing fault traces in the Atlantic Coastal Plain, but evidence relating these features to observable displacements is mostly lacking.

The most significant of the Coastal Plain structures, from the standpoint of seismic activity, is related to the Mississippi embayment trough north of the Monroe uplift and Jackson dome. The record of the sediments deposited in the embayment shows that subsidence was relatively short lived, beginning in Late Cretaceous time and ending by Oligocene time; since then only about a hundred meters of Pleistocene and Holocene alluvial deposits have accumulated. The trend and extent of the basin deep into the central stable region indicate tectonic control rather different from that governing the broader, less linear basins in other parts of the Atlantic and Gulf Coastal Plains. The timing of the tectonic events involved suggests a connection with the Laramide orogeny of the Rocky Mountain region.

An apparent tectonic relationship exists between this pattern of crustal subsidence and the northeast-trending Wabash Valley faults directly along strike in southeastern Illinois. As mentioned previously, these faults appear to represent a somewhat younger, through-going system that transects the Rough Creek and St. Genevieve fault zones near the head of embayment. The northeast-trending faults are overlapped by the Coastal Plain deposits of the embayment, but recent work has shown that movements on the faults continued well into Cretaceous and probably into Tertiary time (Ross, 1963). Thus, although the faults were established by Late Cretaceous time, they appear to have been active tectonically during subsidence of the embayment syncline that began in Late Cretaceous and continued into Eocene time.

For an evaluation of regional seismicity, it is important to know what structural boundary if any exists between the Mississippi embayment and the seismically less active Gulf Coastal Plain to the south. This boundary seems to be marked by a series of transverse uplifts and mafic intrusive bodies in the embayment deposits in northern Louisiana, southeastern Arkansas, and adjacent parts of Mississippi. These features formed in Middle and Late Cretaceous time and served to restrict the embayment trough to the north, where deposition virtually ceased after early Tertiary time. These features indicate the existence of a tectonic boundary between the embayment trough and the region of continued deepening of the Gulf Coast geosyncline. If so, the boundary is considerably south of and seemingly independent of the buried Ouachita tectonic front, which passes beneath the deepest part of the trough.

Quaternary structures

The most recent structures, those of Quaternary age or younger, are of course particularly important in a study of the causes of historical seismicity. Several lines of evidence attest to crustal movements during the Quaternary, although quantitative data are scattered and incomplete. One of the best documented crustal movements is the record of upwarping throughout the northeastern part of the continent as a result of deglaciation and unloading of the crust during and after removal of the Labrador ice sheet. These movements were apparently limited to the region north of a line extending roughly from central Minnesota southeastward through Milwaukee, Detroit, New York, and Long Island, and northeastward to Nova Scotia. The central part of this region, in the Hudson Bay area, appears to be still rising relative to sea level at about 7 mm/yr. Isobases of equal uplift during the past 13,000 years and hinge lines marking successive northward shifts of the limit of warping trend approximately parallel to the outer glacial limit and concentric to the former ice cap.

Similar crustal warping, apparently unrelated to glacial unloading, is suggested by recently completed leveling traverses throughout the eastern United States, as reported by Meade (1971). These results indicate crustal updoming during the period 1930-1970 at a rate of 7 mm/yr centered in northwestern Georgia, as well as subsidence along the Atlantic coast north of Cape Hatteras reaching 5 to 6 mm/yr in New England. Elsewhere in the eastern United States, including Florida, changes of elevation appear to be minimal. Regional tilts indicated by the level measurements show a maximum of 6 mm/yr/100 km in New England and 3.5 mm/yr/100 km in northern Georgia and the Carolinas. In comparison, current tilt rates due to glacial rebound in the Great Lakes region are estimated at only 1.5 mm/yr/100 km (Flint, 1947, p. 425).

Very different evidence of Quaternary deformation is found in faults that offset surficial materials of late Cenozoic age. These include many examples of faults displacing geologically young features, such as alluvial deposits, till, loess, glaciated bedrock surfaces, and river terraces. Although such features have been widely reported from localities in the Appalachian Piedmont and Coastal Plain, most have been observed only in limited exposures such as road cuts and river banks in unconsolidated materials which are subject to rapid erosion and overgrowth of vegetation. Little information has been available as to their regional extent or tectonic significance. A reported exception is that of the glaciated bedrock surfaces in the Champlain-Hudson valley region; these surfaces show systematic reverse-fault displacements of a millimeter or more on southeast-dipping regional cleavage (Oliver, Johnson, and Dorman, 1970).

