Intraplate Seismicity, Reactivation of Preexisting Zones of Weakness, Alkaline Magmatism, and Other Tectonism Postdating Continental Fragmentation

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The distribution of intraplate earthquakes and of igneous rocks postdating continental rifting is summarized and placed into a plate tectonic framework for the following continental areas: eastern and central North America, Africa, Australia, Brazil, Greenland, Antarctica, Norway, Spitsbergen, India, and the margins of the Red Sea and Gulf of Aden. In continents, intraplate earthquakes tend to be concentrated along preexisting zones of weakness within areas affected by the youngest major orogenesis that predates the opening of the present oceans. Many preexisting zones of weakness (including fault zones, suture zones, failed rifts, and other tectonic boundaries), particularly those near continental margins, were reactivated during the early stages of continental separation. In contrast, intraplate shocks rarely occur within the older oceanic lithosphere or within the interiors of ancient cratonic blocks of the continents. In several continental areas, rocks and tectonic features postdating the opening of present-day oceans, including carbonatites, kimberlites, other alkalic rocks, mafic dikes, and ring dikes, as well as some of the largest intraplate shocks, seem to be located preferentially along old zones of weakness near the ends of major oceanic transform faults that were active in the early opening of adjacent oceans. In several places, alkaline magmatism and earthquakes extend several hundred kilometers inland from the ends of oceanic transform faults (but not necessarily with the same strike as the transform fault). Major preexisting zones of weakness that are oriented subparallel to the directions of relative continental separation appear to control the locations of transform faults that develop in a new ocean. In some instances, alkaline magmatism persisted along reactivated features of this type for as long as 100 m.y. after the initial stages of continental fragmentation. Most kimberlites in South Africa seem to have been emplaced along preexisting zones of weakness that were reactivated during the early opening of the South Atlantic. The type of intraplate magmatism appears to be related to the thickness of the lithosphere. Unlike oceanic transform faults where large horizontal movements have occurred, reactivated zones of weakness in continents usually appear to have been the sites of only relatively small displacement. Seismic activity and alkaline magmatism may be controlled by deep fractures that penetrated the entire lithosphere to tap asthenospheric sources of magma. Seismic activity along these zones seems to occur in response to the present-day stress regime, which is not necessarily the same as that which was active during the emplacement of the alkaline rocks. Other intraplate shocks are concentrated along old zones of weakness that are subparallel to continental margins. Such shocks are found in the Appalachians, northeastern and northern Greenland, Norway, Great Britain, Spitsbergen, northern Canada, and Australia. These zones of weakness were also reactivated during continental separation in either the Mesozoic or the Cenozoic. Evidence is now mounting for Cretaceous and Cenozoic deformation along some of these features. Although not many focal mechanism solutions or in situ measurements of stress are available for intraplate areas, horizontal compressive stresses appear to be present today in many of the pre-Mesozoic orogenic belts that were reactivated by continental rifting. This evidence, as well as examples of Cenozoic thrust faulting, indicates that the stress field has changed since rifting commenced. High compressive stresses, the absence of earthquakes in Antarctica, their near absence along the margins of the Gulf of Mexico, and the much lower levels of activity in the oceanic lithosphere adjacent to most continents argue against mere sedimentary loading and the cooling of the oceanic lithosphere as the main source of stress that is reactivating faults of these older fold belts. The large compressive stresses and the uplift found in many continental areas adjacent to continental margins may be caused by a deep-seated source in the mantle of long wavelength or by stresses transmitted in the lithosphere. These effects may be related to either the cooling and underplating of the continental lithosphere adjacent to continental margins, large tractions on the base of the lithosphere in shield areas, stress concentrations related to marked changes in the age and thickness of the lithosphere, convective motions of the mantle beneath these areas, or those regions acting like broad zones of weakness that are being compressed between adjacent areas of greater strength. During the fragmentation of a supercontinent, multibranched rifting usually follows the youngest zone of previous orogenesis and as much as possible avoids passing through old cratonic areas where the lithosphere is thick, cold, and strong. Rift junctions seem to be related to the preexisting mosaic of cratons and younger belts of deformation rather than to a motive force involving mantle plumes. Likewise, many zones of unusually high magmatic activity, i.e., hot spots, appear to be related to nodes or junctions in this mosaic pattern. Thus these hot spots appear to be passive features rather than the surficial expression of mantle plumes. Major transform faults that are active during the early opening of an ocean also tend to develop where the margins of the older cratons undergo an abrupt change in strike. During the early development of an ocean the preexisting mosaic of structural elements within the thick continental lithosphere may result in large normal forces across some plate margins, leaky transform faulting, and localized stress concentrations. The early directions of sea floor spreading and of transforming faulting may be altered by these boundary forces and by the geometrical constraints imposed in separating old cratonic blocks. These constraints are relaxed once old, thick lithosphere is no longer in contact across long transform faults. Since these early directions are strongly influenced by the

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preexisting tectonic framework and may not coincide with the direction of the forces driving the plates apart, early transform faults may have components of extension (or compression) along them in addition to strike slip motion. A small component of extension may be responsible for the formation of volcanic ridges and seamount chains such as the Walvis ridge, Rio Grande rise, and New England seamount chain. These features predate the marked change in the strike of transform faulting that occurred in the North and South Atlantic about 80 m.y. ago as thin oceanic lithosphere finally came in contact across large oceanic transform faults. Several zones of intraplate magmatism in the surrounding continents also ceased at that time.

INTRODUCTION

Although our understanding of the tectonic processes at lithospheric plate boundaries has increased tremendously in the last 10 years, little is known about the present tectonic regime within the interiors of plates. In fact, most simple theories of plate tectonics assume an absence or near absence of deformation within plates. While this gross approximation is useful in comparing, say, the high seismic activity of the Aleutian arc with the lower level of activity in the eastern United States, intraplate seismic activity nevertheless accounts for a few percent of the world's total earthquakes. Also, a number of large and damaging earthquakes have occurred historically in areas so remote from plate boundaries that they can only be classified as intraplate phenomena. These include shocks of magnitude larger than about 6 in central and eastern North America (Figure 1), peninsular India, western and southern Africa, Australia, and several oceanic areas remote from plate boundaries. In contrast, intraplate earthquakes are very rare in Antarctica, interior Greenland, and much of the Canadian shield. Many of the earthquakes in Central Asia, China, and other parts of the broad zone of seismic activity that extends from the Mediterranean to southern Asia are not easily categorized as either intraplate, interplate (where the plate boundary is diffuse and wide), or activity related to a nearby plate boundary. This more questionable category of shocks is not discussed in detail here.

Until a few years ago, intraplate earthquakes had received very little study. In the eastern and central United States, for example, seismograph stations were typically spaced about 300 km apart. For much of the United States east of the Rocky Mountains, almost no detailed studies had been made of parameters such as focal depth, earthquake mechanism, stress drop, state of stress, and possible relationship of earthquakes to faults or to other geological and geophysical features. In fact, many workers assumed that earthquakes in the eastern United States were related to faults even though the evidence for such an association was almost nonexistent. In eastern North America, there are no clearly documented instances of fault displacement at the surface during earthquakes even though several large and damaging shocks have occurred in that region. While evidence is beginning to mount relating earthquakes to faults in areas that were recently instrumented with dense networks of seismographs, such as that operated by Lamont-Doherty in New York State and vicinity [*Fletcher and Sykes*, 1977; *Sbar and Sykes*, 1977; *Yang and Aggarwal*, 1977; *Aggarwal*, 1977; *Aggarwal and Sykes*, 1978], for most intraplate regions it is still not known whether earthquakes do in fact occur along faults and if they do, whether these faults are relatively short or are continuous features.

Some workers have proposed that seismic activity in eastern and central North America is concentrated along particular zones or trends, while others have treated earthquakes as occurring randomly in space. For example, there has been considerable debate about whether a zone of seismic activity extends from the St. Lawrence region to the southwest as far as the New Madrid area of southeastern Missouri. *Sbar and Sykes* [1973], on the other hand, argue for a northwest trending zone of activity that extends from offshore Massachusetts across New England to Ottawa and thence into the Canadian shield as far as Kirkland Lake, Ontario. Other workers propose that a zone of earthquakes follows the Appalachian Mountain system, while others infer that a belt of activity extends northwest across South Carolina from Charleston into eastern Tennessee [*Woollard*, 1958; *Richter*, 1959; *Hodgson*, 1965; *P. B. King*, 1970].

This paper focuses on several continental areas that are remote from plate boundaries but close to continental margins. The historic record of earthquakes in most of these areas, however, is no longer than 100–300 years. Thus historical data and the pattern of earthquakes by themselves do not appear to be adequate at present to answer questions such as, Is the Charleston, South Carolina, earthquake of 1886 related to a particular tectonic structure, or could a future shock of that size occur with equal probability anywhere in eastern and central North America? For this reason, various geological data, such as the presence of igneous rocks that postdate the opening of the Atlantic Ocean and the location of old fault zones, are examined in the context of defining zones of higher seismic risk. Since the number of known large earthquakes in eastern and central North America is relatively small, some principles governing their setting may well be developed by studying intraplate shocks on a global basis. Other areas that have rifted apart by continental fragmentation in the last 200 m.y., such as western and southern Africa and southern Australia, appear to provide clues to a better understanding of seismic risk and recent tectonism in eastern and central North America. One of the main theses of this paper is that many intraplate earthquakes and other types of intraplate tectonism can be placed within a plate tectonic framework.

Figure 2 summarizes schematically several of the major points developed in this paper about the development of continental margins, the emplacement of alkaline rocks, the reactivation of preexisting faults, the occurrence of earthquakes, and the locations of sedimentary basins along margins.

The East African rift system may be one of the best present-day examples of a continent in the process of fragmentation. *Vail* [1967], *Cox* [1970], *McConnell* [1972], and others point out that the various rift zones of East Africa tend to be located along the margins of the older cratonic nuclei and to follow younger belts of deformation. These margins of the older cratonic areas of Africa appear to have been fragmented by continental drift several times in their geologic history and to have subsequently been the sites of the closure of an ocean as they eventually collided with other continental fragments. *Wilson* [1966] concluded that new oceans, such as the Atlantic, tend to develop along or near the suture zone, where an ancestral ocean was closed during the last major orogeny in that area. Hence the belt of most recent orogenesis, usually one of Paleozoic or late Precambrian age, may be thought of as a broad zone of weakness that is reactivated and split apart during the development of a new ocean. Since the strength of

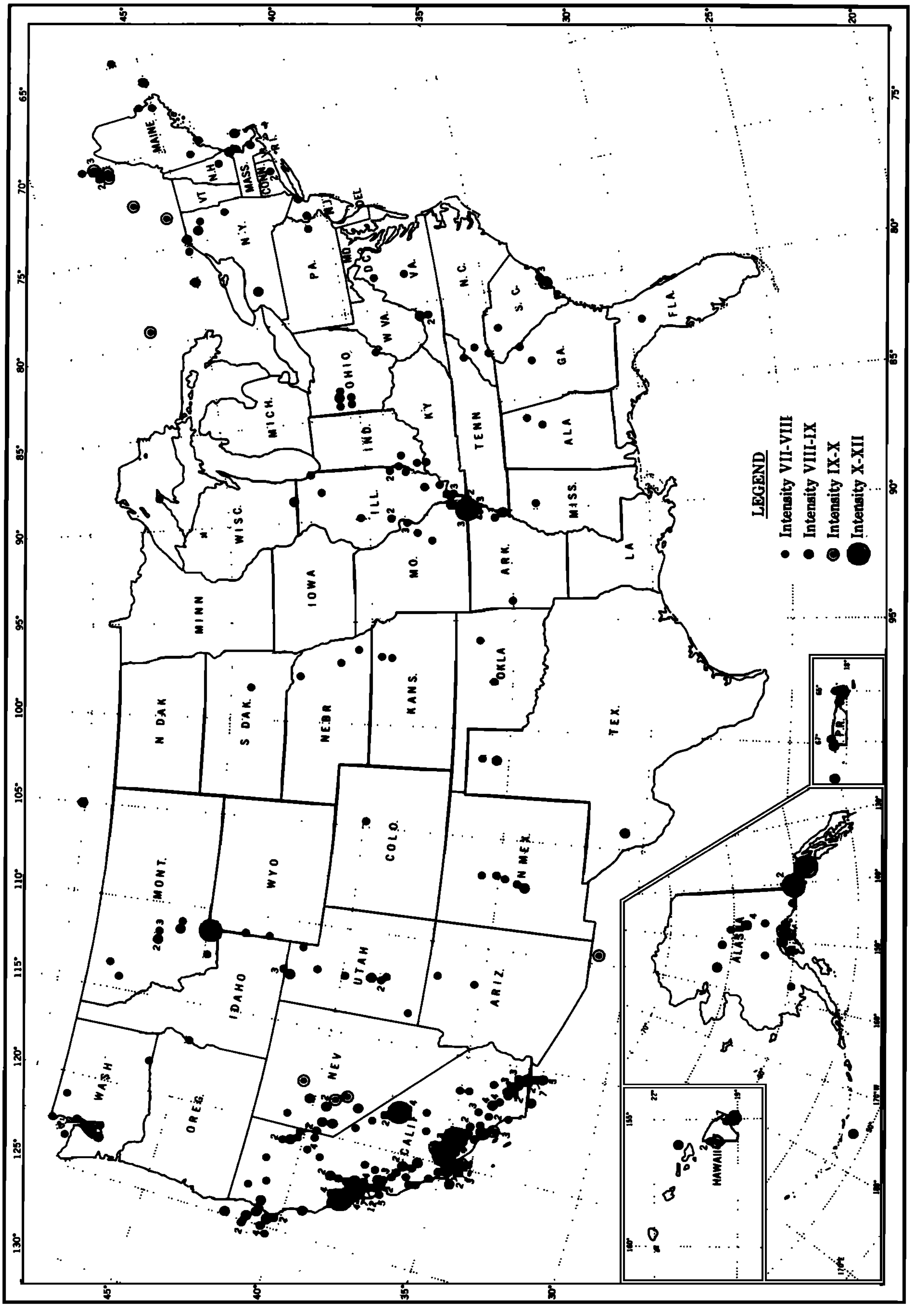

Fig. 1. Damaging earthquakes of the United States through 1969 [after *von Hake and Cloud*, 1969].

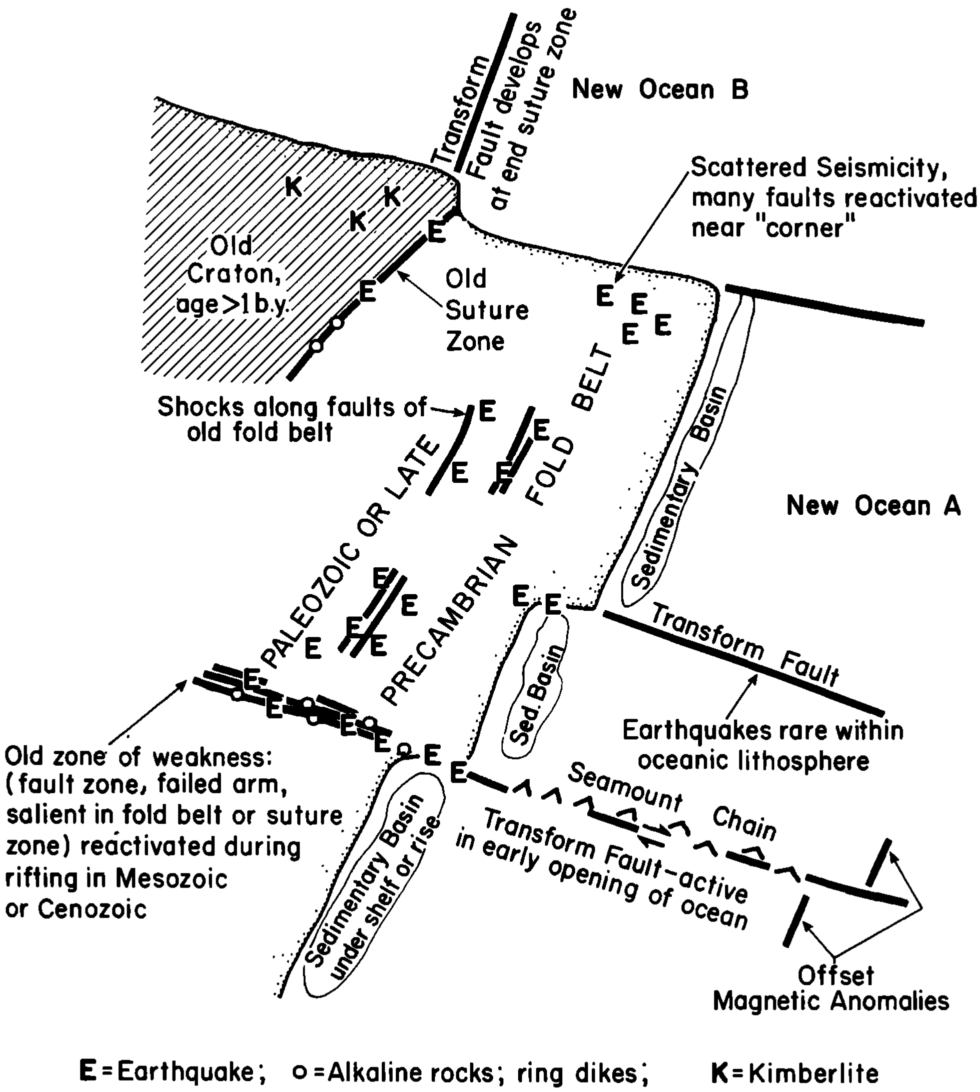

Fig. 2. Schematic drawing of configuration of continental margins developed by continental fragmentation and sea floor spreading. Locations of transform faults (fracture zones) in new oceans are controlled by preexisting zones of weakness in fold belts of Paleozoic or late Precambrian age. These zones of weakness, which are now oriented almost perpendicular to continental margins, and other zones of weakness, now oriented nearly parallel to margins, were reactivated during early development of margins and are active now, judged from locations of earthquakes. Kimberlites (K) are injected into deep fault zones of old craton (hatched area) during opening of ocean B.

fractured rock is much less than that of intact rock, one is tempted to look for various types of preexisting zones of weakness, such as fault zones, suture zones, continental rifts, aulacogens, and monoclines, as having a pronounced influence on the shape of subsequent continental margins, the locations of oceanic transform faults, the emplacement of igneous rocks, and the loci of earthquakes.

In *Wilson*'s [1965] conception of a transform fault, strike slip motion today is confined to that portion of the fault located between two spreading ridges. Those portions of the fault that are situated beyond the two spreading centers were believed to be inactive, since they exhibit very little (if any) seismic activity. These inactive portions appear to represent flow lines of former sea floor growth. *Wilson* [1965] proposes that the apparent offset across several large fracture zones (or transform faults) in the equatorial Atlantic was 'born' into the initial pattern of continental rifting, since the offset of the present axis of the mid-Atlantic ridge is nearly the same as the offset of the margins of either West Africa or Brazil. He also hypothesizes that major fracture zones tend to develop near major preexisting zones of weakness in the continents.

Large oceanic transform faults exerted a pronounced influence on the development of sedimentary basins along several continental margins (Figure 2). *Francheteau and Le Pichon* [1972], *Dingle and Scrutton* [1974], *Gorini and Bryan* [1976], and *Sheridan* [1974, 1976] find that sedimentary basins along the coasts of Africa and North and South America are often sharply bounded by major transform faults. Some basins (and presumably the adjacent parts of the continent) have subsided more than other areas of the continental margin. The spacing of transform faults and the blocklike nature of continental margins appear to be reflected in the present-day pattern of earthquake activity (Figure 2).

One of the principal results of this study is that intraplate earthquakes in several continental regions do not seem to occur randomly in space. Seismic activity in several continental areas is high near the ends of several major fracture zones (Figure 2) that were active in the early fragmentation of a continent and in the development of nascent oceans. Other zones of weakness in the continents appear to control the location at which fracture zones develop in a new ocean. These older zones are reactivated by continental fragmentation and still appear to be moderately active today, judging from the location of large intraplate earthquakes.

The presence of alkalic rocks, ring dikes and diatremes (Figure 2) along zones extending inland from the ends of several oceanic transform faults is indicative of deep-seated sources of magma which probably originated in the uppermost mantle at depths shallower than 200 km. In this paper the term alkaline rocks is taken to include silica-undersaturated igneous rocks of basic and intermediate composition and their derivatives. In many cases, regions of enhanced seismicity and alkaline magmatism are situated along preexisting faults that appear to have been reactivated during continental rifting. The probable deep origin of the alkaline rocks implies that these faults extended (and probably still extend) through the entire thickness of the continental lithosphere. Unlike the ocean floor, where large horizontal offsets have occurred along transform faults, reactivated fault zones within continents probably involve little strike slip offset and only a small amount of horizontal extension during the most recent period of continental rifting and sea floor spreading.

The alkaline rocks found along these reactivated fault zones within continents are quite different from those associated with most plate boundaries. In fact, the only plate boundaries characterized by alkaline magmatism are oceanic transform faults and areas of continental rifting such as East Africa [*Barberi et al.*, 1974; *Bonatti et al.*, 1976]. *Barberi et al.* [1974] find that tholeiitic or intermediate basalts characterize the volcanism of axial rift zones in the Afar triangle of eastern Africa, whereas alkaline basalts, which frequently contain inclusions of peridotites, are found along structures oriented transverse to the rift zones. Alkalic basalts are also characteristic of the earliest stages of volcanism along the rift zones themselves (E. Bonatti, personal communication, 1977). In describing the evolution of the Afro-Arabian dome, *Gass* [1970] concludes that alkalic flood basalts were erupted during an early phase of continental rifting within areas of unattenuated continental crust, whereas tholeiitic magmas were emplaced in the center of the Red Sea and in the Gulf of Aden as the latter opened by sea floor spreading and the emplacement of new oceanic crust. For the Afro-Arabian dome, Gass concludes that alkalic magmatism originates from partial melting at a depth of about 60 km, which is within the uppermost mantle. With the increased development of a lithothermal event, tholeiitic magmas result from a greater degree of partial melting and may originate at depths as shallow as 10 km. Similarly, tholeiitic magmatism is characteristic of the central rift zones of the midoceanic rift system [*Engel and Engel*, 1964], whereas alkaline magmatism is generally confined to seamounts or islands. The latter rocks appear to originate off the axis of the ridge system from partial melting at a depth of about 60 km [*Menard*, 1969].

In discussing the petrogenesis of the carbonatite-alkaline rocks of Mesozoic age in the Lupata-Lebombo area of East Africa, *Woolley and Garson* [1970] conclude that tholeiitic, alkaline olivine basalt and nephelinite-carbonatite magmas originate at respectively greater depths in the upper mantle. Deep faults facilitated their emplacement near the surface. Likewise, *Dawson* [1970] concludes that nondiamondiferous kimberlites in Africa are found mainly within the major cratons and along adjacent fold belts of Kibaran (1100 ± 200 m.y.) age. Diamondiferous kimberlites, which appear to originate at even greater depths in the upper mantle, are confined to the older cratonic areas, where the lithosphere is presumably thick and relatively cold. *Dawson* [1970] finds that kimberlites were often intruded into major fundamental fractures that cut both the cratonic areas and their surrounding Precambrian fold belts during periods of epeirogenic uplift. In southern Africa the most recent episode of kimberlite emplacement occurred during the Cretaceous as the South Atlantic started to open.

Thus the occurrence of kimberlites, carbonatites, nephelinites, melilitites, monchiquites, syenites, alkali granites, and various mafic dikes may be indicative of deep fault zones that extend through the lithosphere. The correlation of presumed depth of origin with the age of the surrounding tectonic province suggests that the type of magmatism is controlled by the age and thickness of the lithosphere through which these magmas penetrated.

This paper focuses on those types of intraplate magmatism that appear to originate within the mantle and which may be indicative of the occurrence of deep faults. Hence more widespread basaltic magmatism, which generally occurs within a relatively short time interval (a few tens of millions of years) during continental fragmentation and during the emplacement of oceanic crust, is not considered here. Only those igneous rocks that postdate continental fragmentation or those of more unusual petrology are discussed. More widespread igneous rocks, such as the basalts and diabases of the Newark series of eastern North America and the Kaoko and Serra Geral basalts of South-West Africa and South America, respectively, whose ages are nearly contemporaneous with continental rifting, are not considered further.

This paper considerably expands the study of *Fletcher et al.* [1978], in which they argue that several zones of seismic activity in eastern and central North America appear to be located along continental extensions of major transform faults. This paper expands upon the seismic evidence of *Fletcher et al.* [1978] and *Sbar and Sykes* [1977] for eastern North America, adds new evidence from the distribution of alkaline rocks, and discusses similar zones of seismic activity and alkaline rocks in other continents. *Fletcher et al.* [1978], however, were not aware of the presence of older zones of weakness in the continents near the ends of many oceanic transform faults. Hence they describe seismic activity as being located along continental extensions of oceanic transform faults. It was difficult for them to devise a mechanism for extending deformation inland from these transforms. To me a more satisfying explanation is that the continental zones of earthquakes and alkaline magmatism are situated along older zones of weakness inherited from the last major orogeny. These zones were reactivated during continental rifting; several remained active as zones of alkaline magmatism for as long as 100 m.y. after initial rifting, and many are now the loci for the release of intraplate stresses during earthquakes. Thus in many ways it is misleading to call these zones continental extensions of oceanic transform faults.

When the distribution of earthquakes in various intraplate areas is examined, however, it is clear that many other earthquakes do not appear to be located landward of major fracture zones. Thus it is necessary to invoke some other type of mechanism or tectonic setting to explain these shocks. Many of these earthquakes are located in older fold belts (Figure 2) adjacent to continental margins that developed in either the

Mesozoic or the Cenozoic. In these fold belts, major preexisting faults, which are now oriented subparallel to the continental margin, also appear to have been reactivated by continental separation and are the loci of earthquakes today. Some intraplate earthquakes also appear to be localized along rift zones (failed arms) that did not develop into full ocean basins [*Burke and Dewey*, 1973].

This paper places considerable emphasis on the occurrence of seismic activity and of rocks that postdate continental fragmentation along major preexisting zones of weakness of various type in the continents. It differs in a number of ways from studies such as those of *Burke and Dewey* [1973], that emphasize failed rifts, triple junctions, and mantle plumes as active elements in the fragmentation of a supercontinent and in the development of ocean basins. Other workers ascribe zones of intraplate magmatism, such as the White Mountain Magma Series in New England and the Monteregian Hills in southern Quebec, to the movement of hot spots. Unlike the clear spatial progression of rock ages along the Hawaiian chain, the published age data for most zones of continental magmatism do not show such a clear migration pattern. Instead, magmatic activity typically occurs for tens of millions of years and then shuts off as it appears to have done in New England and along the New England seamount chain about 80–100 m.y. ago. Magmatism has continued to the present time along the Cameroon line. Thus for these rocks it is difficult to accept an origin related to a moving hot spot. If preexisting zones of weakness in continents play a major role in controlling the sites of magmatism and the locations at which oceanic transforms develop, it seems reasonable to view at least certain hot spots as merely passive centers of increased volcanism. These centers owe their existence to zones of weakness, either on the ocean floor or in the continents, along which magmas can reach the surface if the stress tensor for that area is favorably oriented.

Ideas about continental fragmentation, the occurrence of intraplate seismicity, and the state of stress within plates have a possible bearing upon a number of scientific questions such as the following: What is the source of the stress field? Has it changed with time? Is it residual from the Paleozoic or Precambrian? Are these phenomena related to the driving mechanism of plate tectonics? These ideas also have many social and economic ramifications. For example, most of the nuclear power reactors in the United States (Figure 3) are concentrated in the eastern one third of the country, and future facilities are likely to be placed there. From a seismic risk point of view it is obviously critical to know if large and damaging shocks such as those near Charleston, South Carolina, in 1886; off Cape Ann, Massachusetts, in 1638 and 1755; and in the Mississippi Valley in 1811 and 1812 could occur anywhere in eastern North America or, instead, if they are confined to particular tectonic zones. A knowledge of seismic risk and of the state of stress in the lithosphere is also becoming of increasingly greater interest in the construction of high dams, large underground excavations, facilities for the disposal of fluid waste underground, high-pressure hydraulic mining, and other critical structures. These factors become of greater importance as the population grows and society becomes more complex technologically.

Major fracture zones that intersect continental margins, reactivated faults zones on land, and the distribution of young alkalic rocks may also have considerable economic importance. Structural features near the ends of fracture zones may serve as traps for petroleum. Also, magmatism postdating continental separation may heat organic material of Mesozoic age sufficiently so as to develop a petroleum resource. Metalliferous deposits and diamonds may also be related to some of these features.

TECTONISM IN WEST AFRICA NEAR ENDS OF GREAT EQUATORIAL FRACTURE ZONES

Continuity Across Atlantic Ocean

Heezen and Tharp [1965] recognized a number of great fracture zones that intersect the mid-Atlantic ridge in the equatorial region. Several of these zones are now mapped [*Fail et al.*, 1970; *Mascle and Sibuet*, 1974; *Delteil et al.*, 1974; *Gorini and Bryan*, 1976] almost continuously from Africa to Brazil. Since the amount of apparent offset across a fracture zone appears to be nearly the same for times from the initial opening to the present, and since individual fracture zones can be traced from Africa to Brazil, the equatorial Atlantic seems to have opened in a relatively simple manner compared to the more complex sequence of opening in the North Atlantic. In fact, at least two different centers of rotation are needed to describe the entire sequence of opening of the Atlantic between South America and Africa [*Le Pichon and Hayes*, 1971]. Nevertheless, the history of opening of the North Atlantic appears to be more complicated and to have commenced about 40 m.y. before the opening of the South Atlantic. Some of the major transform faults that were active in the Mesozoic opening of the North Atlantic, such as that along the New England seamount chain, did not continue to be sites of major offset during the Cenozoic. Also, some of the equatorial fracture zones offset the continental margin hundreds of kilometers, whereas such large offsets are rare in the North Atlantic. Hence along the margins of the equatorial Atlantic it is much easier to search for possible tectonic features and earthquakes near the ends of major fracture zones.

Gorini and Bryan [1976] mapped (Figure 4) several of the large equatorial fracture zones and studied the continental areas where those fracture zones intersect the margins of Africa and Brazil. They and *Francheteau and Le Pichon* [1972] conclude that these fracture zones segmented the continental margins such that sedimentary basins on either side of major fracture zones evolved independently. Near the present ridge crest these fracture zones strike somewhat north of east, whereas farther east off the coast of Africa the strike is northeasterly, i.e., parallel to several long, linear, steep segments of the continental margin, which appear to be fault controlled.

Transform Directions for Africa

Francheteau and Le Pichon [1972] use the term transform direction for lines of constant latitude about a center of rotation describing the relative motion of two lithospheric plates. In Figure 5, transform directions were calculated for the west coast of Africa south of 5°N by using the center of rotation of *Le Pichon and Hayes* [1971] for the movement of Africa with respect to South America from 130 to 81 m.y. *Mascle and Sibuet* [1974] computed a somewhat different center of rotation for the early opening of the South Atlantic from the one used in this paper. Nevertheless, their directions do not differ by more than about 10° from most of those of this paper. North of 5°N, transform directions were calculated by using the center of rotation for the Mesozoic rotation of North America with respect to Africa.

The major fracture zones in the equatorial region near the coast of Africa are well known from bathymetric surveys

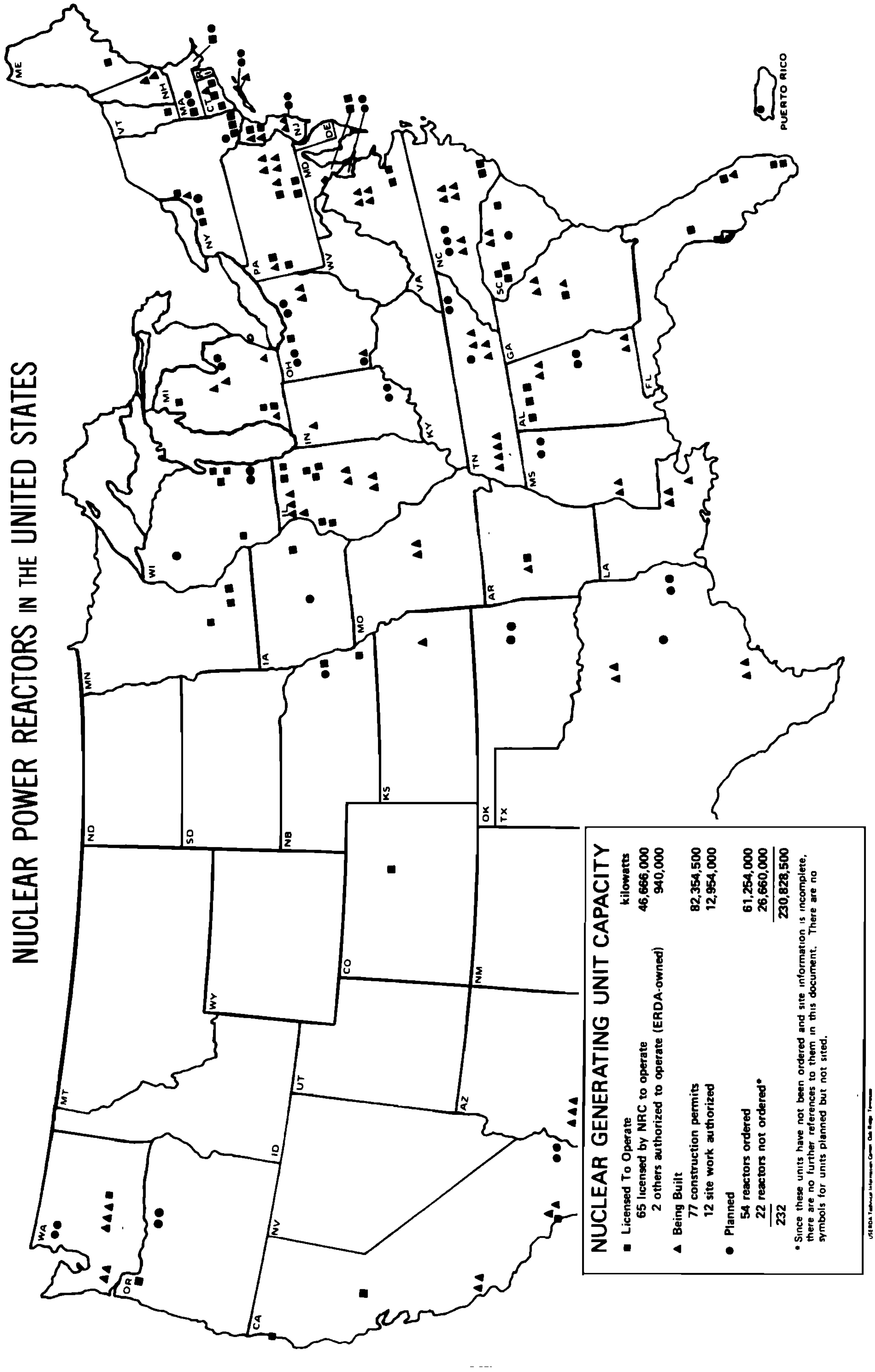

Fig. 3. Nuclear power reactors in the United States (after U.S. Department of Energy). Note the concentration in the eastern third of the country.

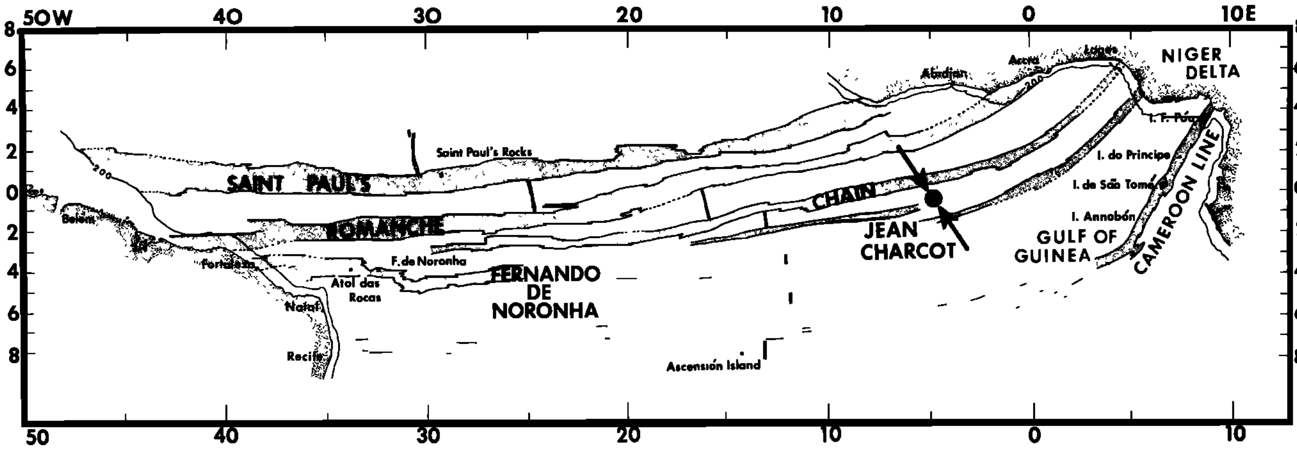

Fig. 4. Large fracture zones in the equatorial Atlantic as mapped from Brazil to Africa, after *Gorini and Bryan* [1976] with additions. Double lines represent the axis of the mid-Atlantic ridge. Solid circle and arrows denote the epicenter of the earthquake of September 30, 1971, and the axis of maximum compression (*P* axis) from the focal mechanism of *Sykes and Sbar* [1974].

[*Mascle and Sibuet,* 1974; *Delteil et al.,* 1974; *Gorini and Bryan,* 1976], as are those that offset the continental margin of Angola between 10° and 15°S and the Agulhas fracture zone, which forms the southeastern continental margin of South Africa [*Francheteau and Le Pichon,* 1972]. Three major offsets in the magnetic anomaly pattern of Mesozoic age near the continental margins of South Africa and South-West Africa were mapped by *Rabinowitz* [1976]. In Figure 5, continuous shaded lines are drawn through these major offsets and are extended a few hundred kilometers inland along transform directions. Since information on fracture zones near the coast of Africa between 0° and 10°S is poor, possible offsets of the present crest of the mid-Atlantic ridge as identified from the locations of earthquakes were extrapolated into Africa by two appropriate finite rotations. These transform directions are indicated by dashed shaded lines in Figure 5. The main purpose of extending these transform directions into the continents is to see either if structures, young rocks, and earthquakes tend to be localized where these features intersect the continental margins or if they seem to have some expression farther inland. This expression, if it exists, undoubtedly will be more complex than a single narrow line with the same strike as the oceanic transform.

In the Gulf of Guinea the continental margin of West Africa is offset about 800 and 600 km, where it is intersected by the Romanche and St. Paul's fracture zones, respectively (Figures 4, 5, and 6). The straight northwestern margin of the Dahomey basin in Figure 6 is located along the strike of the Romanche fracture zone (rather than the Chain fracture zone as *Burke* [1969*a*, *b*] infers). The Benue trough (Figure 5) is located along a transform direction extrapolated into the continent from the Charcot fracture zone. A pronounced gravity high [*Hospers,* 1965; *Gorini and Bryan,* 1976] is found in the Niger delta along the strike of this fracture zone. A deep-seated structural high in the Benue trough, the Abakaliki anticlinorium, which also strikes northeast, is also located along a transform direction extended inland from the Charcot fracture zone [*Gorini and Bryan,* 1976].

Seismicity of West Africa Near the Gulf of Guinea

Figure 5 compares the epicenters of earthquakes for the period 1900–1973, transform directions extrapolated into the continent from oceanic transform faults, and rocks either postdating or approximately equal to the age of opening of the South Atlantic. Earthquakes of magnitude 6 or greater are shown as large solid circles. While the seismicity of the west coast of Africa is low in comparison to that of the East African rift system or to that of most plate boundaries, several large and damaging earthquakes have nevertheless occurred near the ends of some major fracture zones.

The seismicity of the Accra area of Ghana is one of the best examples of activity near the end of a fracture zone and of the reactivation of a major pre-Mesozoic tectonic feature. The isoseismals (Figure 6) of the earthquakes of 1906 and 1939 near Accra indicate that these shocks occurred either along or near the steep northeast trending continental margin, which appears to be an extension of the Romanche fracture zone. The epicenter computer by the International Seismological Summary for the 1939 shock (cross in Figure 6) falls almost exactly on this scarp. Seventeen people were killed and 1000 buildings destroyed in the earthquake of 1939 [*Junner,* 1941; *Gorshkov,* 1963], which was of magnitude 6½ [*Gutenberg and Richter,* 1954] and had a maximum intensity of IX (Modified Mercalli). The 1939 shock appears to have been less severe than the one in 1862, which almost completely destroyed Accra, but more severe than the one in 1906 [*Junner,* 1941]. The large felt areas of the earthquakes of 1906 and 1939 are similar to those of the large historic shocks in central and eastern North America.

Burke [1969*a*, *b*, 1971] states that the Séhoué and Lokossa faults in the Dahomey basin (Figure 6) have each experienced about 100 m of vertical movement. He concludes that the Akwapim fault in Ghana and the Séhoué fault both show evidence of Recent movement. *Burke* [1969*b*] believes that the shocks of 1906 and 1939 occurred on either the Akwapim fault or a subparallel fault about 6 km to the northwest. In 1939 a fissure nearly 10 km long opened along the latter fault. *Tevendale* [1957] concludes that the Akwapim and Toga scarps in this area were formed by faulting. Displacements of up to 130 m have occurred along them since the Cretaceous. He cites the occurrence of earthquakes and the morphology of the Volta River as evidence of faulting during Tertiary to Recent time.

Junner [1941] states that the Accra area experienced major earthquakes in 1636, 1862, 1906, and 1939 and at least 12 smaller shocks between 1858 and 1935. He also mentions that

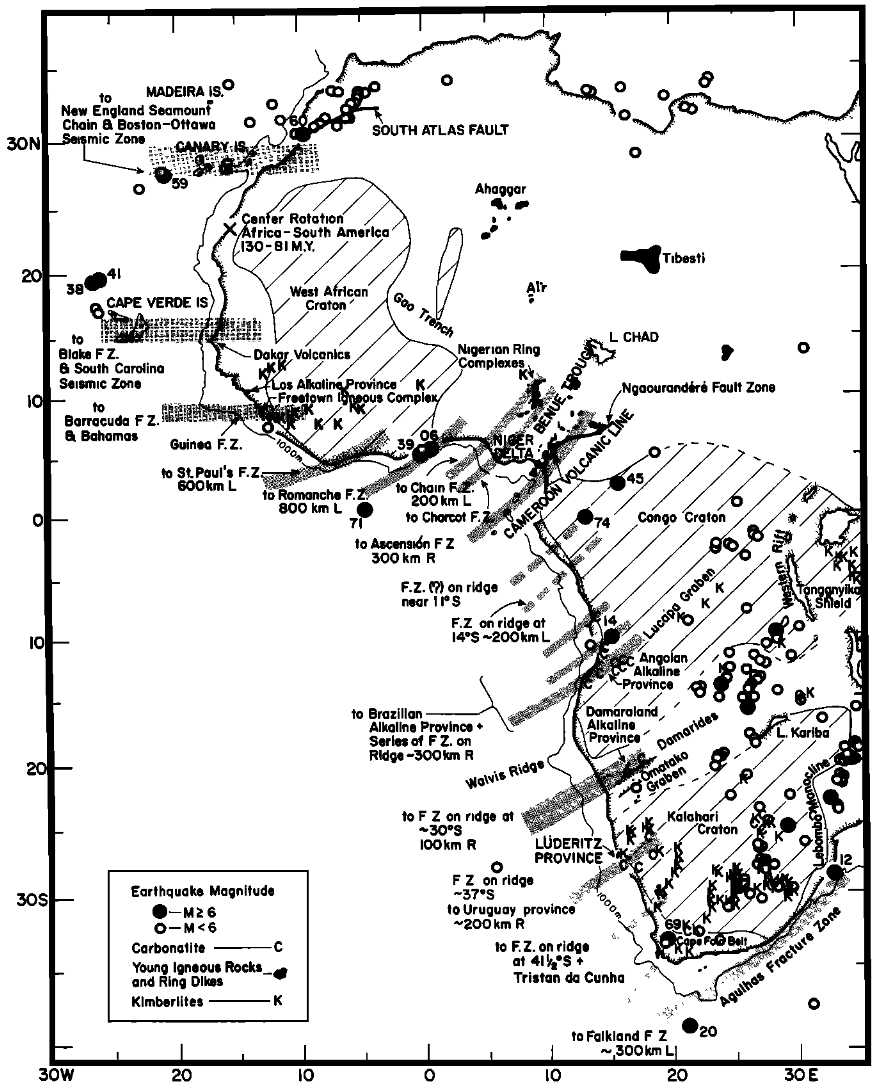

Fig. 5. Earthquakes along Atlantic margin of northern and southern Africa, 1900–1975. Carbonatites, kimberlites, ring dikes, and alkaline igneous rocks [from *Furon*, 1963; *Crockett and Mason*, 1968; *Dawson*, 1970; *Briden et al.*, 1971; *Marsh*, 1973; *van Breemen and Bowden*, 1973; *Cornelissen and Verwoerd*, 1975; *Williams and Williams*, 1977] that are younger than or about equal to the age of separation of Africa and North America are shown along with small circles (transform directions) drawn about the center of rotation for either the movement of Africa with respect to South America from 125 to 81 m.y. (south of 5°N) or the movement of Africa with respect to North America from 81 to 180 m.y. (north of 7°N). Centers of rotation are from *Le Pichon and Hayes* [1971]. Small circles (continuous shaded lines) are drawn such that they pass along either mapped fracture zones in the Gulf of Guinea [*Gorini and Bryan*, 1976], offsets of continental margin [*Francheteau and Le Pichon*, 1972], prominent offsets in the pattern of magnetic anomalies of Mesozic age south of 20°S near the continental margin [*Rabinowitz*, 1976], or zones of recent volcanism in the Canary and Cape Verde islands. Small circles denoted by dashed shaded lines are more questionable. Seismic data are from *Gutenberg and Richter* [1954], *Rothé* [1969], *Sykes* [1970*b*], and recent computer locations of NOAA and USGS. Solid circles are shocks of magnitude 6.0 or greater. Epicenters are omitted near the western rift zone of East Africa and north of 34°N. Several additional earthquakes in Morocco from 1932 to 1951 from Debrach (cited by *Gorshkov* [1963]) are added. Last two digits of date in years are shown beside larger earthquakes that are discussed in the text. The Canary Islands and the Cape Verde Islands have conjugate points in North America near the New England seamount chain and Charleston, South Carolina, in a reconstruction of Africa and North America that closes the Atlantic. The amount of apparent offset at the present ridge axis is indicated in kilometers and as being either left (L) or right (R) lateral. Other features at conjugate points in North America and South America are indicated. Note that a number of kimberlites, carbonatites, other alkalic rocks, and earthquakes are located near the ends of major oceanic transform faults. Cratonic areas (hatching) older than 800 m.y. are from *Kennedy* [1965], *Black and Girod* [1970], *Dawson* [1970], *Grant* [1973], and *Hurley and Rand* [1973].

several earthquakes were recorded by Milne seismographs at Accra during the period 1914 to 1933 when it was in operation and that very slight tremors were noted almost daily.