In the Mississippi embayment region, faults are known in rocks of Pliocene-Pleistocene age (Finch, 1966; Olive, 1963, 1965). Faults in loess and alluvium of Pleistocene age have been interpreted as occurring along topographic lineaments which may represent more extensive faults and fractures in the Coastal Plain sediments (Krinitzsky, 1950). For the most part, however, direct evidence indicating that faults are coincident with these features is lacking. With the exception of the vicinity of New Madrid, Mo., and Reelfoot Lake, Tenn., no faults or other indications of recent faulting are known to be associated with any historic earthquake.

SEISMIC AREAS

A study of the earthquake epicenter map B shows that large areas of background or minimal seismic activity are distinguished from epicentral concentrations representing areas that are seismically more active. The upper limit of activity in the background areas is arbitrarily taken as less than 8 epicenters per 10^4km^2 (in most places less than 4) and the absence of earthquakes with epicentral intensities greater than VI. Not all such areas are wholly aseismic, although some areas considerably larger than 10^4km^2 have generated no recorded earthquakes. In fact, areas with no epicenters within 10^4km^2 could be outlined, but this distinction is believed to be tectonically uninterpretable from a regional point of view.

For meaningful tectonic interpretation one must turn to the more seismic areas and these, classified according to level of seismic activity and degree of known structural control, are shown on map C. They are described in some detail in the following pages.

In establishing the different areas of earthquake activity shown on the seismotectonic map, use is made of three parameters. One is the areal density (seismic frequency) of epicenters, including separate events at the same location. A second is the maximum epicentral intensity, according to the Modified Mercalli scale (Wood and Neumann, 1931), of any earthquake known to have occurred in the given area. Although magnitude determinations are given for many earthquakes that have

occurred since 1960, they are generally not available for earlier events. Moreover, since it is expected that a principal use of the map will be to estimate potential earthquake damage, intensity values are considered to be more useful. The third parameter is the evidence of structural control on epicenter distribution as determined by the tectonic study. From these parameters five levels of seismic activity and three categories of structural control were selected (see map C explanation).

The seismic areas described herein are intended to represent areas designated by notable seismicity and, in so far as possible, the structural features believed to be casually related to the seismicity. In some areas notable seismic activity is not associated with any known structural feature, but is distributed so as to indicate rather clearly that the activity is localized along concealed or unidentified faults or other structures. Seismicity, usually of low level, is scattered in many areas so that no localizing fault or other structure is indicated. In a few places, such as central Virginia, central New England, and the vicinity of Charleston, S.C., seismic activity of intermediate or even high levels cannot, in the light of present information, be considered localized along known structures. In the southern Appalachian fold belt referred to above, the level of seismicity is more uniform, but it is nowhere well enough defined to indicate individually active structures.

It should be emphasized that *the seismic areas shown on the map do not have precisely determinable boundaries.* Few earthquakes in the eastern United States can be shown to be related to specific faults, as are many earthquakes in the west. Moreover scatter of epicenters and lack of correlation with geologic structure may be due in part to the scarcity of instrumental data and the resulting inaccuracy in locating epicenters. Where a good correlation seems to exist between seismic activity and known structures, or where the distribution of epicenters seems to outline a localizing fault zone or other structure, the seismotectonic map boundaries are drawn to conform rather closely to the geologic or seismological information available. Elsewhere boundaries are more generalized and essentially indicate areas where more information is needed to determine the relations between seismic activity and structure.

Seismic areas in the central stable region

Two areas in the central stable region of the continent that have been subject to unusual deformation in Paleozoic or later time are also areas of more than average historic seismicity. The more important of these two is centered in the Mississippi embayment trough and includes parts of western Tennessee, western Kentucky, southern Illinois, southeastern Missouri, and eastern Arkansas. Earthquakes in the southeastern part of this seismic area are believed to have been caused by movements on members of the northeast-trending New Madrid and Wabash River fault systems described previously. Such earthquakes include the New Madrid series of 1811-1812 and the subsequent shocks in the Mississippi embayment, the 1968 earthquake in southern Illinois, and others in the area northeast of the embayment. There is little evidence that historic movement has occurred on the faults of the east-trending Rough Creek fault zone in adjacent parts of Kentucky and Illinois.