Describing a seismic reflection profile, *Blundell and Banson* [1975] and *Blundell* [1976] conclude that a major east striking fault passes a few miles offshore from Accra and intersects the Akwapim fault just to the west of Accra. It then extends northeast as the faulted northern margin of the Dahomey basin. *Blundell* [1976] finds no evidence of a major fault at the continental margin east of Accra near the epicenter shown in Figure 6. He cites past earthquakes and recordings made since 1973 by the Kukurantumi seismograph station as evidence of current activity along the Akwapim fault and the east striking coastal boundary fault. The latter passes between Accra and

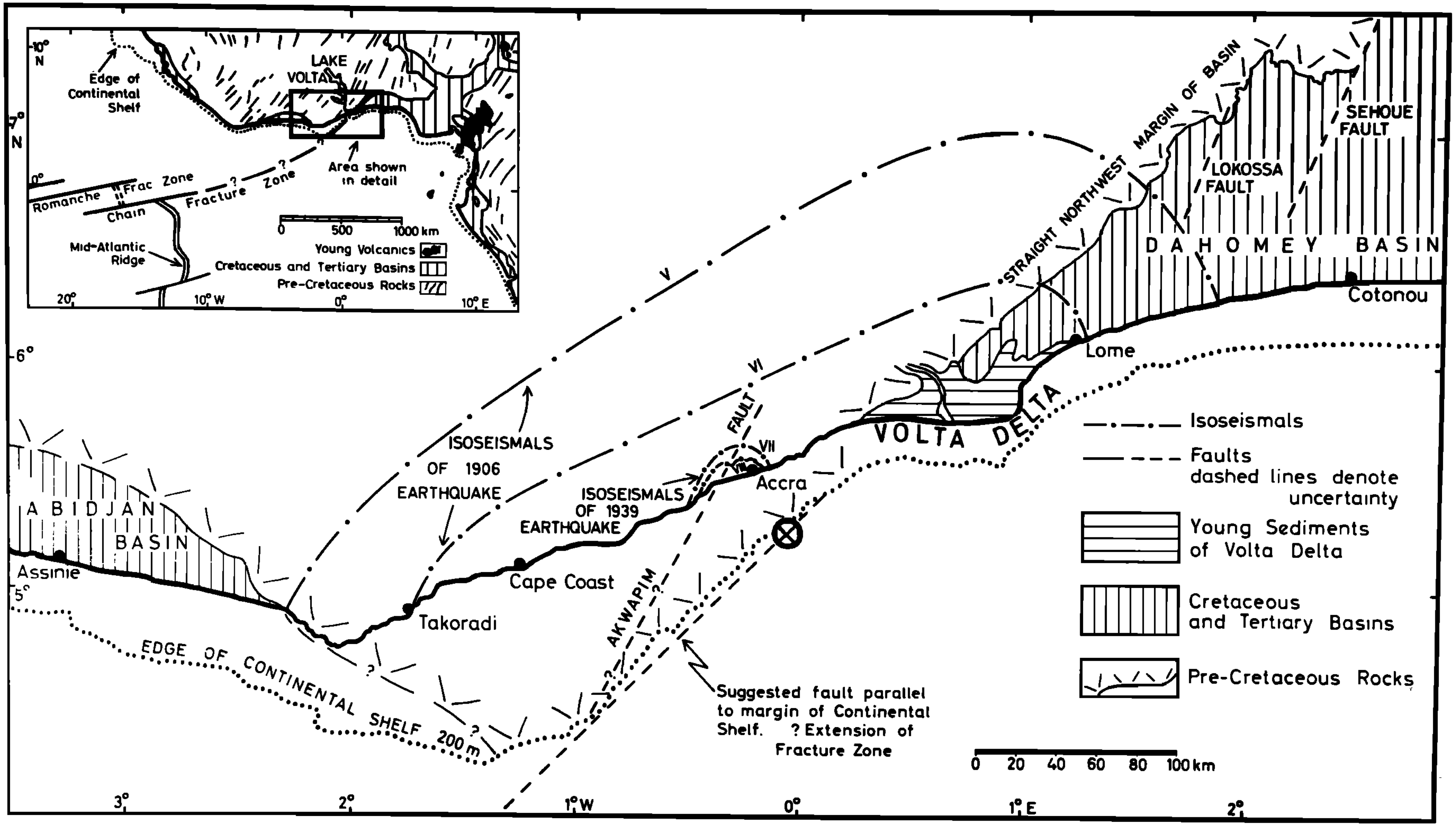

Fig. 6. Structural features near the coast of West Africa in the Gulf of Guinea related to extensions of a northeast trending fracture zone, after *Burke* [1969*b*] with additions. The fracture zone extending into the region probably is the Romanche, not the Chain, fracture zone as Burke suggests. Isoseismals are shown for earthquakes of 1906 and 1939. The straight northwestern margin of the Dahomey basin appears to lie along the continental extension of the same fracture zone. The circle with a cross is the epicenter of the 1939 earthquake as computed by the International Seismological Summary.

the epicenter indicated in Figure 6. Probably, the teleseismic data used to locate the 1939 shock are not accurate enough to assign it unambiguously to one of the above mentioned faults. Regardless of which specific faults are active, seismic activity is present in the continent near the end of the Romanche fracture zone.

Two large earthquakes on September 12, 1945, and September 23, 1974, are located on or close to a transform direction extrapolated inland from a fracture zone on the mid-Atlantic ridge near 11°S (Figure 5). The 1945 shock of magnitude 6 had a maximum felt intensity of VII and a macroseismic area of 200,000 km^2 [*Gutenberg and Richter*, 1954; *Gorshkov*, 1963]. For the earthquakes near Accra, however, it is easier to make a case for an association with a fracture zone, since these shocks were located near the continental margin. Since the shocks of 1945 and 1974 are located well inland and the fracture zone of the ridge near 11°S has not been mapped near the coast of Africa, the relationship of these earthquakes to an inferred continental extension of a fracture zone is speculative.

A focal mechanism solution involving predominantly thrust faulting was obtained (Figure 7) for the 1974 shock of magnitude 6.2, which occurred in Gabon within the Congo craton (Figure 5). Although the two nodal planes are not well determined, they strike northwest and are not parallel to the transform directions of Figure 5. This is the only mechanism solution known to me for western Africa. The northwesterly strike of the two nodal planes is nearly parallel to that of the continental margin. Thus the mechanism solution indicates that the 1974 earthquake probably has no relationship to the inferred transform direction shown in Figure 5.

Longer History of Felt Earthquakes

It is evident from the historic record that the period 1900–1973 for which instrumental locations of earthquakes are available (Figure 5) does not accurately portray the long-term seismicity of the west coast of Africa. *Krenkel*'s [1923] map of the average annual frequency of felt earthquakes in Africa (Figure 8) shows a zone of moderate seismic activity in West Africa near the Gulf of Guinea. He indicates that the highest activity in that area is found near the Cameroon line. In his Figure 50, *Gorshkov* [1963] reproduces a map of *Sieberg* [1932] showing the areas of perception of five widely felt earthquakes in Cameroon from 1903 to 1913. These shocks were located on or close to the Cameroon line.

Suture Zone Near Accra

In examining the seismicity of West Africa in Figure 5 the Accra region has been active both instrumentally and historically; no earthquakes have been detected instrumentally near the end of the St. Paul's fracture zone, although historic activity in the area is apparent in Figure 8. The eastern end of the Romanche fracture zone is located near the southern end of a major tectonic boundary that passes near Accra. Across the latter, which is of more northerly strike (about NNE), rocks of Pan-African age (450–750 m.y.) are juxtaposed against rocks of the West African craton (Figure 5) of 1950 m.y. age or older [*Kennedy*, 1965; *Grant*, 1973]. The pronounced gravity anomalies along this boundary suggest that it may represent a suture zone of late Precambrian–early Paleozoic age. Thus this major structural boundary appears to have controlled the location (but not the strike) of the Romanche fracture zone. At least a part of it may have been reactivated during rifting between South America and Africa. The higher known seismicity of the Accra region compared to that near the end of the St. Paul's fracture zone may be related to rejuvenation of movement along this ancient tectonic boundary. Such a well-developed zone of weakness does not appear to be present near the end of the St. Paul's fracture zone. Thus the presence or absence of older tectonic boundaries may be a guide to the relative seismic risk of areas within continents near the ends of oceanic transform faults.

Young Igneous Rocks in Equatorial Africa and Gulf of Guinea

Several areas of widespread igneous activity of Jurassic to Recent age are found (Figure 5) near the margins of the West African craton [*Furon*, 1963; *Black and Girod*, 1970; *Burke and Wilson*, 1972]. Most of these rocks have a pronounced alkaline affinity [*Black and Girod*, 1970]. They include the Nigerian ring complexes, the Cameroon volcanic line, and young rocks associated with the Benue trough, all of which are located between the older West African and Congo cratons in a broad area that was affected by the Pan-African orogenic event. In contrast, the West African craton itself has experienced very little magmatism of Mesozoic or more recent age. Other alkaline igneous rocks are found much farther to the northwest near Dakar and in the Cape Verde and Canary islands (Figure 5). Hence it seems most likely that the spatial pattern of young igneous rocks is largely controlled by a mosaic of tectonic elements that consists of thick lithosphere beneath the older (>2 b.y.) cratons and relatively thin deformable lithosphere beneath the surrounding orogenic belts of Pan-African or younger age. In this sense the emplacement of the young igneous rocks is viewed as a passive phenomenon rather than as the surficial manifestation of mantle plumes, as *Burke and Dewey* [1973] and others propose.

Nigerian ring complexes. The younger granites and ring dikes of Nigeria are similar in many ways to the White Mountain magma series in New England [*Rhodes*, 1971]. Both have a pronounced alkaline character. As in New Hampshire, ring dikes are the most striking feature of the Nigerian province. In New Hampshire the oldest rocks of the White Mountain magma series (about 200 m.y.) are a few tens of millions of years older than the oldest oceanic crust dated thus far in the western Atlantic; the youngest are about 100 m.y. [*Foland and Faul*, 1977]. The younger granites and syenites of Nigeria have an age range of 156–167 m.y. [*Jacobson et al.*, 1964; *van Breemen and Bowden*, 1973], which is about 30 m.y. older than the onset of sea floor spreading in the South Atlantic as identified by magnetic anomalies [*Larson and Ladd*, 1973; *Ladd et al.*, 1973] but younger than the oldest sea floor in the North Atlantic. The Nigerian rocks may reflect an early stage of rifting between Africa and South America that preceded the development of the first oceanic crust by a few tens of millions of years. Such an interval of continental rifting prior to the development of oceanic crust appears to be common in many areas of the world. Isotopic studies indicate a mantle origin for the younger granites of Nigeria and for many of the younger igneous rocks found near the margins of the West African craton [*Bowden and van Breemen*, 1972; *Burke and Whiteman*, 1973].

Cameroon volcanic line. The Cameroon volcanic line, which extends for about 1100 km in a northeasterly direction across the Gulf of Guinea and into Africa, is one of the major tectonic features of Africa (Figures 4 and 5). The tectonic character and rock types along the line [*Burke et al.*, 1971] do not change appreciably in crossing from an oceanic to a continental environment. Mt. Cameroon, which is located along

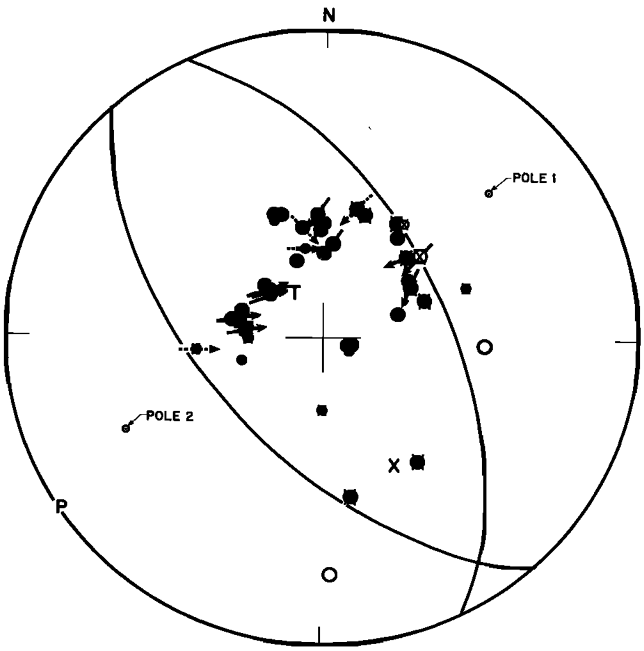

Fig. 7. Focal mechanism solution for earthquake of magnitude 6.2 in Gabon, west central Africa, on September 23, 1974. Diagram is an equal-area projection of the lower hemisphere. Open circles indicate dilatations; closed circles compressions, and crosses arrivals judged to be near a nodal plane. Smaller symbols denote questionable readings. Arrows indicate polarization of *S* waves. Nonorthogonal nodal planes were fit to the data, but a solution involving orthogonal planes probably can be found if the velocity structure of the crust and upper mantle is allowed to vary. *T* and *P* are inferred axes of minimum and maximum compressive stress. Azimuths and plunges are as follows: *T*, 328°, 74°; *P*, 236°, 01°; pole 1, 48°, 30°; pole 2, 244°, 32°. The mechanism is predominantly thrust faulting, but the strikes of nodal planes and the azimuth of the *P* axis are not well determined.

this line near the Gulf of Guinea, is an active volcano [*Gutenberg and Richter*, 1954]. Radiometric ages for the Cameroon line range from 35 m.y. to Recent [*Burke and Wilson*, 1972; *Gorini and Bryan*, 1976]. A seismic section on the continental shelf shows that the Cameroon line has formed an effective barrier to sedimentation on either side of it since the Cretaceous [*Reyre*, 1966; *Gorini and Bryan*, 1976]. Thus this tectonic line appears to have been in existence since the initial opening of the South Atlantic.

The present Cameroon line appears to involve the reactivation of the Precambrian Ngaourandéré fault zone (Figure 5) of Cameroon during the early opening of the South Atlantic [*Gorini and Bryan*, 1976]. In a pre-Mesozoic reconstruction of Africa and South America this fault zone appears to be continuous with the Pernambuco lineament, which trends west (Figures 4 and 9) near Recife, Brazil [*Gorini and Bryan*, 1976]. The latter is also of Precambrian age and was reactivated during the early opening of the South Atlantic. The Cameroon line and the Ngaourandéré fault zone are situated (Figure 5) near the boundary between the Congo craton and a belt of Pan-African deformation that extends as far west as Accra, Ghana [*McConnell*, 1969].

Not enough bathymetric mapping has been done to ascertain if the Ascensión fracture zone, a major offset at the crest of the mid-Atlantic ridge (Figure 4), can be traced into the Cameroon line or if it intersects the coast of Africa 100–200 km south of that line. The trend of the Cameroon line is about 15° more northerly than the strikes of transform directions in Figure 5 and about 25° more northerly than the trend obtained by *Mascle and Sibuet* [1974].

Benue trough. The Benue trough of Nigeria is located inland from the end of the Charcot fracture zone, which appears to extend under the Niger delta (Figure 5). Over 5 km of Cretaceous sediments are found in the trough [*Black and Girod*, 1970]. This feature appears to have originated as a tensional graben system during the Cretaceous from about 102 to 80 m.y. and to have experienced a period of compressional deformation at the end of the Cretaceous [*Wright*, 1968; *Burke et al.*, 1971; *Burke and Dewey*, 1974]. *Wright* [1968] believes that the trough and its associated sediments were formed in response to high stresses at the point where the initial coast lines of Africa and South America changed from predominantly east-west to predominantly north-south at the time that the two continents were being wedged apart. He associated the period of compression at the end of the Cretaceous with the final separation of the two continents as Africa south of the Benue trough tended to swing back with respect to northern Africa.

Burke et al. [1971] and *Burke and Dewey* [1973, 1974] interpret the Benue trough as a failed arm (failed rift) of a Cretaceous ridge-ridge-ridge (RRR) triple junction. The other two rifts continued to develop as the Atlantic opened. Citing evidence from a broad region of Africa northeast of the Benue trough, *Burke and Dewey* [1974] conclude that Africa did not behave as a single rigid plate during the Cretaceous. Nevertheless, the measurable shortening in the Benue trough is only about 5 km, although it is not possible to assess the amount of oceanic crust that might have been consumed by subduction [*Burke et al.*, 1971]. In the Gulf of Guinea, however, there is no evidence of large distortions of the ocean floor that might have been associated with a subduction zone extending from the mid-Atlantic ridge into the Benue trough (Figure 4).

A focal mechanism solution for an earthquake in the Gulf of Guinea (Figure 4) shows thrust faulting with the maximum compressive stress oriented nearly horizontal and northwest [*Sykes and Sbar*, 1974]. With present bathymetric mapping it is difficult to ascertain whether the earthquake was located on one of the branches of the Charcot fracture zone or it occurred between the Charcot and the Chain fracture zone (Figure 4). While this focal mechanism solution might be interpreted as a release of stress that was related to the closing of the Benue trough, other interpretations are possible. Horizontal compression is commonly found in many intraplate regions. Since large distortions on the ocean floor in the Gulf of Guinea do not appear to be present, it seems reasonable to conclude that large amounts of oceanic crust were not consumed in the Benue trough at the end of the Cretaceous.

Kennedy [1965], *Sutton* [1968], and others point out that much of the present western continental margin of Africa developed along zones of Pan-African (600 ± 150 m.y.) or younger deformation. The presence of Mesozoic rifting and magmatism on both the western and the eastern side of the West African craton suggests that multibranched rifting (perhaps like that in East Africa on the two sides of the Tanganyika shield), initially occurred on both sides of the West African craton but that rifting on the west side continued to develop into a single oceanic plate boundary. The Gao trench (Figure 5) on the eastern side of the craton appears to be a rift structure of Cretaceous age, which may have been an incipient, but abortive, marginal oceanic fracture system [*Kennedy*, 1965; *McConnell*, 1969].

The presence of Cenozoic magmatism along the east side of

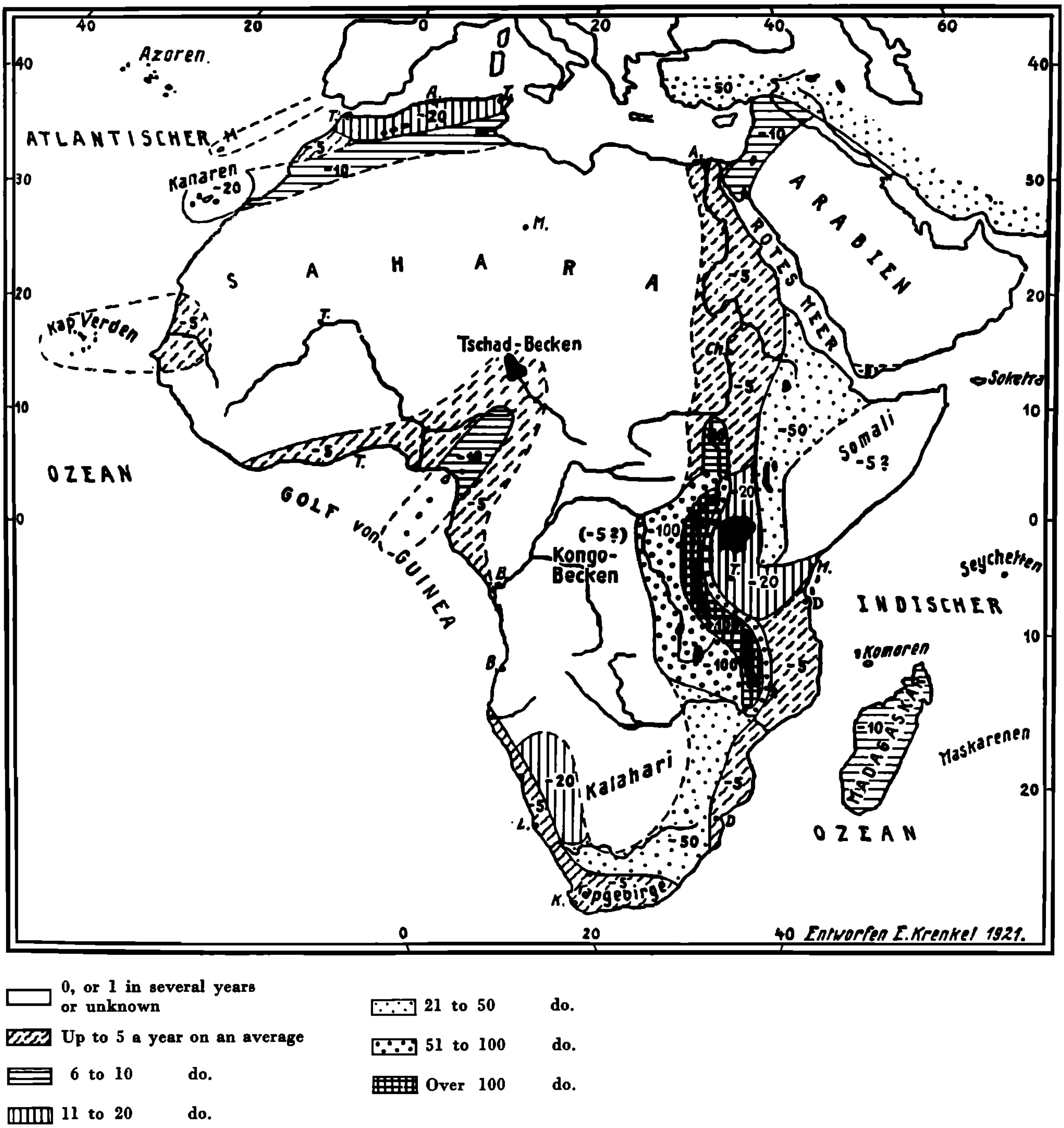

Fig. 8. Average annual frequency of earthquakes in Africa [after *Krenkel*, 1923]. In addition to higher levels of activity associated with the East African rift system and with the plate boundary in northwestern Africa, note the higher levels of activity near the Canary Islands, Cape Verde Islands, Cameroon line in the Gulf of Guinea, and near the Damaraland alkaline province and Omatako graben in South-West Africa (Figure 5).

the West African craton near the Benue trough, along the Cameroon line, and in the Ahaggar, Aïr, and Tibesti areas (Figure 5) suggests that preexisting fault zones within that Pan-African terrain, which were reactivated during the early separation of North and South America from Africa, continued to be zones of intraplate magmatism long after continental separation occurred. This is not too surprising, since it is unlikely that reactivated faults in those areas were healed by metamorphism since the Mesozoic. Continued magmatism along such fault zones may be largely controlled by the orientation of the lithospheric stress tensor.

The present Atlantic Ocean developed near the south side of the West African craton. A series of great fracture zones connected zones of rifting separated by about 3000 km along the west sides of the West African and Congo cratons (Figure 5). Thus it seems reasonable to associate the presence of these

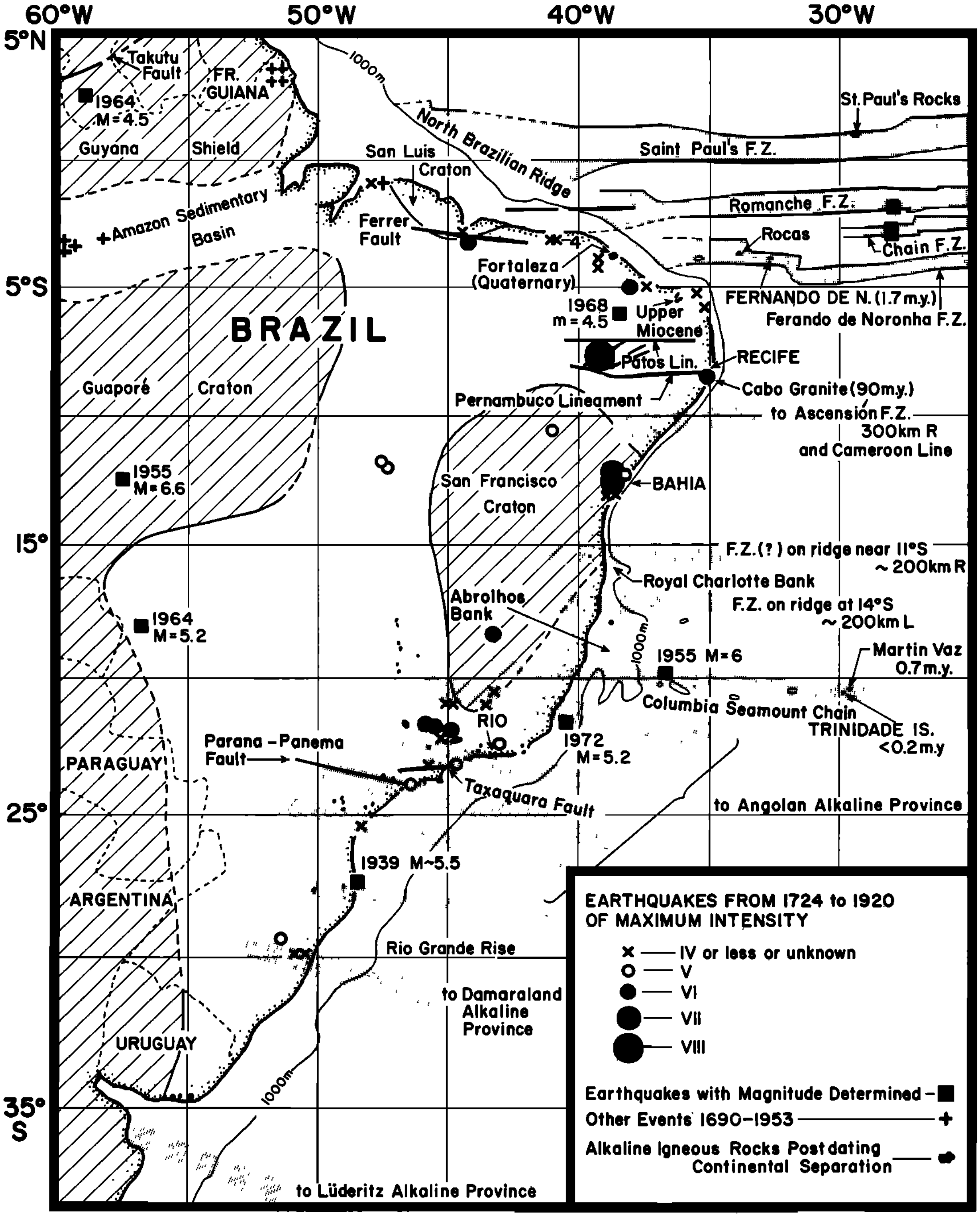

Fig. 9. Seismicity, igneous rocks postdating continental separation, and fracture zones near continental margin of Brazil. Earthquake locations are as follows: 1724–1920, felt reports of *Branner* [1912, 1920] with intensities in Rossi-Forel scale; felt reports 1690–1953, for the Amazon basin, *O'Reilly Sternberg* [1955]; instrumental locations with magnitude determined (squares), *Gutenberg and Richter* [1954]; *Rothé* [1969] and reports of U.S. Geological Survey through 1975. Some felt reports (not shown) for western Brazil may be from shocks located farther west in the Andean region. Fracture zones near continental margin (continuous shaded lines) are from *Francheteau and Le Pichon* [1972], *Gorini and Bryan* [1976], and *Kumar et al.* [1977]. Fracture zones inferred from offsets of mid-Atlantic ridge by appropriate finite rotations are indicated by dashed shaded lines. Young igneous rocks are from *Almeida* [1971], *Marsh* [1973], *Neill* [1973], *Asmus and Ponte* [1973], *Baker* [1973], and *Gorini and Bryan* [1976]. Tectonic features on land are from *de Loczy* [1970], *Gorini and Bryan* [1976], and *Kumar et al.* [1977]. Cratonic areas (hatched) older than 800 m.y. are from *Almeida et al.* [1973] and *Hurley and Rand* [1973].

great fracture zones with the preexisting mosaic of tectonic elements—cratons and younger orogenic belts—rather than with one or more underlying mantle plumes.

SOUTH-WEST AFRICA, ANGOLA, AND SOUTH AFRICA

Fracture Zones and Previous Zones of Weakness

One of the major offsets of the continental margin and of the magnetic anomaly pattern in the South Atlantic is found at the Agulhas fracture zone, which forms the southeastern continental margin of South Africa. The Falkland plateau separated from South Africa along this fracture zone, which has its counterpart, the Falkland fracture zone, along the continental margin of South America. The location of the Agulhas fracture zone may have been controlled by the preexisting Cape fold belt (Figure 5) of Paleozoic age [*Kennedy*, 1965; *A. O Fuller*, 1971].

The three transform directions that pass through the Damaraland and Lüderitz alkaline provinces and along the Agulhas fracture zone in Figure 5 were drawn through prominent offsets in the magnetic anomaly pattern of Mesozoic age. They have a similar location to three marginal offsets proposed by *Dingle and Scrutton* [1974] that appear to have influenced the areas of maximum sedimentation along the continental margin. *A. O. Fuller* [1971] proposes that several of the major fracture zones along the mid-Atlantic ridge south of 25°S have their counterpart within the African continent along old lines of weakness.

Late Precambrian (Pan-African) trends within the Nama fold belt along the west coast of South Africa and South-West Africa strike nearly parallel to the present continental margin [*Kennedy*, 1965; *Martin*, 1973]. Thus the continental margin in this area appears to have opened along a relatively young fold belt located to the west of the older rocks of the Kalahari craton [*Kennedy*, 1965]. The Damaraland alkaline province appears to have developed along the northeast trending Damara fold belt (Figure 5) which is also of Pan-African age [*Kennedy*, 1965]. The Damara belt is interpreted as a suture zone between the Congo and the Kalahari craton by *Burke et al.* [1977] and others. Whether that orogenic belt and others represent suture zones or zones of younger intracratonic deformation is not crucial, however, to the ideas of preexisting zones of weakness discussed here. Thus the locations of the continental margin, fracture zones, and centers of alkaline magmatism seem to be strongly controlled by preexisting zones of weakness. The Walvis ridge and the Rio Grande rise (Figures 5 and 9) are among the largest bathymetric features in the South Atlantic, yet they do not appear to be related to major offsets of the continental margins of Africa or Brazil, respectively. Both features are situated on oceanic crust that is older than about 80 m.y. They appear neither to continue onto younger sea floor nor to intersect one another on the mid-Atlantic ridge near Tristan da Cunha as some proponents of an origin related to moving hot spots have maintained. Major segments of the Walvis ridge and Rio Grande rise appear to be oriented along transform faults, while other portions strike nearly perpendicular to transform directions, i.e., they appear to be oriented parallel to spreading centers [*Francheteau and Le Pichon*, 1972]. The rise and ridge appear to join as nearly one continuous feature when the South Atlantic is closed to the 80-m.y. isochron [*Francheteau and Le Pichon*, 1972].

The various segments of the Walvis ridge, which collectively trend southwest from the coast of Africa, are situated seaward of the Damara fold belt (Figure 5). Thus the various segments of the Walvis ridge and Rio Grande rise appear to have developed at the former intersection of three mobile belts of Pan-African age that separated the Congo, Kalahari, and Guaporé cratons of Africa and South America. In a later section I suggest that the Walvis ridge and Rio Grande rise may have developed along leaky transform faults which ceased to be active after a major reorganization in spreading directions occurred about 80 m.y. ago.

Seismicity

Seismic activity in Africa south of 10°S from 1900 to 1973 (Figure 5) does not show a very clear correlation with oceanic fracture zones or transform directions. It is not clear whether the relatively high level of seismic activity in South Africa in Figure 5 is related to the East African rift system [*Vail*, 1967; *de Beer et al.*, 1975; *Scholz et al.*, 1976] or perhaps to zones of weakness that were reactivated during either the separation of Antarctica from Africa or of South America from Africa. The possible association of seismic activity with Cretaceous kimberlites, epeirogenic uplift of the southern part of the continent, and subsidence of the continental margins will be discussed later in the section on alkaline rocks and kimberlites.

Seismic activity associated with the western rift zone of East Africa is omitted in Figure 5 but is shown in Figure 10. Much of the other activity shown in Figure 5 to the west and southwest of the western rift is probably related to the continued development of the rift system rather than to the opening of the South Atlantic [*Scholz et al.*, 1976]. The rift system will be discussed in a later section. Insofar as possible, rock bursts related to mining in South Africa were eliminated in Figure 5.

Nevertheless, if we focus only on those parts of Africa close to the Atlantic continental margin, a possible correlation of seismicity, young rocks, and transform faults becomes more apparent for South-West Africa and Angola. The earthquake of magnitude 6 on May 24, 1914 (Figure 5), occurred within the Angolan alkaline province near the end of a major fracture zone and near a major offset of the continental margin of Angola [*Francheteau and Le Pichon*, 1972]. Two of the larger South African earthquakes, those of 1912 and 1920 (Figure 5), are located along the Agulhas fracture zone.

Felt Earthquakes

The historic record of felt earthquakes also suggests a higher level of activity in South-West Africa than does the map of instrumental locations for the period 1900–1973 (Figure 5). This is not surprising, since the detection level for instrumental locations has been poor for this area. *Krenkel* [1923] shows a region of moderately high activity (Figure 8) near the west coast of South-West Africa and South Africa. *Gorshkov* [1963] reproduces a map from *Sieberg* [1932] of the felt areas of seven moderate to large earthquakes in South-West Africa that occurred in 1850 and between 1906 and 1913. A shock in 1906 had a maximum felt intensity of VII–VIII. Several of these historic earthquakes appear to be associated with the Omatako graben [*Gorshkov*, 1963], the Damaraland alkaline province (Figure 5), and the Damara fold belt. *Korn and Martin*'s [1951] map of the felt areas of 223 earthquakes in South-West Africa from 1911 to 1938 shows a pronounced concentration near the Omatako graben. *Scholz et al.* [1976] also show a number of epicenters in this area as compiled from the seismological bulletins of the Rhodesian Meteorological Services.

Alkalic Rocks and Kimberlites

Marsh [1973] proposes that the igneous rocks of the Angolan, Damaraland, and Lüderitz alkaline provinces (Figure 5) form distinct northeast trending lineaments in Angola and South-West Africa. He found that these lineaments lie along small circles about the Cretaceous center of rotation for the opening of the South Atlantic. They appear to be correlated with transform faults that offset the mid-Atlantic ridge and with similar alkaline igneous complexes in Brazil and Uruguay (Figure 9). The three southernmost transform directions shown in Figure 5 by solid lines also intersect the continental margin of Africa, where *Rabinowitz* [1976] finds offsets in the magnetic pattern of Mesozoic age. The NNW trending pattern of linear magnetic anomalies [*Rabinowitz*, 1976] is disturbed for about 100–200 km along the continental margin offshore from the Damaraland alkaline province (Figure 5).

Marsh [1973] tabulates four radiometric ages from the Damaraland province of 123–136 m.y. and single measurements of 112 and 130 m.y. for the Angolan and Lüderitz provinces, respectively. The ages for the Damaraland province are similar to the 110- to 128-m.y. ages obtained by *Gidskehaug et al.* [1975] for the Kaoko basalts of this region, which are thought to be contemporaneous with the early opening of

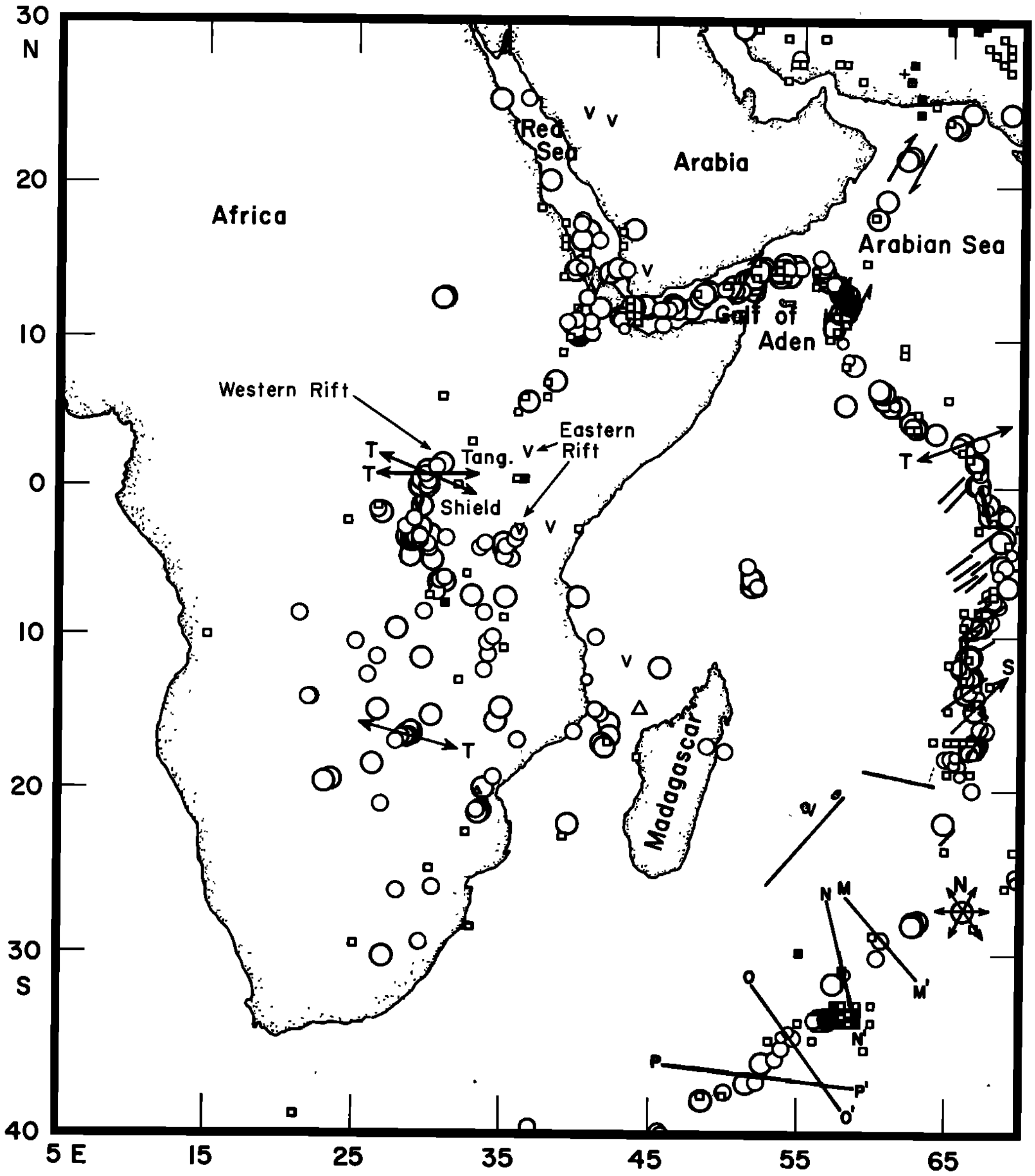

Fig. 10. Seismicity of East Africa and adjacent oceanic areas, after *Sykes* [1970*b*]. Circles denote relocated epicenters of shallow earthquakes from 1950 to 1966. Larger symbols denote more precise locations. Squares indicate locations for events before 1950. Solid symbols indicate events larger than magnitude 7, and V depicts historically active volcanoes from *Gutenberg and Richter* [1954]. T denotes tensional axis from focal mechanism solutions. Note that seismic activity along the mid-Indian-Ocean ridge is concentrated in a single narrow zone, whereas that in East Africa is generally multibranched and is spread over a large area. The multibranched nature of the East African rift system, particularly south of the equator, may be a model for the initial rupturing of continents.

the South Atlantic. Marsh lists a single age of 120 m.y. for the Uruguayan province of South America, which is opposite (conjugate to) the Lüderitz province, and 45–85 m.y. and 120–145 m.y. for alkaline complexes in Brazil, which are conjugate to the Angolan province. He states that the available isotopic age determinations for the three alkaline provinces in western Africa and for those in Brazil and Uruguay do not show any distinct spatial progression, which argues against an origin related to moving hot spots. *Kröner* [1973] reports radiometric ages on phonolites and melilite basalt plugs of 35.7–37 m.y. from the Lüderitz province, South-West Africa, and 38.5 m.y. from western South Africa.

The alkalic complexes of Angola form a well-defined northeasterly trend [*Neill,* 1973]. When this line is extrapolated to the northeast (Figure 5), it passes along the Lucapa graben of northeastern Angola, where kimberlites believed to be of Upper Jurassic age are found, and into the adjacent part of Zaire, where kimberlites of Mesozoic age occur [*Dawson,* 1970; *Neill,* 1973].

Two other transform directions are indicated by shaded dashed lines in Figure 5 near the west coast of South Africa. The southernmost of the two, which was identified by *Rabinowitz* [1976] from an offset in the pattern of Mesozoic magnetic anomalies, might have an expression inland in the occurrence of numerous kimberlites as well as seismic activity. The northernmost transform direction indicated by shaded dashed lines in Figure 5, which forms the southern boundary of the Orange basin on the continental margin [*Dingle and Scrutton,* 1974] and which matches a fracture zone near Tristan da Cunha at 41½°S on the mid-Atlantic ridge [*Ladd et al.*, 1973], passes near a number of kimberlite localities that trend northeast in western South Africa [*Cornelissen and Verwoerd*, 1975]. *Dingle and Gentle* [1972] also identified igneous rocks about 58 m.y. old on Agulhas bank off the south coast of South Africa. Thus a number of occurrences of rocks younger than the initial age (125–130 m.y.) of sea floor spreading in the South Atlantic are present in South-West Africa and western South Africa. Several of these occurrences appear to be located along zones extending inland from the ends of oceanic fracture zones. The trend of individual zones of kimberlites or of alkaline rocks is only well developed for the zone that intersects the coastline near 31°S. Its strike differs from that of the computed transform direction in Figure 5 by about 35°. *Kröner* [1973] and *Cornelissen and Verwoerd* [1975], however, believe that the younger alkaline rocks of South Africa and South-West Africa define a northwesterly trend associated with the incipient stages of continental separation. The latter authors remark, however, that near 30.2°S, 18.5°E, kimberlite pipe lineaments trend both northeast and northwest.

Emplacement of Kimberlites During Opening of the South Atlantic

Most kimberlites in South Africa and South-West Africa are of Cretaceous age [*Crockett and Mason*, 1968; *Dawson*, 1970; *Cornelissen and Verwoerd*, 1975]. *Allsopp and Barrett* [1975] report ages of 86 m.y. for five kimberlite pipes in South Africa and 127 and 147 m.y. for two other pipes. Thus the 86-m.y. dates are considerably younger than the 125- to 130-m.y. age of initial opening of the South Atlantic inferred from the oldest marine magnetic anomalies [*Larson and Ladd*, 1973].

Crockett and Mason [1968] observe that seismic activity in South Africa is concentrated in a distinct belt that closely approximates the distribution of lavas of the Karroo system as well as the younger kimberlite pipes. The broad zone of kimberlite occurrence in South Africa (Figure 5) is a region of relatively high seismic activity, which includes several shocks of magnitude 6 or greater. *Oliver*'s [1956] map of South African earthquakes from 1953 through 1955 shows a similar distribution.

Only one focal mechanism solution is known to me for the entire area of South Africa, South-West Africa, and Angola. A strike slip solution was obtained by *Fairhead and Girdler* [1971] for the Ceres earthquake near Capetown in 1969 (Figure 5). Of the two nodal planes the one striking N41°W appears to be the fault plane from the distribution of aftershocks. That direction suggests faulting subparallel to the coast rather than along a transform direction.

Dawson [1970] states that 'most kimberlites in South Africa are of Cretaceous age and are believed to have been intruded during a period of strong uplift of the continent with attendant downwarping or faulting around the periphery as a result of deepening of the contiguous ocean basins.' He concludes that kimberlites were intruded for the most part into major fundamental fractures that cut both the cratonic areas and the circumcratonic belts during periods of epeirogenic uplift. Some of these fractures were controlled by existing planes of weakness, but others transect the deepest known structures and apparently were independent of earlier structural weaknesses. These fundamental fractures probably extend into the uppermost mantle, perhaps to depths as great as 200 km beneath cratonic areas to account for the presence of diamonds and of pyroxene compositions of kimberlites [*Boyd and Nixon*, 1975].

It is not surprising that kimberlite pipes were emplaced in great numbers in South Africa starting at approximately the initial stage of opening of the South Atlantic and continuing for about 50 m.y. Probably a number of preexisting faults were reactivated as continental fragmentation occurred along both western and southeastern margins of South Africa and as the Falkland platform continued to move by South Africa for several tens of millions of years along the Agulhas fracture zone. It is understandable that faults near a 90° corner in the shape of the continental margin, as occurs off South Africa, might be extensively reactivated during two episodes of continental separation. This probably accounts for the widespread distribution of earthquakes and kimberlites in South Africa and South-West Africa. A mechanism related to the development of continental margins can explain the emplacement of kimberlites in the adjacent craton, where the source area of kimberlites is at a depth of about 100–200 km. Many workers on kimberlites have equated a cratonic setting with tectonic stability. This apparent enigma can be overcome if deep, reactivated faults control the emplacement of kimberlites.

Dawson [1970] points out that many of the kimberlites in Africa are concentrated within the older cratons. In East Africa, for example, kimberlites tend to be concentrated on the Tanganyika shield (Figures 5 and 10) between the western and the eastern branch of the East African rift system and to be rare along the branches of the rift system itself. In Siberia, kimberlites of Mesozoic age are also found in platform areas adjacent to belts of Mesozoic deformation. Likewise, kimberlites of Mesozoic age in West Africa are found within the West African craton, whereas other types of Mesozoic igneous rocks, which appear to originate at a shallower depth in the mantle, are concentrated along the younger belts of deformation surrounding that craton. Thus the time of emplacement of kimberlites in cratonic areas suggests that they are related to continental rifting and sea floor spreading near the margins of those cratons. Hence at least some deep fractures within cratons appear to be reactivated by continental fragmentation. Also, the type of magmatism appears to be related to the thickness of the lithosphere beneath various geologic provinces. Kimberlites are emplaced where the lithosphere is relatively thick, cold, and old.