The lateral boundaries of the New Madrid seismic zone are determined by geologic evidence of consistent fault trends, including the through-going northeast-trending faults in southern Illinois, and by surface evidence of recent (post-Pliocene) faulting. The northeastern limit of the zone does not appear to be structurally controlled and is arbitrarily placed where the seismic frequency drops to 8. The maximum seismic activity coincides approximately with the crestal part of the Pascola arch and also with a possible extension beneath the embayment of a group of northwest-trending faults west of the seismic zone. Scattered surface evidence of northwest-trending post-Eocene faults within the embayment (Farrar and McManamy, 1937) suggests that part of the seismic activity in the New Madrid area has resulted from movements on northwest-trending structures.

Although the St. Genevieve fault, as mapped, seems to have been largely inactive according to the seismic record, a group of epicenters including two of MM VII in southwestern Illinois suggests that earthquakes have been produced on a branching fault or faults on the southwest flank of the Illinois basin. There seems to be, however, no geologic record of a major fault with this location and orientation.

Evidence from fault plane solutions of 25 earthquakes recorded since 1962 in the Mississippi embayment and adjoining areas (Stauder and Nuttli, 1970; Street and Herrmann, 1974) indicates that northwest directed compressional fault movements dominate in the embayment north of the crest of the Pascola arch and also in southeastern Illinois. Earthquakes in the embayment south of the crest, however, indicate normal fault movements on north-trending faults. Strike-slip faulting was found in only two earthquakes, both along the northwestern edge of the New Madrid fault zone north-northwest of New Madrid, Mo. This evidence, if confirmed, would indicate that earthquakes near and northeast of New Madrid arise from tectonic mechanisms different from those controlling faulting farther southwest.

Several much less seismic areas within the central stable region deserve mention. The largest of these is in northwestern Ohio, where a northeast-trending cluster of epicenters includes those of the earthquakes near Anna, Ohio, and others of MM VII or higher. This seismic area lies a little west of the crestline of the Findlay and Cincinnati arches in an area where the arches broaden and merge westward with the less well defined Kankakee arch. The seismic plot shows that the epicenters are crowded, reaching 32 per 10^4km^2, in contrast to the broad and poorly defined Paleozoic structure at this point. The local concentration of epicenters and the marked northeast trend, which is about parallel to that of the Wabash River faults of southeastern Illinois, suggest control by a basement fault or faults related to but not necessarily continuous with the New Madrid fault system.

Two still smaller clusters of epicenters, involving much less seismic frequency and no earthquakes greater than MM VI, appear in northeastern Ohio east of Cleveland and in south-central Ohio. No specific structural control is indicated but the weak alinements suggest that they may have resulted from movements on minor faults.

An area of low regional seismic activity, which includes epicenters of two earthquakes of intensity VII, is located in western Illinois and southern Michigan. This is unusual in that earthquakes of this intensity generally have not occurred in areas whose seismic frequency is less than 8. The area is one of slightly deformed Paleozoic rocks, although its northern end, which is the locus of one of the larger earthquakes, is in a region of subsurface folds and high-angle faults in the Paleozoic rocks near the western end of the poorly defined Kankakee arch. Recorded earthquake epicenters, however, are too few to define what the structural control may be.

A small seismic area in western New York and adjoining Ontario suggests control by an unidentified westward-trending structure about parallel to the strike of the south-dipping Paleozoic rocks of the region. Earthquake intensities are low except for an earthquake of MM VIII near Attica, N.Y., at the eastern end of the area. This earthquake and some lesser ones nearby may be related to a buried north-trending fault that displaces the lower part of the Paleozoic sequence and probably the underlying basement; it is represented by a west-dipping monocline in the surface rocks known as the Clarendon-Linden structure (Chadwick, 1920, 1932). This structure trends north-south, and its relation to the east-west seismic trend is not clear.