EAST AFRICAN RIFT SYSTEM

A number of authors point out that the present rift system in East Africa, which has been in existence since the Tertiary, generally tends to follow preexisting zones of weakness along the margins of the old cratons. Some of the mobile belts along the margins of cratons in Africa appear to have been reactivated many times in the last 3 b.y. [*McConnell*, 1972]. *McConnell* [1972] calls some of the vertical shear zones of Africa perennial deep faults, since they have been reactivated many times. The western and eastern rifts and their associated earthquakes and volcanism (Figure 10) are located along the margins of the Tanganyika (Tanzanian) shield (Figures 5 and 10). The western rift [*McConnell*, 1972] follows the Ubendian orogenic belt of Precambrian age (~2 b.y.). Large parts of the eastern rift [*McConnell*, 1972] and the Lebombo monocline (Figure 5) of Mesozoic age [*Cox*, 1970] are located along the western side of the Mozambique belt (900–450 m.y.). Portions of the present rift system follow rift zones and centers of magmatism of Karroo (Mesozoic) age, which in turn represent a reactivation of yet older lines of weakness. Karroo deformation appears to be associated with the separation of Antarctica (and perhaps other fragments of Gondwanaland) from Africa. Evidently, a wide zone of rifting and igneous activity was present during the Mesozoic in East Africa as well as near the southeastern margins of Africa. The development of a throughgoing plate boundary near the present coast of southeastern Africa apparently stranded the Karroo rifts of East Africa within the interior of the African plate.

McConnell [1972] points out that Cenozoic and Holocene volcanic rocks are absent from the western rift except for very local occurrences that are associated with nodal points, i.e., junctions, of older mobile belts. Some of the present zones of abundant magmatism, i.e., hot spots, also appear to have been thermal nodes during the Precambrian and early Paleozoic. Hence the present hot spots do not appear to be related to active tectonic elements like mantle plumes but to preexisting nodes in the older tectonic fabric.

Seismic activity in southern Africa is high near the Lebombo monocline, which was the locus of abundant alkaline magmatism of Mesozoic (Karroo) age [*Cox*, 1970; *Woolley and Garson*, 1970]. Earthquakes are also numerous to the southwest of the East African rift system between Lake Kariba and the Congo craton (Figure 5). They are situated at the northeastern end of the northeast striking Damaran-Katangan orogenic (700–450 m.y.) belt [*McConnell*, 1972]. Portions of that belt were also reactivated during the Karroo period [*McConnell*, 1972]. Thus zones of seismic activity and Cenozoic magmatism in eastern and southern Africa tend to be associated with preexisting belts of deformation and tend to avoid passing through the older cratonic nuclei.

NORTHWEST AFRICA AND CANARY AND CAPE VERDE ISLANDS

Transform Directions and Hot Spots

As is described in a later section, small circles (transform directions) were drawn through the Newfoundland fracture zone, New England seamount chain, and Blake fracture zone using the center of rotation for the Mesozoic opening of the western Atlantic. These three features appear to be among the largest interruptions in the pattern of Mesozoic anomalies in the western Atlantic. Using the same center of rotation, transform directions in west Africa (Figure 5) were drawn passing through the New England seamount chain and Blake fracture zone. These two lines pass through the Canary and Cape Verde islands, respectively. These two conjugate points on the African side are sites of historic and Cenozoic volcanism [*Gutenberg and Richter*, 1954; *Abdel-Monem et al.*, 1971; *Burke and Wilson*, 1972; *Gunn and Watkins*, 1976] and are thought to be mantle hot spots by many workers. A conjugate point in Africa to the Newfoundland fracture zone (not shown in Figure 5) would lie in the vicinity of Gibraltar and northernmost Morocco, near the presently active plate boundary that follows the Azores-Gibraltar ridge.

Hayes and Rabinowitz [1975] mapped the Mesozoic M sequence of anomalies in considerable detail off the west coast of Africa between 34° and 11°N. In addition to numerous small offsets of tens of kilometers, which they interpret as fracture zones trending approximately east-west, they find that the magnetic pattern is offset about 100 km to the north of the Canary Islands between 29° and 30°N. They also show an offset of about 75 km across the Cape Verde Islands. Both of these offsets are most prominent for the oldest anomalies (M22 to M24) that they mapped. Across the Canary Islands themselves the oldest anomalies are not offset noticeably, nor do they change strike appreciably. Thus magnetic anomalies experience their greatest offset both at the Cape Verde Islands and about 100–200 km north of the Canary Islands. In Figure 5 a zone of shading about 200 km wide is drawn to include both the magmatism of the Canary Islands and the zone of offset magnetic anomalies north of the islands. In addition, the trend of magnetic anomalies changes abruptly from northeast to north off Cape Blanc at 23°N, which is conjugate to the Norfolk fracture zone of the western Atlantic and the central Virginia seismic zone.

Seismicity

Figure 5 shows a number of instrumentally recorded earthquakes in northwest Africa in the vicinity of the South Atlas fault zone. A very destructive earthquake of magnitude 5.9 occurred in 1960 at Agadir, Morocco, where this fault zone reaches the Atlantic Ocean [*Rothé*, 1969]. Although the relationship of the South Atlas fault zone to the tectonics of the Canary Islands or to the oceanic fracture zone between 29° and 30°N is not known in detail, these features appear to define a major discontinuity in the Mesozoic development of the eastern Atlantic. *Dewey et al.* [1973] conclude that a plate boundary passed along this zone between the African and the Moroccan plate during the early development of the Atlantic Ocean in the Mesozoic. Right lateral movement along the South Atlas fault zone is needed to close the western Atlantic between the New England seamount chain and the Newfoundland fracture zone. The map of magnetic anomalies of *Hayes and Rabinowitz* [1975] indicates that this Mesozoic movement may be reflected in the more than 100 km of right lateral offset of the older M sequence of anomalies just to the north of the Canary Islands. If this fracture zone is extrapolated to the coast, the Agadir earthquake of 1960 would fall close to it.

The South Atlas fault appears to have developed along a major suture zone of Pan-African age on the northern side of the West African craton [*Dillon and Sougy*, 1974; *Burke et al.*, 1977]. The western part of the South Atlas fault zone (the Tizi n'Test fault of *Mattauer et al.* [1972]), along which epicenters are shown in Figure 5, experienced large right lateral strike slip motion during the late Paleozoic [*Rod*, 1962] and was again reactivated in post-Triassic time [*Mattauer et al.*, 1972; *Le Pichon et al.*, 1977] during the opening of the North Atlantic. *Le Pichon et al.* [1977] mention that the fault zone has been traced offshore by geophysical methods to south of Grand Canary Island in the Canary Islands. They conclude, as is also evident in Figure 5, that the strike of the South Atlas (or Tizi n'Test) fault zone differs significantly from the direction of early opening of the North Atlantic. Thus transform faulting to the north of the Canary Islands and magmatism in the islands themselves appear to be related to the presence of a major preexisting zone of weakness in Morocco. Seismic activity along the reactivated Tizi n'Test fault zone in Morocco follows the trend of the late Paleozoic zone of weakness rather than the direction of Mesozoic opening of the North Atlantic. In central Morocco the earthquake belt leaves the South Atlas fault (Figure 5), follows the Middle Atlas mountains, and then joins the zone of Cenozoic deformation in the Rif area of northern Morocco.

Several earthquakes in Figure 5, including one of magnitude 6.2 in 1959, are located either in or to the west of the Canary Islands. They fall, however, about 100–200 km south of the major offset of the magnetic pattern that is located between 29° and 30°N.

Most of the intraplate earthquakes studied in this paper are located either within continents or along those parts of continental margins floored by continental crust. Intraplate shocks within the oceanic lithosphere are relatively rare. The two shocks of magnitude 6¼–6¾ in 1938 and 1941 that are located northwest of the Cape Verde Islands in Figure 5 occurred well within oceanic lithosphere. Although they fall close to one of the smaller fracture zones mapped by *Hayes and Rabinowitz* [1975], these shocks are not situated on the fracture zone shown in Figure 5 that passes through the Cape Verde Islands. Several other shocks in Figure 5 are located within oceanic lithosphere between Madeira and the coast of Morocco.

Krenkel's [1923] map of earthquakes in Africa (Figure 8), however, does indicate moderate historic seismic activity in the Cape Verde and Canary islands, in Morocco along the South Atlas fault zone, and to the east of the Cape Verde Islands, where young volcanism is found near Dakar. *Jacobi* [1976] describes several major submarine slides on the continental margin north of Dakar that he believes may have been triggered by earthquakes associated with the Cape Verde Islands, Cape Verde plateau, and adjacent fracture zones.

Guinea Fracture Zone

In Figure 5 the 1000-m depth contour is offset several hundred kilometers near 8°N at the Guinea fracture zone. One well-located earthquake near the coast of Sierra Leone [*Sykes and Landisman,* 1964] is found near the end of this fracture zone. Nevertheless, the known seismicity of this region as well as that of its conjugate point near the western end of the Bahamas fracture zone is extremely low. *Grantham and Allen* [1963] and *Hawkes* [1972] describe a swarm of kimberlite dikes striking ENE to northeast, some of which contain diamonds, in Sierra Leone near the end of the Guinea fracture zone (Figure 5). These dikes intrude and displace dolerite dikes of Triassic-Jurassic age and themselves are thought to be of Cretaceous age. *Lazarenkov* [1976] reports alkaline lamprophyres dated as 96 m.y. and ring fractures in the Los Alkaline province of Guinea, which is located about 100 km north of the Guinea fracture zone (Figure 5). The Freetown, Sierra Leone, basic igneous complex of Triassic age [*Briden et al.,* 1971] is also located near the end of the Guinea fracture zone (Figure 5). *Williams and Williams* [1977] conclude that a number of other occurrences of kimberlites of Mesozoic age in West Africa (Figure 5) are situated along the inferred eastward extension of oceanic fracture zones.

BRAZIL

Transform Directions

A number of great fracture zones in the equatorial Atlantic (Figures 4 and 9) have been traced into the continental margin of northeastern Brazil [*Gorini and Bryan,* 1976; *Kumar et al.,* 1976]. In this area the North Brazilian ridge consists of two east-west segments that can be traced into the St. Paul's and Romanche fracture zones and an intervening northwest trending segment (Figure 9) that may represent a foundered continental fragment.

The expressions of fracture zones near the coast of northeastern Brazil and also off Argentina and the Falkland plateau are more difficult to see in the bathymetry than are several of the fracture zones off the west coast of Africa, since the former are generally blanketed by thicker sediments. Nevertheless, as off Africa, fracture zones divide the continental margin of Brazil into segments along which sedimentary basins appear to have evolved independently [*Asmus and Ponte,* 1973; *Gorini and Bryan,* 1976].

Using the center of rotation for the Mesozoic opening of the South Atlantic of *Le Pichon and Hayes* [1971], the transform directions shown as solid shaded lines in Figure 9 were extended into South America between 19° and 30°S from offsets of the continental margin identified by *Francheteau and Le Pichon* [1972]. *Kumar et al.* [1977] mapped two of these fracture zones farther east off the continental margin. Since offsets of the continental margin are not obvious from 5° to 17°S and since fracture zones in the westernmost South Atlantic have not been detected from offsets of magnetic anomalies, it is difficult to identify fracture zones near the margins of eastern South America between 5° and 17°S and from 31° to 50°S. The four transform directions shown by dashed shaded lines in Figure 9 were extrapolated from offsets of the mid-Atlantic ridge by making appropriate finite rotations to close the South Atlantic. Obviously, much more work remains to be done in mapping fracture zones near the continental margin before much can be said about tectonism in the continent near the ends of oceanic transform faults.

Young Igneous Rocks and Tectonism Near the Continental Margin

Although they are not great in areal extent, a large number of igneous rocks postdating the separation of South America and Africa are found near the east coasts of Brazil and Uruguay and on several islands in the western South Atlantic (Figure 9). *Gorini and Bryan* [1976] find that several very marked geologic features occur along the inferred westward continuation of the St. Paul's, Romanche, Chain, and Fernando de Noronha (Charcot) fracture zones. The St. Paul's fracture zone appears to extend under the very thick sediments of the continental shelf near the mouth of the Amazon. *Gorini and Bryan* [1976] tentatively correlate a westward extension of the Romanche fracture zone with a flexure zone consisting of horsts and grabens beneath the continental shelf. The alkaline rocks of Mecejana near Fortaleza, which are believed to be of Quaternary age [*Almeida,* 1971], appear to lie near the end of either the Chain or Fernando de Noronha fracture zones [*Gorini and Bryan,* 1976].

Igneous rocks near Macau (5°S, 37°W) of upper Miocene age (Figure 9) occur along the south flank of the west trending Potiguar basin of northeastern Brazil [*Asmus and Ponte,* 1973]. These rocks, the southern margin of the basin, and the abrupt change in strike of the continental margin from easterly to southerly may be associated with the westward end of the Fernando de Noronha fracture zone (Figure 9). Igneous activity on Fernando de Noronha dated as 1.7–12 m.y. [*Baker,* 1973] is much younger than that calculated for nearby sea floor generated at the crest of the mid-Atlantic ridge. Rocos atoll [*Wilson,* 1963] appears to lie along the same fracture zone as Fernando de Noronha.

A complex of extrusive and intrusive rocks dated as 87–99 m.y. is found in the Cabo area near Recife [*Asmus and Ponte,* 1973]. In a reconstruction closing the South Atlantic, *Gorini and Bryan* [1976] find that the young rocks of the Cabo area, the Recife plateau, and the Pernambuco lineament (Figure 9) are situated approximately along the strike of the Cameroon line of Africa (Figure 5). They conclude that the Pernambuco lineament, a Precambrian fault zone that was reactivated in the Cretaceous, and the Ngaourandéré fault zone of Africa

(Figure 5) form a continuous lineament upon closing the South Atlantic. In Brazil these features, as well as two of the larger historic earthquakes (Figure 9), lie about 200 km north of a transform direction obtained by rotating the Ascensión fracture zone of the mid-Atlantic ridge to Brazil by the two finite rotations of *Le Pichon and Hayes* [1971] that close the South Atlantic. *Asmus and Ponte* [1973] indicate a saddle near Cabo between two of the major basins of the continental margin. This saddle, the change in the strike of the continental margin near Cabo, the young igneous activity, and the seismicity (Figure 9) suggest that the end of a major fracture zone occurs near Cabo and Recife rather than 200 km south as indicated by the dashed shaded line in Figure 9. The bathymetry (N. Kumar and M. Gorini, personal communication, 1977) also suggests that a major fracture zone exists off the continental margin to the east of Recife. Thus seismic activity and young igneous rocks appear to be associated with the eastern end of the reactivated Pernambuco lineament in the same way as they also are related to its counterpart in Africa, the Ngaourandéré fault.

Asmus and Ponte [1973] mention lavas encountered in a well near 18°S on Abrolhos bank (Figure 9) that penetrate rocks believed to be of Upper Cretaceous age. They conclude that igneous activity in the area may have extended into the Eocene as indicated by K-Ar ages of 42–52 m.y. [*Herz,* 1977] on intrusive rocks from Santa Barbara and nearby islands. *Asmus and Ponte* [1973] speculate that Royal Charlotte bank near 16°S also consists of volcanics, which are probably of Upper Cretaceous age. A major submarine ridge that includes the Columbia seamount chain, as well as rocks less than 0.7 m.y. old on the islands of Trinidade and Martin Vaz [*Baker,* 1973; *Herz,* 1977], probably is the site of a major fracture zone. Thus most of the examples of young alkaline magmatism to the north of 20°S in Brazil are found either on or near the continental margin near the ends of major oceanic transform faults.

Alkaline Rocks in Southern Brazil

A large number of young igneous bodies are found in southern Brazil (Figure 9) to the northwest, east, and southwest of Rio de Janeiro. This greater frequency of young rocks coincides generally with the higher density of fracture zones near the continental margin and with the higher seismicity of the area. These rocks fall into two age groups, of about 45–110 m.y. and 120–140 m.y. [*Amaral et al.,* 1967; *Marsh,* 1973; *Neill,* 1973; *Herz,* 1977], the older of which includes the period of more widespread volcanism associated with continental rifting. The Pocos de Caldos alkaline complex, a nearly circular feature about 30 km in diameter, is the largest structure that is formed by the younger series of igneous rocks [*Opdyke and MacDonald,* 1973]. *Herz* [1977] divides the younger alkalic rocks into three groups. He infers that these three groups, as well as the older alkalic rocks, are related to the development of a triple junction near Rio de Janeiro and to the movements of hot spots. One of the problems with this hypothesis is that it must explain alkalic magmatism that occurs in several different areas over a period of about 90 m.y.

Diamonds are found in placer deposits in several areas in southern Brazil. It is possible that the sources may be kimberlite pipes of Cretaceous age, since alkaline rocks of that age are found in the same general area as the diamonds [*Meyer and Svisero,* 1975].

Seismicity

Worldwide compilations of instrumental locations of earthquakes, such as the compilation of *Gutenberg and Richter* [1954], indicate very few shocks in South America to the east of the Andes. In large part this can be attributed to the relatively low activity of the region compared to that of most plate boundaries. Nevertheless, events as large as magnitude 5½–6 may well not have been catalogued by them, since the coverage of seismic stations in the area was poor for much of the twentieth century. Three events with instrumentally determined magnitudes of 5.2–6 occurred near the coast of southern Brazil from 1939 to 1976 (Figure 9). The longer history of felt reports of earthquakes in Brazil [*Branner,* 1912, 1920; *O'Reilly Sternberg,* 1955] indicates maximum intensities of at least VIII (Rossi-Forel), that is, magnitudes of about 6. While not high, the seismic activity is not negligible in designing important structures.

Most of the felt events in Figure 9 are located near the coast of Brazil. While population bias probably does influence this map of historic felt shocks, several instrumentally located events are nevertheless found in the same areas. Thus it is reasonable to conclude that the coastal areas of Brazil are generally more active than much of the interior.

Southern Brazil. A high concentration of earthquakes is found in Figure 9 in southern Brazil, where a large number of fracture zones are found near the continental margin. That area is also characterized by the occurrence of young alkaline rocks. All three of the instrumentally located earthquakes in southern Brazil in Figure 9 are situated near the coast very close to the ends of fracture zones. Of course, more detailed mapping of oceanic fracture zones must be done in this area before it can be concluded that this coincidence is not a result of mere chance. Nevertheless, it is reasonable that the density of fracture zones is high between 20° and 30°S, since the continental margin is far from being perpendicular to the inferred direction of Mesozoic opening of the South Atlantic. Likewise, the density of fracture zones and of young alkaline rocks is also high in Angola, conjugate to southern Brazil. Some of the historic seismicity to the west of Rio de Janeiro (Figure 9) appears to be located on or near the Parana-Panema and Taxaquara faults.

Northeastern Brazil. A large number of historic shocks are also found in northeastern Brazil (Figure 9), another area of abundant alkaline magmatism. As was discussed earlier, a series of great fracture zones intersects the continental margin in this region. The epicenters of some of the historic shocks, such as the one of intensity VII–IX? [*Branner,* 1912] in northeastern Brazil in 1824 (Figure 9), are, of course, not as well located as most of the instrumentally recorded events. The earthquake of 1824 occurred on or near one of several faults [*Almeida et al.,* 1973] connecting the Patos and Pernambuco lineaments (both of which appear to have been reactivated in the Cretaceous [*Gorini and Bryan,* 1976; *Sadowski,* 1978]).

The series of felt earthquakes near Bahia in Figure 9 is not known to be located near the end of a fracture zone. Nevertheless, the presence of one is suspected, since the continental margin changes strike near there.

Reactivation of Old Zones of Weakness

I have already mentioned that the Precambrian Pernambuco lineament of northeastern Brazil was reactivated during continental rifting and has been the locus of more recent magmatism and at least one historic earthquake (Figure 9). *Asmus and Ponte* [1973] show a number of major structural features in Brazil and Africa that appear to be continuous in a predrift reconstruction of the two continents. Fracture zones near 2°S, 21°S, and 30°S in Figure 9 fall close to three of the west trending lineaments that they describe.

Sadowski [1978] proposes that a number of late Precambrian and early Paleozoic tectonic features, which are oriented either subparallel to the present continental margin or subparallel to Mesozoic transform directions, were reactivated during the Cretaceous. He contrasts the type of Mesozoic magmatism with differences in the age and strength of various pre-Mesozoic parts of the continental crust. He deduces that the pattern of fracture zones in the South Atlantic is strongly governed by the distribution of older zones of weakness.

Offield et al. [1977] identified a previously unnoticed structural lineament on Landsat images that trends east-west near 31°S in southern Brazil. They believe that it is associated with the offset of the continental margin shown in Figure 9 near that latitude, offsets of the magnetic anomaly pattern on the shelf near 30.5°S, the sharp southern boundary of the Rio Grande rise, and the southern boundary of the Damara orogenic belt in Africa. As in Africa, the lineament in Brazil appears to be a major tectonic boundary separating rocks to the north of Pan-African age from those to the south that are older than 2000 m.y. They infer that mineral deposits are localized along that lineament in both Brazil and South-West Africa. They state that it is likely that the lineament is a fault zone that was formed in the Precambrian and along which only the effects of its latest adjustment are now observed. Two epicenters in Figure 9 are located along that lineament.

The Amazon basin, which contains thick sequences of Paleozoic to Cenozic sediments and igneous activity dated as 170–293 m.y., lies between the Brazilian and the Guyana shield [*Bigarella*, 1973]. Regardless of its tectonic interpretation, be it rift, suture zone, or intercratonic basin, it seems to represent a major zone of weakness of pre-Mesozoic age. The St. Paul's and Romanche fracture zones are located nearly along the strike of the Amazon basin. Several historic earthquakes are known within the basin (Figure 9).

McConnell [1969] describes a major fault zone along the Takutu rift in Guyana (near 4°N, 58°W in Figure 9), across which Precambrian terrains do not appear to be continuous. This zone also appears to have been reactivated in the Cretaceous or Cenozoic. He describes a similar feature, the El Pao fault zone, in Venezuela. These fault zones are situated inland from several major fracture zones in the western Atlantic (to the north of the area shown in Figure 9).

Thus many young igneous rocks and earthquakes appear to be located in Brazil near the western ends of several major oceanic transform faults. The continent itself appears to have been affected by renewed tectonism along old lines of weakness that appear to have governed the way in which the South Atlantic Ocean developed. As in Africa, these lines of weakness apparently controlled the locations of the major transform faults involved in the early opening of the South Atlantic. Alkalic magmatism appears to be especially concentrated on the continents near the ends of major transform faults. While some of the larger earthquakes in Brazil also appear to be located at such places, other shocks may be related to faults that strike subparallel to the continental margin. Most of the earthquakes, as well as the occurrences of young alkaline rocks in Figure 9, are confined to the younger mobile belts (white areas) surrounding the older cratons (hatched areas).

AUSTRALIA

The eastern, western, and southern margins of Australia developed by continental separation in the Mesozoic and Cenozoic. Hence it is reasonable to ask if earthquakes and Cenozoic igneous activity are in any way related to one or more of these episodes of continental fragmentation along the margins of Australia.

Seismicity and Possible Relationship to Oceanic Fracture Zones

Earthquakes in Australia from 1897 to 1972 (solid circles) and shocks in the southeastern Indian Ocean from 1950 to 1966 (open circles) are shown in Figure 11. A series of great fracture zones (solid lines) intersect the mid-Indian-Ocean ridge south of Australia and Tasmania between 138° and 160°E. Seismic activity along these transform faults is very high between actively spreading ridge segments but is nearly absent in Figure 11 along the northern or southern prolongations of these fracture zones toward Australia and Antarctica. No earthquakes were found along the nearly straight east-west portion of the ridge (zone C) between 127° and 138°E. This segment of the ridge, like much of the East Pacific rise, is also characterized by fast spreading, smooth topography, and an apparent absence of major fracture zones [*Weissel and Hayes*, 1971, 1974]. The more seismically active segment of the ridge (zone B) between 120° and 127°E is characterized by rough topography, a greater density of small fracture zones (not shown in Figure 11), and asymmetrical sea floor spreading from 10 to 4.7 m.y. ago.