The other principal region of historic earthquakes in the central stable region is located in northeastern New York and adjoining parts of Quebec; it extends northwestward beyond Ottawa and northeastward beyond Quebec City. The region has an exceptionally long history of recorded earthquakes, beginning in the 17th centry, and several of these have been rated as intensity IX or higher on the basis of contemporary but obviously rather limited accounts. The continued occurrence of earthquakes, however, and the high seismic frequency--more than 64 in the Montreal area and the area northeast of Quebec City--attest to unusual seismic activity in the region.

The region is characterized by high-angle faults belonging to an east-west or west-northwest-trending set that extends well into the Laurentian shield and a northeast-trending set limited mainly to the eastern

border of the shield along its boundary with the Appalachian tectonic system. Epicenter alinements indicate control by faults belonging to each of these sets of faults, although specific faults cannot be identified, and in the area of highest earthquake incidence, epicenters are so numerous that individual trends cannot even be detected. The distribution does show, however, that most of the earthquakes have occurred in the region of high-angle faults associated with relatively late disruption of the shield. There is little evidence that distinctively Precambrian structures are involved.

Seismic areas in the Appalachian tectonic province

Historic earthquakes have occurred throughout much of the Appalachian tectonic province in the eastern United States, and, in general, this province has been considerably more active than adjoining parts of the central stable region and the Coastal Plain.

Seismic activity in the faulted part of the Appalachian fold belt is characteristically higher than in the less faulted central part, and epicenter alinements in several places suggest that one or more of the faults may have been factors in releasing seismic energy. Most of the known faults are believed to extend to relatively shallow depths, probably less than 7 km, and two recent earthquakes in the belt are estimated to have originated within this limit (Bollinger, 1968; Bollinger and Hopper, 1970). In addition to having higher average seismic frequency, four of the five earthquakes of intensity VII that have occurred in the southern part of the Appalachian tectonic province originated in this faulted part of the fold belt. In contrast, the less faulted central part has shown relatively little seismic activity. Seismic frequency is less than 4 in most of it, and large areas have had less than 2 epicenters per 10^4km^2. The few earthquakes known provide no indication of tectonic control.

For seismotectonic purposes the Blue Ridge and Piedmont form a single unit with complex and incompletely known tectonic features. Within it there is much diversity of structure and of seismic activity but relatively little information on structural control. The most concentrated activity occurs in an east-trending area in central Virginia, within which the seismic frequency is 16 to 32 or higher. Most of the earthquakes in this area have been of intensity VI or less, although one near Richmond, Va., is reported to have been MM VI-VII (Bollinger and Hopper, 1971). The area consists largely of crystalline rocks with northeast-trending structures and includes parts of the James River and Arvonia synclinoria. It also spans three north-trending faulted basins of Triassic sedimentary rocks. Epicentral trends are not well defined, but their distribution indicates that most of the earthquakes originated in the crystalline rocks rather than in the Triassic basins or their bordering faults. The seismic area terminates near the edge of the Coastal Plain, which is puzzling since one would expect greater macroseismic effects in the Coastal Plain sediments than in the crystalline rocks of the Piedmont. The overall trend of the seismic area crosses the regional structure of the near-surface rocks; perhaps it represents a deeper crustal structure of east-west trend, but such a structure is not shown by most geophysical data.

Another area of more than average seismicity in the Blue Ridge-Piedmont province covers parts of western North Carolina and northwestern South Carolina. Structurally the area includes part of the Murphy syncline, much of the great allochthonous terrane of Blue Ridge rocks with its large negative gravity anomaly, and segments of the Brevard fault zone. Weak epicentral alinements suggest that longitudinal faults within the Murphy syncline may have controlled a few earthquakes; other alinements may be related to east-west transverse displacements that have been observed in the Blue Ridge part of the seismic area. A few epicenters are located close to the Brevard fault zone, which may have had a local influence on the distribution of macroseismic effects of nearby earthquakes, but over most of its extent the fault zone has been notably lacking in seismic activity. There is also no evidence that the Triassic(?) fault-breccia zones in Polk County, N.C., and in Greeneville and Spartanburg Counties, S.C., have been sources of historic earthquakes.