Although seismic activity in Australia is low in comparison with that of most plate margins, several shocks of magnitude greater than 6 have occurred during the twentieth century within and near Australia. All well-recorded earthquakes in Australia are located in the crust, and most are in the upper part of the crust [*Doyle et al.*, 1968]. Several major zones of earthquakes can be seen in Figure 11: (1) along the east coast and in Bass Strait and Tasmania, (2) surrounding the Yilgarn and Pilbara blocks of the western Australian shield, and (3) in a zone extending from Adelaide along 138°E into the center of Australia.

Cleary and Simpson [1971] speculate that the seismic zone along 138°E also extends from central Australia to the northwest coast of Australia. They propose that the three main seismic zones of Australia and their inferred extensions southward to the mid-Indian-Ocean ridge divide the Australian plate into four subplates. They believe that these are characterized by different spreading rates at the ridge crest. As *Doyle* [1971] and others point out, however, observed seismic activity (Figure 11) is nearly absent between the ridge and the south coast of Australia. Also, magnetic anomalies indicate that various segments of the ridge crest have opened without significant internal deformation of the surrounding lithospheric plates [*Weissel and Hayes*, 1971, 1974]. Thus significant differential movement does not appear to be occurring between the ridge and southern Australia; the continental margin marks the seaward extent of earthquake activity. Seismic activity in the western two thirds of the continent, however, does surround several of the older cratonic nuclei even though these zones do not extend outside the continent to join surrounding plate margins.

Adelaide seismic zone. A north trending zone of seismic activity along 138°E is located close to the boundary between the western Australian sheild (i.e., the eastern edge of the Gawler block in Figure 11) and a series of Paleozoic orogenic belts to the east. The 138°E zone is situated along the continuation of the strike of one or more major fracture zones in the southeast Indian Ocean between 140° and 145°E. The entire series of great fracture zones on the ridge between 140° and 155°E appears to have developed to the south of the series of Paleozoic orogenic belts in the eastern third of the country.

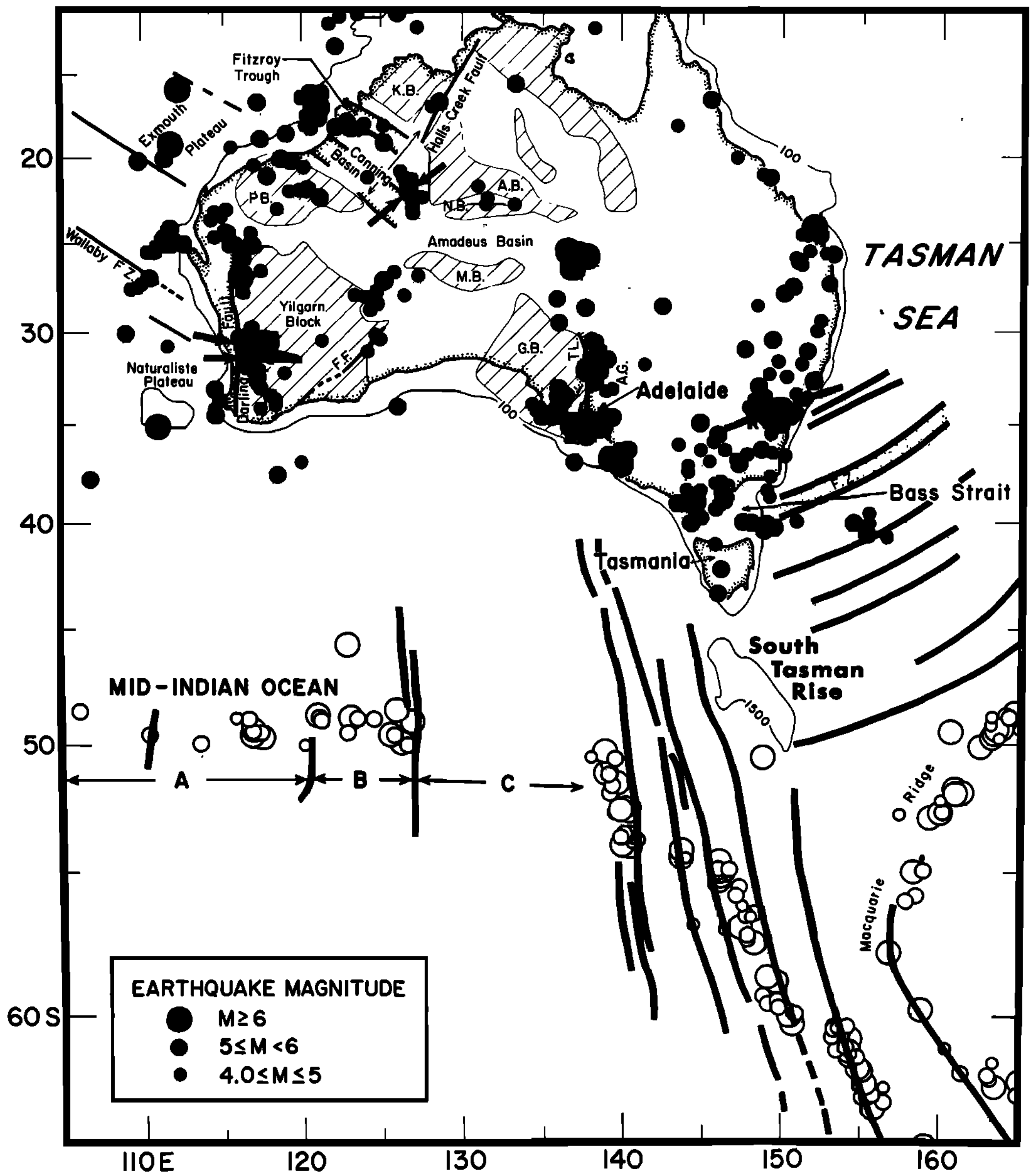

Fig. 11. Earthquakes in Australia (solid circles) from 1897 to 1972 from *Doyle et al.* [1968] and *Lilley* [1975] and the eastern Indian Ocean from 1950 to 1966 (open circles) from *Sykes* [1970*b*]. Seismic activity in the Adelaide region along 138°E and in southeastern Australia and Tasmania is located landward of several great fracture zones (solid lines) that offset the mid-Indian-Ocean ridge south of Australia. Note that fracture zones tend to develop seaward of major tectonic boundaries in continents that predate formation of surrounding oceans. Segments of ridge denoted A, B, and C and fracture zones (solid lines) south of Australia are from *Weissel and Hayes* [1971, 1974]. Fracture zones in the Tasman Sea are from *Weissel et al.* [1977]; those off western Australia are from *Larson* [1977, and personal communication, 1978]. Horizontal arrows denote direction of maximum compressive stress inferred from well-determined focal mechanism solutions of earthquakes [*Fitch et al.*, 1973; *Mills and Fitch*, 1977]. Kimberlite occurrences of possible Quaternary age indicated by K are from *Ferguson et al.* [1977]. Old cratonic areas experiencing their last major deformation prior to 1 b.y. ago, indicated by hatching, are from the *Geological Society of Australia* [1971], as are the faults. KB is Kimberley block, AB Arunta block, MB Musgrave block, PB Pilbara block, GB Gawler block, AG Adelaide geosyncline, FF Frasier fault; TL Torrens lineament, and NB Ngalia basin.

The 138°E zone nearly coincides with a belt of late Precambrian to early Paleozoic deformation in the Adelaide geosyncline, which is interpreted by some authors as an aulacogen [*Burke et al.*, 1977], and with a prominent anomaly in electrical conductivity along the Flinders range [*Lilley*, 1975]. Some of the activity also extends into the southeastern part of the Gawler block. The Paleozoic belts to the east of 138°E are characterized by higher heat flow, a better developed low-velocity zone in the upper mantle, and positive travel time residuals (slow velocities) compared to the shield [*Cleary et al.*, 1972]. It is likely that this major preexisting structural boundary influenced the latest pattern of continental fragmentation, the shape of the continental margin, and the location of major fracture zones near the continental margin [*Coward*, 1976].

Thus this ancient zone of weakness within Australia appears to have been reactivated in the Tertiary and to be the locus of seismic activity today. The presence of kimberlites and carbonatites [*Colchester,* 1972; *Tucker and Collerson,* 1972; *Stracke et al.*, 1977] indicates that this ancient zone of weakness probably extended through the entire lithosphere.

Within the Adelaide seismic zone, Pleistocene and Recent movements have occurred along the Mt. Lofty and Flinders horsts, which developed in the late Tertiary, and along the downdropped blocks of St. Vincent and Spencer gulfs [*David,* 1950; *Campana,* 1957]. David mentions a north striking fault in Adelaide that displaces deltaic sediments of late Cenozoic age. He states that the 1932 Mornington earthquake occurred on the Selwyns fault. Thus tectonic movements that started in the Adelaide area during the Tertiary and that appear to have continued into the Quaternary are expressed in the physiography, pattern of faults, and occurrence of earthquakes.

The absence of epicenters in Figure 11 along the south coast of Australia between 126° and 135°E matches a similar lack of activity along zone C of the mid-Indian-Ocean ridge between 127° and 138°E. I conclude that preexisting zones of weakness of northerly trend are not abundant along that segment of the continental margin of Australia; hence seismic activity is very low, and major transform faults did not develop. The regions of higher seismic activity in southwestern Australia and in adjacent offshore areas are situated north of the zones of higher activity, zones A and B, along the mid-Indian-Ocean ridge, where several fracture zones have been mapped (Figure 11). The presence of the Darling fault and other zones of weakness may have resulted in the development of a more complex pattern of transform faults in the adjacent ocean than in segment C.

Eastern Australia. The zone of high seismic activity near the east coast of Australia is about 350 km wide. Orogenic belts and old lines of weakness in eastern Australia generally strike north to NNW [*Geological Society of Australia,* 1971]. Sea floor spreading in the Tasman Sea, which occurred between 80 and 60 m.y. ago [*Hayes and Ringis,* 1973] also took place along NNW trending rift zones with fracture zones oriented ENE (Figure 11). Nevertheless, the overall strike of the continental margin of eastern Australia differs from that of spreading centers in the Tasman Sea by about 45°–50° [*Ringis,* 1975]. More northerly spreading centers in the Tasman Sea are offset progressively to the east in an en echelon pattern. Ringis concludes that many major structural features in southeastern Australia intersect the coast at or near the ends of major fracture zones in the Tasman Sea. Several discordant igneous intrusive masses on the coast or shelf that predate the opening of the Tasman Sea are situated along fracture zone trends. Hence Ringis deduces that the locations of fracture zones were governed by old zones of weakness in the continent.

With the exception of a group of earthquakes near 40°S, 155°E, seismic activity cuts off very abruptly in Figure 11 near the steep continental slope of eastern Australia. This may be contrasted with the west coast, where activity extends well off the coast into a region of complex plateaus and basins.

Minor volcanism started in eastern Australia about 70 m.y. ago during the opening of the Tasman Sea [*Wellman and McDougall,* 1974*b*]. From 55 to 10 m.y. ago, volcanism was nearly continuous in New South Wales, and its onset coincides closely in time with the initiation of sea floor spreading between Australia and Antarctica. More recent lavas of Pliocene-Pleistocene age occur over extensive areas of western Victoria and northeastern Queensland [*Wellman and McDougall,* 1974*a*, *b*]. *Wellman and McDougall* [1974*a*] find that central volcano provinces (but not lava field provinces) in eastern Australia show an apparent southerly migration of 6.6 cm/yr with decreasing age, which they attribute to the northerly movement of Australia with respect to an asthenospheric source.

The area affected by earthquakes in eastern Australia has also been uplifted in the Mesozoic and Cenozoic, the most recent being the post-middle-Miocene Kosciusko uplift [*Wellman and McDougall,* 1974*b*]. *David* [1950] states that faulting along the east coast of Australia or downwarping, which downdropped the platform on which the Great Barrier reefs were built, must have been active during the Pleistocene and even during Recent time. In a study of the northeast striking Tawonga fault in Victoria, *Beavis* [1960] concludes that Quaternary gravels were involved in a more recent episode of thrust faulting. The fault also experienced deformation in the early Paleozoic.

Apparently, tensional forces and the reactivation of old lines of weakness related to rifting between Australia and Antarctica and to the development of the Tasman Sea allowed magmas to reach the surface in eastern Australia. Although some volcanic features exhibit a space-time migration, it is difficult to explain the persistence of magmatism from 70 to 10 m.y. solely by the movement of a mantle hot spot. The cessation of volcanism in New South Wales about 10–15 m.y. ago may relate to dominantly compressional forces occurring in the lithosphere during the post-middle-Miocene Kosciusko uplift [*Wellman and McDougall,* 1974*a*, *b*]. Evidence for contemporary compression also comes from several focal mechanism solutions, which are discussed later. As in eastern North America, the stress field in eastern Australia appears to have changed with time.

Several kimberlite occurrences of Permian to late Cenozoic age have been reported for southeastern Australia and Tasmania [*Stracke et al.*, 1977; *Ferguson et al.*, 1977]. These and similar occurrences near Adelaide are situated in areas of seismic activity. They may be indicative of deep faults that cut through the lithosphere.

High seismic activity in the Bass Strait coincides spatially with the prominent Otway anomaly in electrical conductivity [*Bennett and Lilley,* 1974] and with several faults shown on the tectonic map of Australia and New Guinea [Geological Society of Australia, 1971]. The Gippsland and Otway basins in the Bass Strait, which contain thick sequences of Cretaceous and Cenozoic sediments, appear to have formed in response to the opening of the Tasman Sea and southeast Indian Ocean, respectively; they are interpreted by *Burke and Dewey* [1973] as a failed arm of a triple junction. Seismic activity in the eastern part of the Bass Strait may be located near the end of one of the major fracture zones in the Tasman Sea (Figure 11). *David* [1950] mentions that Recent faulting may have occurred along several of the coasts of Tasmania.

Western Australia. The western two thirds of Australia consists of a series of cratonic nuclei, most of which have been little deformed during the past 1 b.y. In Figure 11, hatching denotes known occurrences of rocks older than 1 b.y. as well as regions that are not believed to have experienced significant deformation in the last 1 b.y. Some of these blocks are known to be separated by younger zones of deformation, i.e., the Amadeus basin [*Geological Society of Australia,* 1971]. Some of them probably are more extensive than their surface exposure would suggest. Hence younger belts of deformation may not be present between all of the blocks of Figure 11.

Seismic activity in Figure 11 nearly surrounds the large Yilgarn block. Although earthquakes are found within that block, most of them occur near its margins.

The continental margin of western Australia developed during the Mesozoic [*Veevers and Cotterill*, 1976; *Larson*, 1977]. Cretaceous centers of sea floor spreading trend about N30°E, and associated transform faults strike about N60°W [*Larson*, 1977]. Thus the Cretaceous direction of sea floor opening was oblique to the Darling fault, one of the great preexisting fault zones of Australia. A band of earthquakes follows the northerly trend of the Darling fault and the trend of a prominent belt of gravity anomalies associated with the fault zone [*Australian Department of Natural Resources*, 1977]. Nevertheless, the earthquakes are not situated exactly on the Darling fault. Instead, they occur both within the Perth and Carnarvon basins, which are of Permian to Mesozoic age, and to the east of the Darling fault in the westernmost part of the western Australian shield. The Meckering earthquake of magnitude 7 in 1968 produced an unusually definite thrust fault scarp striking north. It occurred within the westernmost part of the shield [*Everingham et al.*, 1969].

Unlike most continental margins, seismic activity extends well off the coast of western Australia into a region of complex plateaus and basins which resemble the borderland off southern California [*Larson*, 1977]. The oblique opening of the ocean off western Australia with respect to the Darling fault may be reflected in the complex nature of the continental margin and in the broad region of seismic activity. One of the largest fracture zones to the west of Australia is situated along the southern margin of the Exmouth plateau. The distribution of magnetic anomalies indicates that one or more major fracture zones (dashed lines) are present seaward of the Canning basin and north of the Exmouth plateau. When the trends of fracture zones are projected into western Australia, they fall along either side of the Pilbara block within belts that were deformed within the last billion years.

One of the most active seismic zones in Australia is associated with the Fitzroy trough (Figure 11) in the Canning basin. The seismic zone, as well as faults mapped in the trough, strikes northwest, that is, nearly along the continuation of the strike of inferred fracture zones on the north side of the Exmouth plateau. A series of lamproites of Tertiary age occur along the trough [*Geological Society of Australia*, 1971]. The trough, which is located between the Kimberley and Pilbara cratonic blocks (Figure 11), experienced deformation along its northeastern side in late Precambrian–Early Cambrian time [*Geological Society of Australia*, 1971]. Prominent gravity anomalies striking northwest are found along the two sides of the trough as well as along the southwestern side of the Canning basin [*Australian Department of Natural Resources*, 1977]. Thus the present state of stress in the Canning basin may result from its being compressed between older cratonic blocks.

Also, mechanism solutions involving thrust faulting were obtained for two earthquakes that occurred just to the east of the Darling fault (Figure 11). That sense of movement differs from normal faulting that took place along the Darling fault during the Mesozoic. A number of normal faults (not shown in Figure 11) are indicated on the tectonic map of Australia and New Guinea [*Geological Society of Australia*, 1971] within the Pilbara block. The occurrence of earthquakes within that small block suggests that some of those faults may still be active.

Three shocks in the central part of Australia near 23°S, 131°–134°E (Figure 11) are situated near the boundary between the Arunta cratonic block and the Ngalia basin. Those events may be part of a zone of activity that extends from the eastern end of the Canning basin to the northern end of the Adelaide seismic zone. Three very prominent west striking gravity anomalies that are shown on the gravity map of Australia [*Australian Department of Natural Resources*, 1977] are associated with the Amadeus and Ngalia basins and with the cratonic blocks on their northern and southern sides. *Burke et al.* [1977] conclude that continent rifting occurred in the Amadeus basin about 1400 m.y. ago and that a suture zone of late Precambrian age is located along the northern side of the Musgrave block. Deformation of middle Paleozoic age also occurred in the Amadeus and Ngalia basins [*Geological Society of Australia*, 1971]. Hence seismic activity within Australia appears to follow mobile belts of late Precambrian to mid-Paleozoic age from Adelaide to the Amadeus and Ngalia basins and thence to the Canning basin.

Another zone of seismic activity in Figure 11 extends southwest from the western end of the Musgrave block. The seismic zone is nearly coincident with a prominent belt of gravity anomalies on the gravity map of Australia [*Australian Department of Natural Resources*, 1977]. Seismic activity follows a portion of the Frasier fault along the southeastern side of the Yilgarn block and then steps en echelon to the northwest at a similar step in the configuration of that cratonic block.

Thus old faults within mobile belts that are located along the edges of ancient cratonic nuclei appear to have been reactivated during the opening of oceans along the margins of Australia in the Mesozoic and early Cenozoic. Some of the seismic activity extends into the margins of the older blocks. Several of the most prominent gravity anomalies in Australia appear to follow old mobile belts and to be nearly coincident with seismic zones. Thus a more detailed analysis of the gravity field of Australia and of other continental areas may be of great value in identifying zones of weakness of Precambrian and Paleozoic age. Earthquakes appear to be occurring along some of those zones today in response to the present stress field.

Focal Mechanisms and State of Stress

Several focal mechanism solutions are available for Australian earthquakes (Figure 11). All of the better solutions known to me involve thrust faulting. *Fitch et al.* [1973] speculate that the near cessation of the subduction process north of Australia has led to high horizontal compressive stresses within Australia. Nevertheless, neither this mechanism nor one involving stresses related to the northward movement of the Australian-Indian plate explains the eastward directed maximum compressive stress inferred for three of the four focal mechanisms in Figure 11. Since the two stress directions for western Australia are almost perpendicular to the nearby continental margin and the stress in eastern Australia is also approximately perpendicular to its nearby margin, these stresses may be related in some way to processes beneath those parts of the continent that are adjacent to continental margins.

Stewart and Mount [1972] obtained several focal mechanism solutions for earthquakes along the Adelaide seismic zone, which they interpret in terms of contemporary movements. Their solutions seem to be questionable, since they are based on amplitudes of seismic waves at only a few stations. Since only a few well-constrained mechanism solutions are available for Australian earthquakes, additional solutions can be ex-

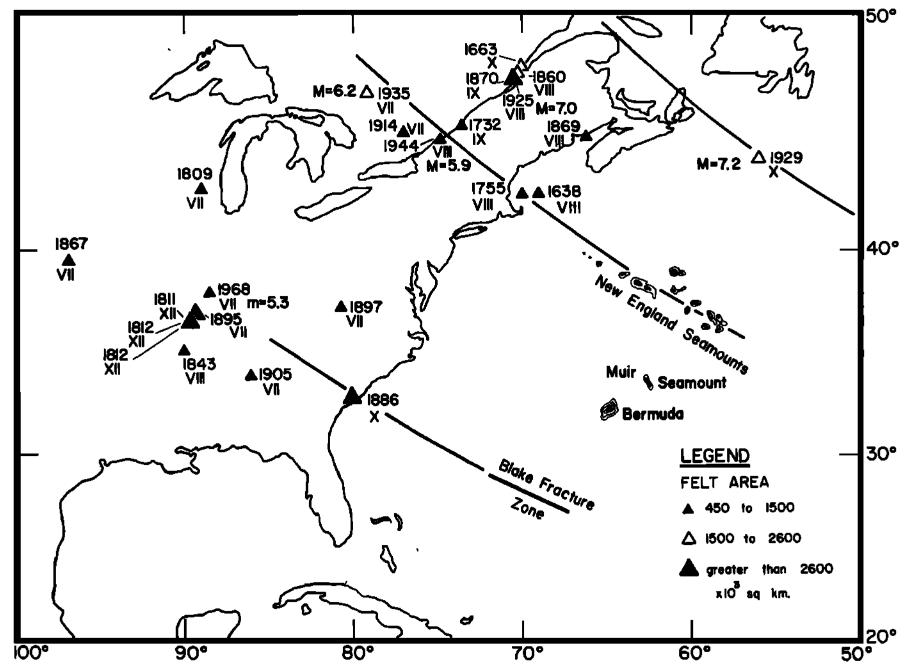

Fig. 12. Historic earthquakes in eastern North America [after *Fletcher et al.*, 1978]. Only earthquakes with felt areas greater than 450,000 km² and intensities of VII or greater are shown. Three small circles (solid lines) were drawn about a single center of rotation for the early opening of the Atlantic Ocean. These three small circles (transform directions) were started at the three major offsets (transform faults) in the magnetic lineation pattern of Mesozoic age in the western Atlantic and extended into eastern North America. Note that several of the larger historic shocks fall close to the small circle passing along the New England (Kelvin) seamount chain.

pected to increase considerably the understanding of current intraplate processes in that continent.

EASTERN AND CENTRAL NORTH AMERICA

Historic Earthquakes and Their Size

Although the rate of occurrence of earthquakes in the central and eastern United States and in eastern Canada is not as high as that of California or Alaska, several large and damaging earthquakes have nevertheless occurred there. Figure 12 shows known historic earthquakes in eastern and central North America of maximum intensity VII or greater that also were felt over an area of 450,000 km² or greater. The choice of only very large felt areas in Figure 12 was an attempt to select only the most energetic shocks. Not included are a few earthquakes for which maximum intensities of VII or greater were reported but for which the felt areas were small. These cases of strong shaking and smaller felt areas are probably related to very shallow focal depths. Some of the largest and most notable shocks in Figure 12 include the New Madrid earthquakes of 1811–1812 in southeastern Missouri and adjacent areas; the Charleston, South Carolina, earthquake of 1886; two earthquakes of intensity VIII or greater off the coast of eastern Massachussetts in 1638 and 1755; a number of large and destructive earthquakes in the St. Lawrence Valley; and the Grand Banks earthquake of 1929. A. E. Stevens (P. Basham, written communication, 1976) reanalyzed the felt reports of the shocks of 1869 and 1914 in the Bay of Fundy and near Ottawa, respectively, and concluded that the maximum intensities of those events were VI and V, respectively. If those lower intensities are adopted, those two shocks should be removed from Figure 12.

On the basis of the very large felt areas, *Richter* [1958] states that the New Madrid series of earthquakes in the central United States in 1811 and 1812 may have been some of the largest earthquakes in the history of the United States. *Nuttli* [1973*a*, 1974] attempted to calibrate the energy and magnitudes of some of the larger historic earthquakes in eastern and central North America by studying the magnitudes determined at teleseismic distances for waves of about 1-s period for recent earthquakes of moderate size. He finds that earthquakes in these areas are characterized by felt areas about 10–100 times as great as those of shocks of the same magnitude in western North America. Both he and *Evernden* [1975] conclude that the attenuation of short-period (about 0.3–3 s) surface waves (the main contribution to perception at large distances), is much greater for the western than for the central and eastern United States. Thus the New Madrid and the Charleston earthquakes liberated a smaller amount of energy for periods near 1 s than the San Francisco earthquake of 1906.

O. W. Nuttli (written communication, 1978) believes that a consideration only of magnitudes (m_b) determined from 1-s waves tends to downplay the earthquake hazard for the central and eastern United States. The spectral scaling of seismic waves is such that no earthquake will have an m_b greater than about 7. Nevertheless, events of m_b greater than 7 can have surface wave magnitudes as great as 8.5, which corresponds to the greatest of earthquakes. Thus long-period spectral estimations, including seismic moment, are not available for some of the largest intraplate shocks like the New Madrid events of 1811–1812. Nevertheless, events of clearly intraplate character are not known to have exceeded surface wave magnitudes of 7.8 anywhere in the world during the period 1904–1952 [*Gutenberg and Richter,* 1954].

Using a relationship between felt area and body wave magnitude m_b derived for earthquakes in central and eastern North America, *Nuttli* [1973*a*, 1976] obtained magnitudes for the following shocks from intensity data: 7.1–7.4 for the three largest events at New Madrid, 1811–1812; 6.5 for Charleston, 1886; 5.2 for Attica, New York, 1929; 5.5 for Anna, Ohio, 1937; and 5.5 for Ossipee, New Hampshire, 1940. Nuttli implicitly assumes, however, that teleseismically determined magnitudes m_b can be compared for source regions as different as western and central North America. As *Evernden* [1975, 1976] argues, regional biases in m_b probably exist, and calibration is necessary if the magnitudes or calculated energies of events from different tectonic provinces are to be compared in a meaningful way. To correct for regional bias in m_b and to permit a comparison with events in the western United States, the magnitudes given here for shocks in central and eastern North America probably should be reduced a few tenths of a magnitude unit. Nevertheless, this is only an approximate regional correction; it may vary with locality and tectonic setting.

Magnitudes determined at teleseismic distances are available for several earthquakes in eastern North America including the following: 7.0 for St. Lawrence Valley, 1925; 7.2 for Grand Banks (off eastern Canada), 1929; 6.2 for Timiskaming, Ontario, 1935; 5.9 for Massena, New York–Cornwall, Ontario, 1944; and 5.5 for south central Illinois, 1968. Thus the sizes of these shocks are by no means negligible, even when regional bias in teleseismic magnitudes is allowed for. The importance of these events in terms of seismic risk is underscored by their very large felt areas.

Seismic Trends

Many workers have attempted to recognize patterns or trends in the distribution of historically or instrumentally located earthquakes in eastern and central North America. *Woollard* [1958], *Hodgson* [1965], and others postulate a southwest trending seismic zone extending from the St. Lawrence Valley to southeastern Missouri. This trend is certainly suggestive to the eye in Figure 1 and in other similar maps of damaging earthquakes in the United States by the U.S. Coast and Geodetic Survey. One of the shocks in Figure 1 that is crucial in suggesting a southwesterly trend is the earthquake of 1638, which is shown at 47°N, 72½°W. *Smith* [1962] located the event in the St. Lawrence Valley, but in a subsequent publication [*Smith,* 1966], he relocated it off Cape Ann, Massachusetts. Evidence from recent instrumental recording in New York State indicates that seismicity near the eastern end of Lake Ontario is not continuous with that in westernmost New York State near Attica [*Sbar and Sykes,* 1977]. Likewise, microearthquake observations indicate that seismic activity northeast of Quebec City appears to be concentrated within a large meteorite crater. Thus activity does not appear to be continuous along the St. Lawrence River between Montreal and Quebec City [*Le Blanc et al.*, 1973; *Stevens,* 1974]. Although the area of seismic activity in western New York does not appear to join other seismic regions to the northeast along the St. Lawrence Valley, at the present time, instrumental data are not sufficient to ascertain if it joins or is separate from activity in west central Ohio and along the Wabash River in southern Illinois and Indiana.

In his map of historic earthquakes, *Woollard* [1965] suggests that a major zone of activity follows the Appalachian Mountain system from Alabama to eastern Canada. This belt of activity can be seen in Figures 13 and 14, which also show historic earthquakes. Since maps of historic earthquakes are based mainly on felt reports, they probably have some bias related to the population density. In Figures 13 and 14 the concentration of felt reports near the coast of New England and the mid-Atlantic states may be related in part to the high population density. Nevertheless, it is clear that several large earthquakes have occurred near the coasts of New England and the mid-Atlantic states, including the two large events off Cape Ann, Massachusetts, in 1638 and 1755 (Figure 12) and earthquakes in the vicinity of New York City of intensity VII in 1737, 1884, and 1927. Thus the Appalachian seismic belt still remains when only events of intensity VII or larger are plotted.

In a study of historic earthquakes, *Bollinger* [1972, 1973] identified four seismic zones in the southeastern United States (Figure 15). One follows the trend of the southern Appalachians from southwestern Virginia to central Alabama. He also defines a South Carolina–Georgia zone of northwesterly trend and a central Virginia zone trending west. *Richter* [1959] and others suggest that a northwesterly zone of high activity may extend across Tennessee from South Carolina to the New Madrid region of southeastern Missouri. Nevertheless, the record of historic activity and also more recent instrumental locations indicate that the South Carolina–Georgia zone does not extend farther west than eastern Tennessee.

Boston-Ottawa Seismic Zone

Diment et al. [1972] and *Sbar and Sykes* [1973] propose that a seismic zone trending across New England in a northwesterly direction is located along a continental extension of a line drawn along the New England seamount chain. Within that trend they include a region of higher activity off the coast of Massachusetts containing the large shocks of 1638 and 1755, the intensity VII Ossipee earthquake of 1940 in central New Hampshire, and a region of higher activity extending from northern New York State to Ottawa and thence to Kirkland Lake, Ontario (Figure 13). Six of the large earthquakes in Figure 12 are located along that zone. A northwest trending zone of activity is also apparent in the maps of historic earthquakes in Figures 13 and 14 and in the map of instrumentally located earthquakes from 1961 to 1974 (Figure 16). Although the instrumentally located earthquakes in Figure 16 suggest a nearly continuous zone of activity extending from offshore Massachusetts into Canada, the historic data shown in Figures 13 and 14 indicate a region of lower activity in Vermont and western New Hampshire.

Sbar and Sykes [1977] describe a network of about 30 short-period seismograph stations that has been in operation since 1970 in the states of New York, New Jersey, and Vermont. Additional stations have also been installed in New England since 1976 as part of a northeast U.S. seismic network. Epicentral locations of events in that area from 1970 to 1977 are shown in Figure 17. Detectability of events is still better for New York State, northern New Jersey, and Connecticut than it is for most other parts of New England. Thus Figure 17 probably contains some bias resulting from different detection capabilities and the fact that the stations in New York were installed earlier than most of those in New England.

A region of higher activity in northern New York State (Figure 17), which extends to the northwest into Canada, is well defined by the recent epicenters and is located along what *Sbar and Sykes* [1973] call the Boston-Ottawa seismic zone. Only two earthquakes, however, were detected in Vermont during the period, while many events were detected from adjacent parts of Canada at comparable distances from the network. Although the coverage of seismograph stations in Vermont and western New Hampshire was poor until late 1976,

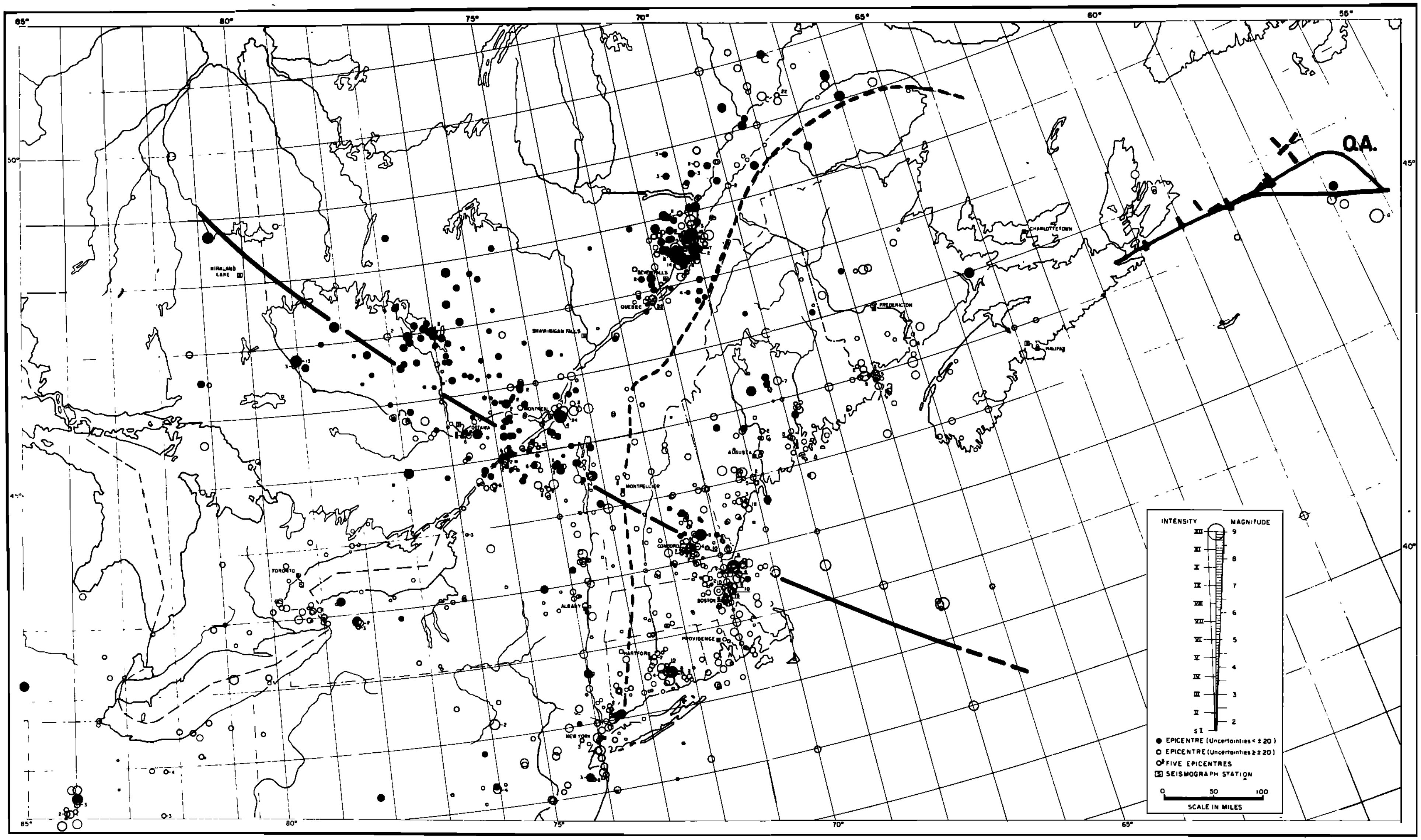

Fig. 13. Seismicity of eastern Canada and northeastern United States, 1534–1959 (after *Smith* [1966], with additions). Although only felt reports were available for most of the earthquakes that occurred before 1927, many later events were located by using instrumental data. Solid line is the same small circle indicated in Figure 12 passing along the New England seamount chain. The northwest trending zone of high seismicity that starts offshore of eastern Massachusetts and extends into Canada to the northwest of Ottawa appears to be nearly coincident with the circle. This data set suggests a gap or lower level of activity along this line in Vermont. Serpentine belt and inferred Ordovician suture zone of *Hess* [1955] and *St. Julien and Hubert* [1975] are indicated by the dashed line. Solid lines labeled OA are faults inferred by *L. H. King and B. McLean* [1970] near the Orpheus gravity anomaly.

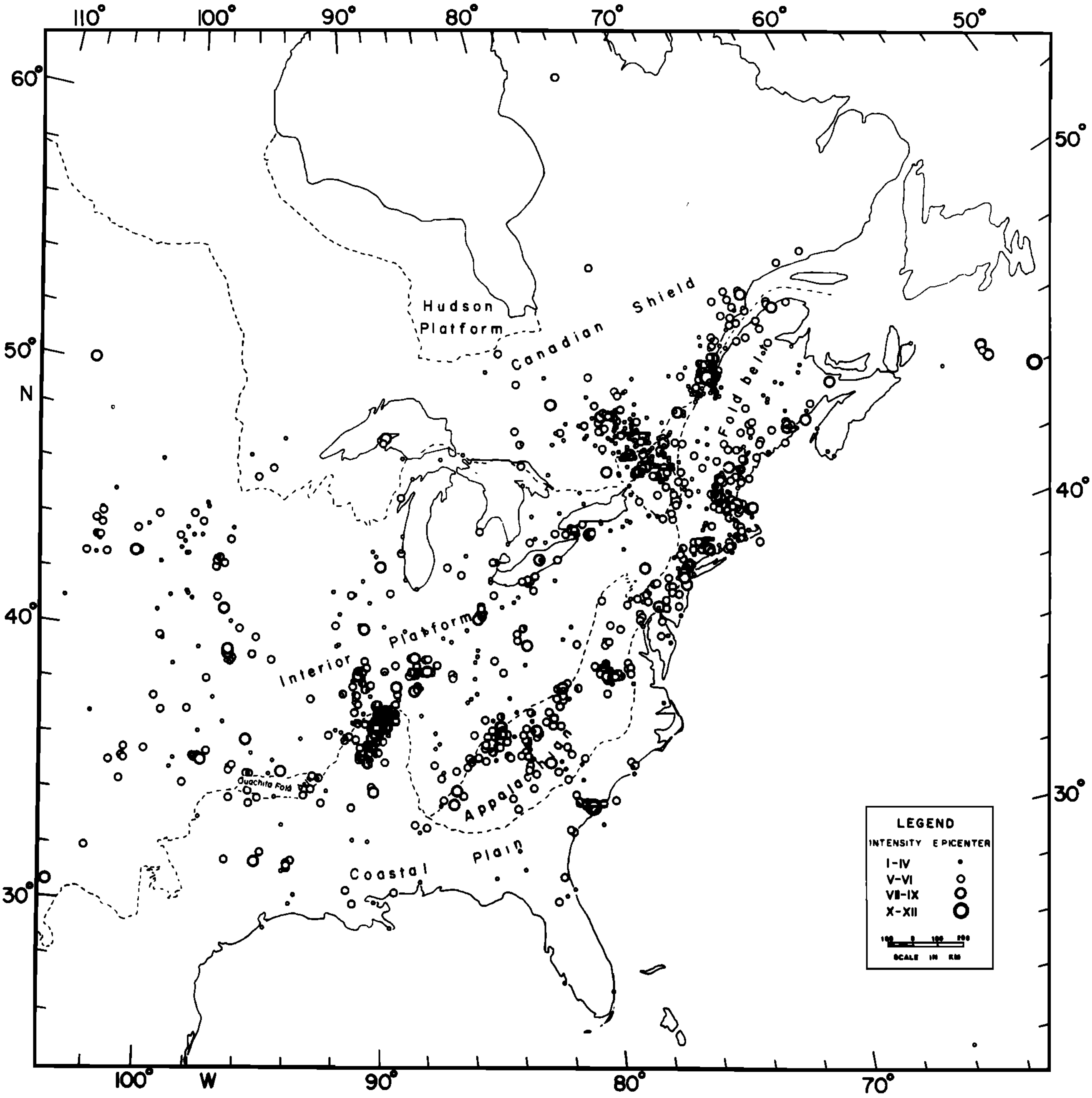

Fig. 14. Distribution of reported earthquakes in eastern North America, 1534–1971, from historical and instrumental data [after *York and Oliver,* 1976]. Note activity along the Appalachian fold belt, the northwest trending zone in New England and southern Quebec, and the northwesterly trend in South Carolina.

the existing evidence from instrumental locations argues for a region of low activity in Vermont. Thus new and historic data indicate that the Boston-Ottawa seismic zone is composed of two distinct zones of high activity: one extending from offshore Massachusetts into central New Hampshire and another extending northwest from northern New York State to Kirkland Lake, Ontario. As will be discussed later, however, alkalic rocks postdating the opening of the western Atlantic extend across the gap in seismic activity in Vermont and western New Hampshire.

Younger Igneous Rocks in New England and Southern Quebec

Although major tectonism and magmatism are commonly thought to have ended in most of the eastern United States in the Triassic, igneous rocks with ages postdating the initial separation of North America from Africa are found in the White Mountain magma series of New England, in the Monteregian Hills of southern Quebec, and along the New England seamount chain (Figure 18). The White Mountain magma series extends NNW across New Hampshire and ranges in age from about 200 m.y. (the initial stage of rifting of North America from Africa) to about 100 m.y. [*Foland and Faul,* 1977]. A small percentage of the radiometric ages are between 220 and 235 m.y. The Monteregian Hills, a group of alkalic mafic and ultramafic rocks in southern Quebec, trend WNW from north of New Hampshire to Montreal. Carbonatites and diatremes are found in the western part of the Monteregian province [*Gold,* 1967]. The $^{87}Sr/^{86}Sr$ ratio of 0.704 obtained by *Fairbairn et al.* [1963] for rocks of the Monteregian province,

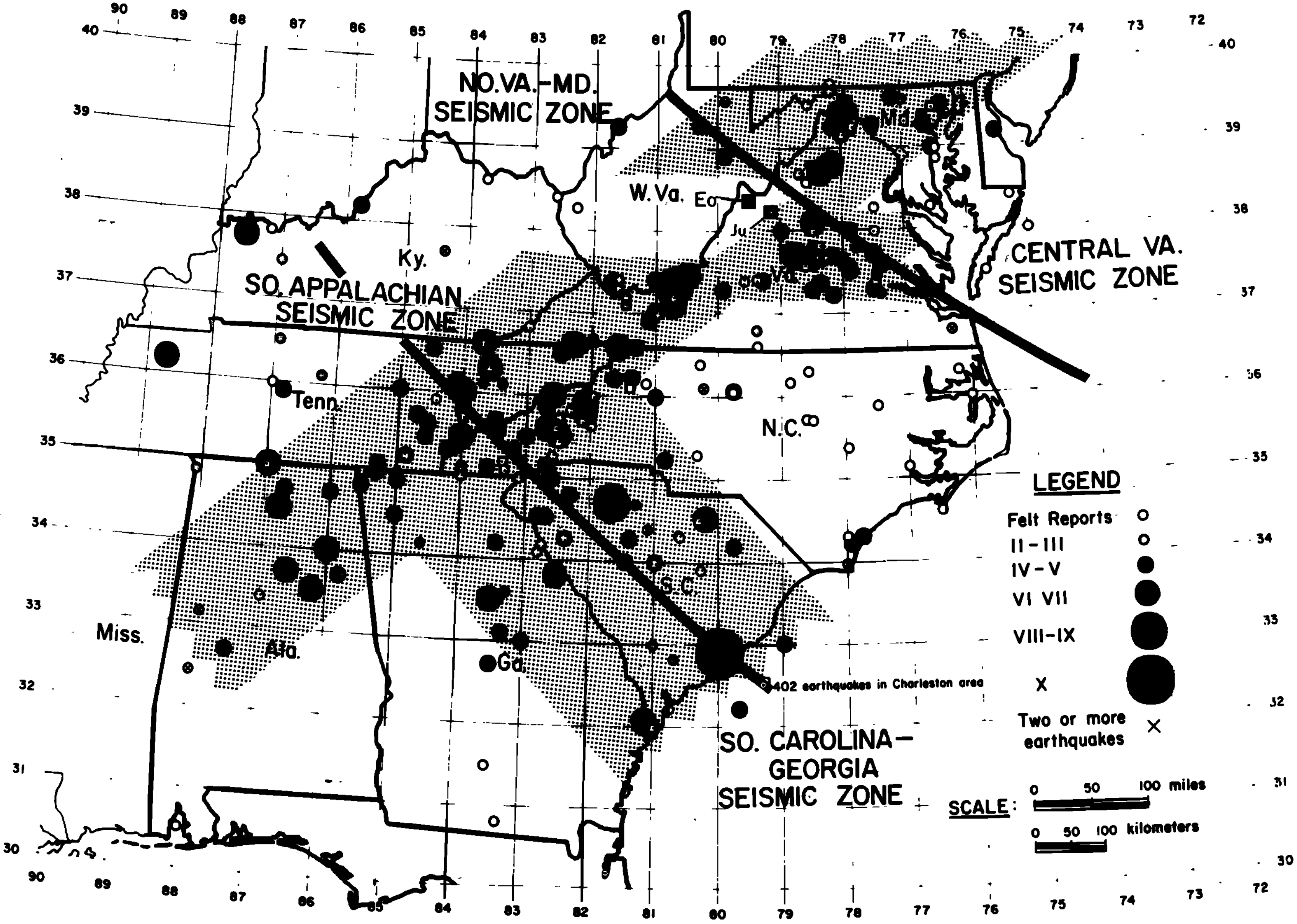

Fig. 15. Historic seismicity for southeastern United States, 1754–1970, after *Bollinger* [1973], with additions. Eo and Ju are localities of alkaline rocks of Eocene and Jurassic age in Virginia. Note that the South Carolina–Georgia seismic belt and central Virginia seismic zone appear to be located along inferred continental extensions of the Blake and Norfolk fracture zones, respectively (solid lines).