Northeast of central Virginia, the Piedmont belt is an area of low-level seismic activity with no strong earthquakes except one of intensity VII near the head of Delaware Bay. A cluster of epicenters in the eastern part of the Gettysburg Triassic basin suggests, but does not strongly indicate, movement on faults within or bordering the basin. Elsewhere in the Appalachian Piedmont and Blue Ridge regions, earthquake frequency is less than 8, and in central and northern Georgia it is less than 2. An exception to this may be developing in central South Carolina, where several minor earthquakes have occurred recently (Bollinger, 1972).

The structural geology of the New England-Maritime region is dominated by north- and northeast-trending folds, faults, elongated intrusive bodies, and metamorphic features. These trends are reflected also in extensive and detailed aeromagnetic data of the region and to some extent in gravity data. Special significance is attached to a series of northeast-trending faults in central and southwestern Connecticut; these faults cut Triassic rocks and appear in a zone of prominent magnetic lineaments extending northeastward into Massachusetts and southeastern New Hampshire (P.J. Barosh, written commun., 1973). While displacements on these faults took place before Triassic time, they appear to have remained active during the Triassic, and epicentral alinements suggest that they may have been a source of earthquakes in historic time. Minor earthquakes recently recorded by the Lamont seismograph network have been interpreted as originating along the Ramapo fault, which marks the northwestern edge of the Newark Triassic basin (Page, Molnar, and Oliver, 1968).

Seismic activity of more than background frequency and intensity is confined to the southeastern part of the region in Connecticut, eastern Massachusetts, southern New Hampshire, and southern Maine. Seismic frequency is unusually high as compared with the rest of the Appalachian system, although this may have resulted in part from the longer period during which earthquakes have been recorded. The lack of resolution in locating and plotting epicenters, combined with relatively high seismic frequency, makes it especially difficult to correlate earthquakes with specific structures in this part of the Appalachian region, despite the availability of better structural data. The most likely correlation seems to be between faulted terranes and northeast-trending epicentral alinements, although a small area of epicenters along the international boundary in southeastern Maine fits known faults of northwest trend. Otherwise, much of the seismic activity is classed without correlation to known structures.

Seismic areas in the Atlantic Coastal Plain

Although rocks and structures typical of the Blue Ridge and Piedmont provinces extend beneath the Atlantic and Gulf Coastal Plains, seismic frequency in the Coastal Plains is much lower than in the Appalachian belts to the west, except locally in areas adjacent to seismically active parts of those belts. Low seismic activity is also characteristic of more distant parts of the Coastal Plain, such as those on the Florida platform. This is true even where the Coastal Plain sequence is considerably faulted, as in the Gulf Coastal Plain of southern Arkansas, Mississippi, and southwestern Alabama. Most of these faults are thought to have been caused by seaward sliding within the sedimentary sequence; whose rocks are not strong enough to accumulate large strains and thus generate only minor earthquakes. This situation probably explains why few earthquakes strong enough to be locally reported or teleseismically located have occurred in large parts of the Coastal Plain.

A highly anomalous exception to the characteristic seismic inactivity of most of the Coastal Plain is the well known record of earthquakes in the vicinity of Charleston, S.C. This record includes several unusual features that distinguish it from seismic activity elsewhere in the eastern United States. The intensity IX-X earthquake of 1886 seems to have occurred after a centry or more of little seismic activity in the region, and the many subsequent shocks that occurred within a 20- to 30-year period all took place in the area of the original earthquake; thus, they have been referred to as a long sequence of aftershocks of the 1886 event. No other part of the Coastal Plain or other locality in the eastern United States shows a similar record.

The wide extent of high-intensity effects of the 1886 earthquake, as well as the large region in which it was felt, indicate that it originated at considerable depth in the crustal rocks below the Coastal Plain sedimentary

cover. The lithologic character and structure of the basement rocks in the region are, however, poorly known, and there is very little evidence to indicate what kind of structure was involved. The seismic area shown on the map is therefore merely a representation of the areal frequency of many imprecisely located epicenters reported in the Charleston area. The computer plot of epicenters indicates a weak northwest trend that may or may not reflect real structural conditions.

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