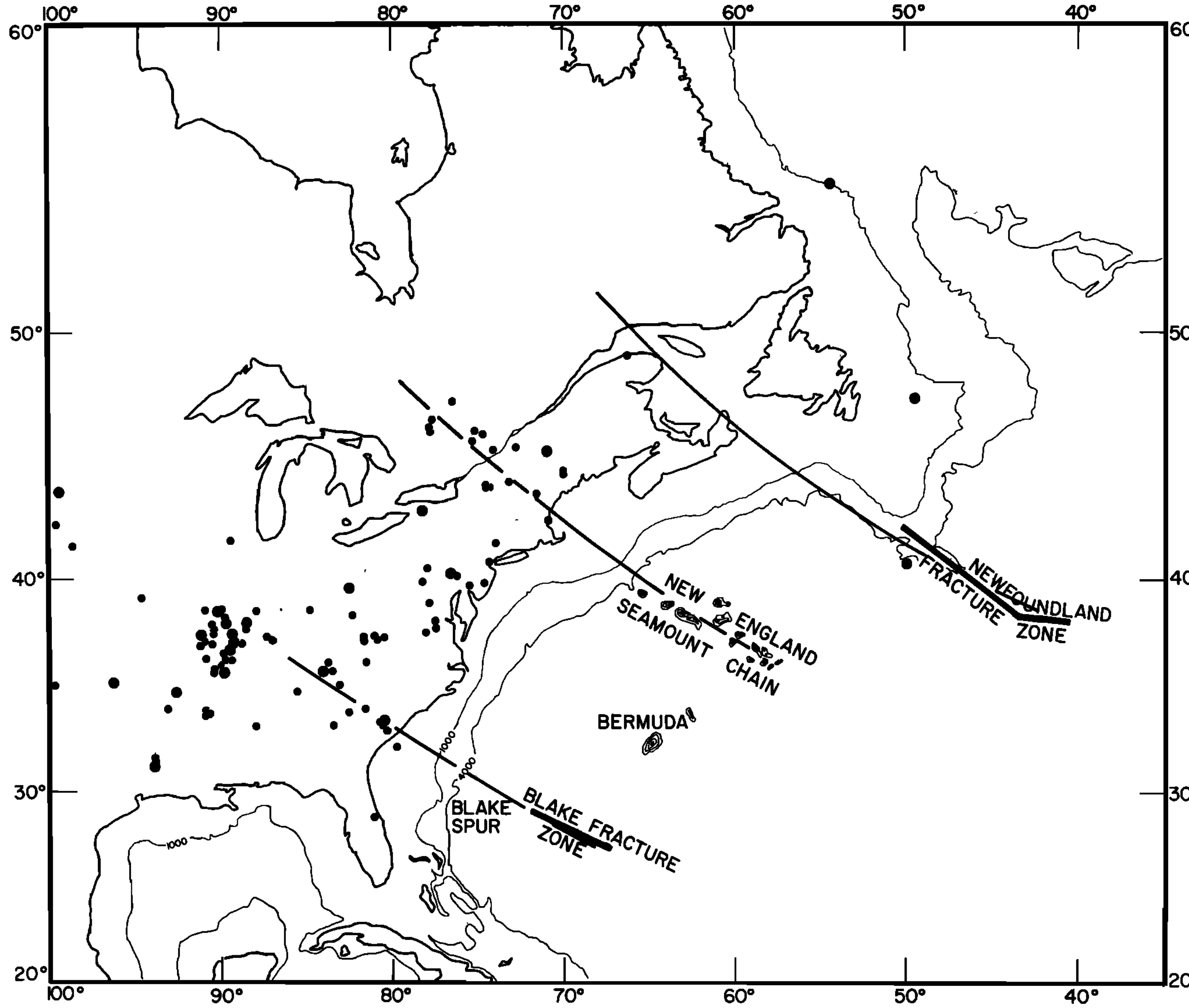

Fig. 16. Seismicity of eastern and central North America, 1961–1974, from data of the National Oceanic and Atmospheric Administration, after *Fletcher et al.* [1978]. Larger dots are earthquakes of magnitude (m_b) greater than or equal to 4.5; the smaller dots are those of magnitude less than 4.5. Lines drawn through three major discontinuities of the magnetic lineation pattern of Mesozoic age in the western Atlantic are as in Figure 12. Depth is in meters. Note the spatial correlation of earthquake epicenters in New England and in adjacent parts of Canada with the small circle extended northwest from the New England seamount chain. The South Carolina–Georgia seismic zone, although not as readily distinguished on this map, is also nearly coincident with the small circle drawn through the Blake fracture zone.

which is nearly identical to that for alkaline igneous rocks in general, is indicative of a mantle origin.

A series of nearly circular positive gravity anomalies in northernmost New York and in adjacent parts of Canada (triangles in Figure 18) trend WNW, subparallel to the Monteregian Hills, which are located about 100 km to the northeast [*Thompson and Miller,* 1958; *Simmons,* 1964; *Diment,* 1968]. *Diment* [1968] suggests that these anomalies are indicative of buried equivalents of the Monteregian Hills.

In addition to the large plutonic features of the White Mountain magma series, a number of smaller stocks and dikes of Mesozoic age have been identified in New England and in southern Quebec. Figure 19 shows a compilation by *McHone* [1975] and *McHone et al.* [1976] of 338 localities of lamprophyre dikes of Mesozoic age in New England and in the Monteregian province. Although the large plutonic bodies of the White Mountain magma series themselves have a somewhat different trend than the transform direction indicated by the solid line in Figure 18, the areal distribution of the entire suite of igneous rocks in Figures 18 and 19 is centered along that line, which also follows the New England seamount chain. Thus in this area the case for tectonic continuity is strong on the basis of the distribution of igneous rocks, while it is weak on the basis of the distribution of earthquakes.

Ottawa-Bonnechere Graben

The Ottawa-Bonnechere graben, one of the major grabens of North America, is located to the WNW of the Monteregian Hills (Figure 18). This fault system is located along the southwestern edge of the region of higher seismic activity shown in Figures 13, 14, and 17. The seismic zone, however, is broader and extends much farther northeast than the mapped faults of the graben. *Kay* [1942] tentatively correlated the age of the graben with that of the Monteregian Hills, i.e., Cretaceous. *Burke and Dewey* [1973], however, suggest that the graben may be of early Paleozoic age, that is, equivalent in age to the youngest sediments that have been displaced stratigraphically.

Rankin [1976] proposes that the Ottawa-Bonnechere graben

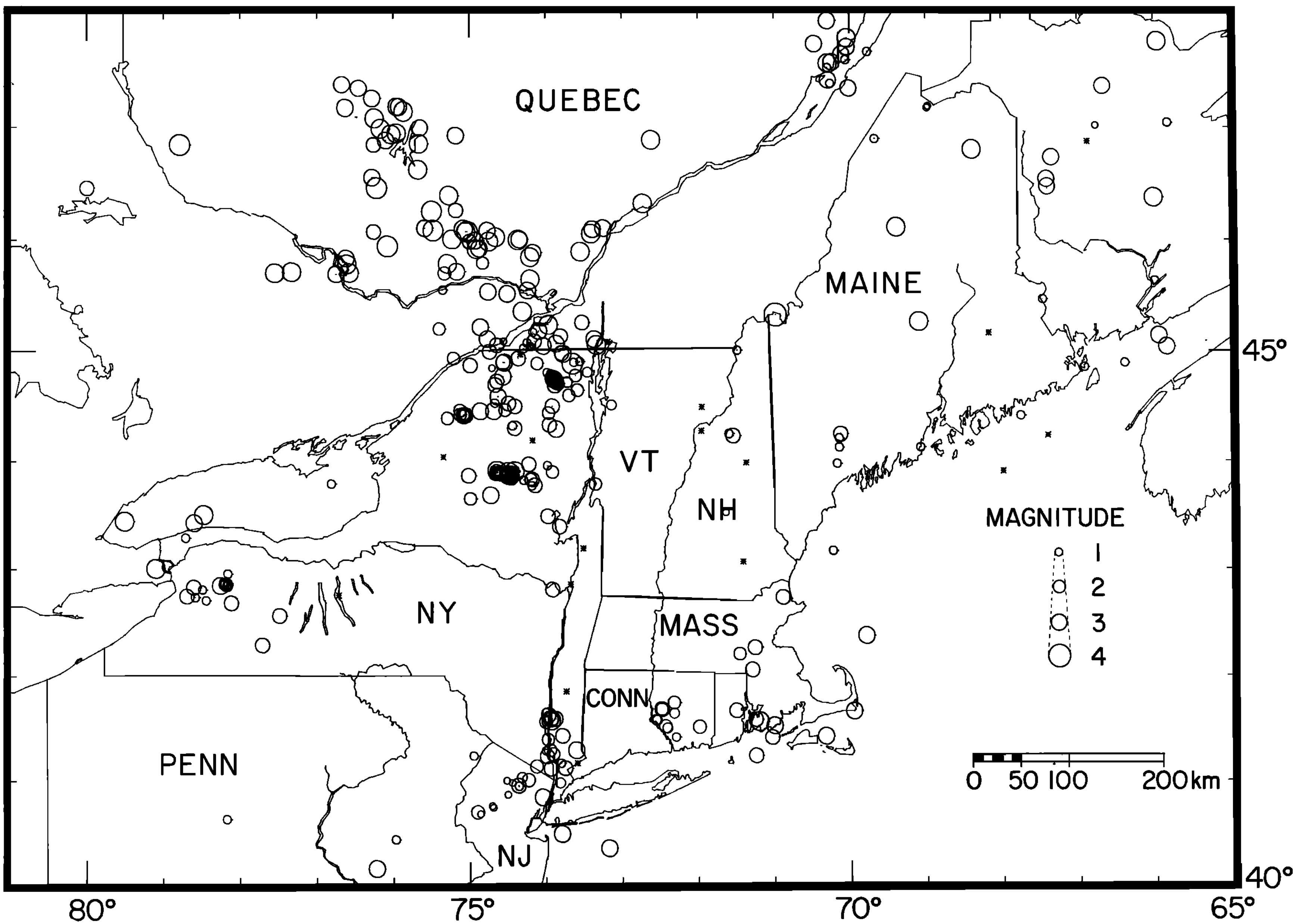

Fig. 17. Epicenters of earthquakes (1970–1977) in northeastern North America located by various networks in the area. For New York State and adjacent areas the coverage is probably complete for magnitudes of >2; for New England the coverage is poorer. Note the northeast alignment of earthquakes in northern New Jersey and southern New York. Asterisks denote unidentified events [after *Aggarwal and Sykes*, 1978].

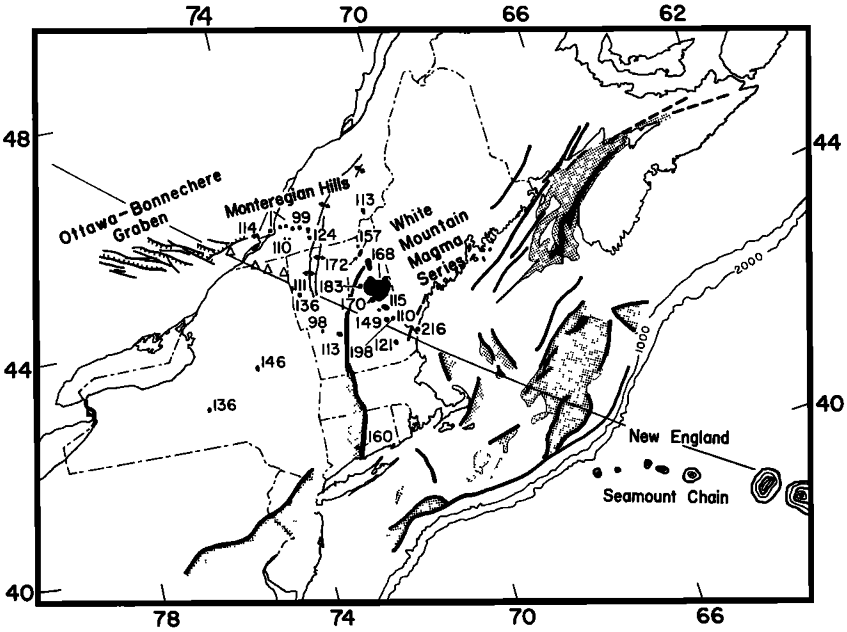

Fig. 18. Geologic features in northeastern United States and eastern Canada with emphasis on those postdating the opening of the Atlantic Ocean, after *Fletcher et al.* [1978]. White Mountain magma series and Monteregian Hills are shown with radiometric ages in millions of years. Triangles mark locations of positive gravity anomalies that may reflect buried equivalents of the Monteregian Hills. Basins of Triassic-Jurassic age are shown by shading, and solid lines are faults of similar age. Depths are in meters. New England salient (light lines) in Vermont and Quebec is shown by trends of main anticlinoria. The White Mountain magma series, Monteregian Hills, and New England seamount chain appear to form a lineation that may be described by a small circle (shown as a thin line trending northwest) drawn about the center of rotation for the early opening of the Atlantic.

developed as a failed arm or rift in the late Precambrian about an RRR triple junction centered near Montreal. He associates Cambrian carbonatites with the development of this failed rift. Evidently, the Mesozoic rocks of the Monteregian Hills represent a reactivation of this older rift. The other two rifts about the triple junction developed into an ocean, of which segments of its continental margin strike north from New York City to Montreal and northeast from Montreal to the Gaspé Peninsula, Quebec. This ocean is thought to have closed in the mid-Ordovician along the suture zone indicated by dashes in Figure 13 [*St. Julien and Hubert,* 1975].

New England Seamount Chain

The New England (Kelvin) seamount chain, which extends for about 1500 km off the east coast of New England (Figures 12 and 18), is one of the major topographic and tectonic features of the Atlantic Ocean. *Drake and Woodward* [1963] speculate that a major strike slip fault follows the seamount chain, curves to the west, and passes south of Long Island. *Dennis* [1956, 1972], *Diment et al.* [1972], *Sbar and Sykes* [1973], and *Fletcher et al.* [1978], however, associate the younger alkalic rocks of New England and the Monteregian province with a lineament extending into the continent from the western end of the New England seamount chain.

In a study of magnetic lineations in the North Atlantic, *Pitman and Talwani* [1972] concluded that the seamount chain was a major transform fault during the early opening of the North Atlantic in the Mesozoic. *Barrett and Keen* [1976] find that magnetic anomalies of the Mesozoic M series are offset across the New England seamount chain and that the strike of these anomalies is quite different north and south of the chain. Thus the seamount chain appears to be a major discontinuity in the magnetic anomaly pattern of Mesozoic age in the western Atlantic. In detail, however, the seamount chain strikes about 15° more west than the nearby fracture zones of Mesozoic age [*Schouten and Klitgord,* 1977]. Probably more than one fracture zone falls near the seamount chain, which is about 100 km wide.

Volcanic breccia from Nashville seamount near the eastern end of the New England seamount chain were dated by the *Deep Sea Drilling Project* [1975] as 78–86 m.y., i.e., about 20 m.y. younger than the oceanic crust surrounding it as determined from magnetic anomalies. A minimum age of 74 m.y. was inferred for Vogel seamount near the center of the chain. These two dates are somewhat younger than the youngest Mesozoic intrusives of New England and southern Quebec (Figure 18). There is no suggestion of a progression in age along the New England seamount chain. Likewise, there is no indication of a spatial progression in age of the young igneous rocks in New England and southern Quebec [*Foland and Faul,* 1977]. Thus the lack of a clear space-time progression seems to rule out an emplacement associated with a moving hot spot beneath the lithosphere. Also it would seem that any reasonable mechanism for the formation of the White Mountain magma series must involve the occurrence of magmatism over a time interval of at least 100 m.y.

Lineation Defined by New England Seamount Chain, Igneous Rocks, and Earthquakes

Sbar and Sykes [1973] and *Fletcher et al.* [1978] calculated a small circle (transform direction) passing along the New England seamount chain and extended it into New England about the center of rotation given by *Pitman and Talwani* [1972] for the Mesozoic movement of North America with respect to Africa. This small circle, i.e., a line of equal latitude about the center of rotation, follows the trend of the New England

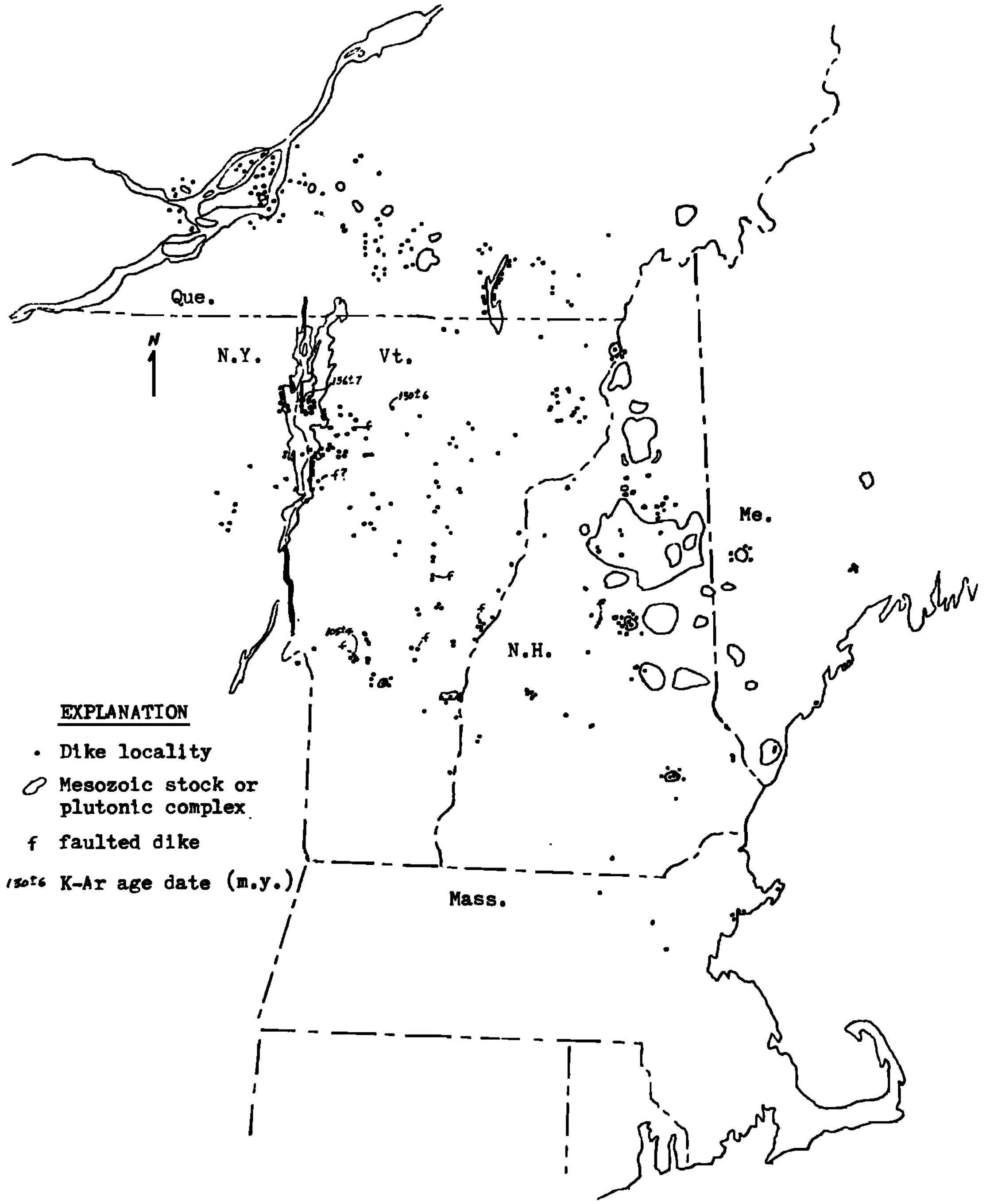

Fig. 19. Localities of 338 mapped dikes in the New England–Monteregian lamprophyre province, after *McHone* [1975] and *McHone et al.* [1976].

seamounts and passes through the region of higher seismic activity extending from offshore Massachusetts to central New Hampshire and that extending from northern New York State to Ottawa and thence to Kirkland Lake, Ontario. Nearly all of the Mesozoic rocks in New England and southern Quebec shown in Figures 18 and 19 and six of the larger historic earthquakes in eastern North America (Figure 12) are also located within 100 km of the same small circle. Thus the New England seamount chain, the Monteregian Hills, the White Mountain magma series, dikes of Mesozoic age in New England and possibly the Ottawa-Bonnechere graben appear to define a tectonic lineament which is nearly 2000 km long and about 100–200 km wide.

The zone of Mesozoic magmatism in the White Mountain magma series and the Monteregian Hills is unusual in that it extends far inland in comparison to magmatism postdating rifting in many other areas. Also, the New England seamount chain with its vast amount of volcanism is also unusual for an oceanic fracture zone. The marked change in strike of Mesozoic magnetic anomalies across the seamount chain suggests that it was a leaky transform fault during at least part of the Mesozoic [*Barrett and Keen*, 1976]. Magnetic anomalies of latest Cretaceous age (anomalies 31 and 32), however, are not offset significantly, nor do they change strike on crossing the seamount chain. This age nearly coincides with the youngest magmatism both in New England and along the seamount chain.

A question arises, If Mesozoic magmatism is nearly continuous along this lineament, why does the present seismic activity appear to be high along two parts of the lineament but low in Vermont and western New Hampshire? One possibility is that the latter is a seismic gap and the low activity for the past 300 years is not representative of the longer-term seismicity. Before leaning toward that interpretation, however, it seems reasonable to entertain other hypotheses that might explain the low activity as a persistent, long-term phenomenon related to differences in the geological and geophysical properties of the lithosphere.

As I mentioned earlier, a late Precambrian to early Paleozoic ocean developed along the east side of the North America continent. When this ocean (or back arc basin) closed in the mid-Ordovician [*St. Julien and Hubert,* 1975], the change in trend of the old continental margin near the Montreal triple junction may have influenced the trends of Paleozoic deformation within those terrains to the east that appear to have been sutured onto the older parts of the continent. The prominent salient in the Appalachian fold belt at the Vermont-Quebec border (Figure 18) may have developed in this way. Major tectonic features of Paleozoic age in New England and Quebec change strike in a similar fashion near a line connecting the New England seamount chain with the Ottawa-Bonnechere graben. When the present Atlantic started to open in the Mesozoic, this Paleozoic lineament apparently was reactivated, and a major transform fault developed seaward of this line along what is now the New England seamount chain.

It is interesting that magmatism in New England and the Monteregian Hills changes drastically both in petrology and in trend near the mid-Ordovician suture zone of Figure 13. Kimberlites and carbonatites are found only to the northwest of the suture on Grenville basement about 1100 m.y. old. The source depth of the Monteregian rocks appears to increase from ESE to WNW [*Gold,* 1967]. To the southeast of this suture, granites, syenites, and ring dikes of the White Mountain magma series are found. This is not surprising, since the age of the basement (and presumably the thickness of the lithosphere) changes from about 1100 to 450 m.y. near the suture zone. Thus the physical properties, age of most recent metamorphism, age of basement rocks, thickness of the lithosphere, and degree to which preexisting faults were reactivated probably change appreciably in passing from the New England seamount chain to the White Mountains and thence to the Monteregian Hills and Ottawa-Bonnechere graben. These factors may greatly influence the present state of stress and the occurrence of earthquakes.

Seismic activity also increases markedly just to the northwest of the Ordovician suture zone (Figures 13, 14, 17, and 20). It appears to be low, however, to the southeast of the suture in Vermont and western New Hampshire. One explanation would be that faults striking northwest along the old failed rift or aulacogen (the Ottawa-Bonnechere graben) were not healed by metamorphism in the Paleozoic, were reactivated in the Mesozoic, and are still active today in response to the present stress field. Faults to the southeast of the suture zone may well have been healed by metamorphism during the early and middle Paleozoic.

Focal mechanism solutions of earthquakes in northern New York and a solution for a shock northwest of Ottawa indicate thrust faulting along northwest to NNW striking fault planes [*Yang and Aggarwal,* 1977]. Focal mechanism solutions are not yet available for shocks in New Hampshire or offshore Massachusetts.

Seismic activity in the zone extending from offshore Massachusetts to central New Hampshire is similar to that seen in other continental areas in that it is located near the end of a major oceanic transform fault. Using marine geophysical data, *Kane et al.* [1972] and *Ballard and Uchupi* [1972, 1974, 1975] infer NNW and north trending faults (Figure 18) off the east coast of Massachusetts and in the Gulf of Maine. Unfortunately, the epicenters of the large shocks of 1638 and 1755 in this area (Figure 12) are not known well enough to relate them to specific faults. Although several magnetic anomalies at sea along this lineament have been interpreted as arising from rocks similar to those of the White Mountain magma series [*Ballard and Uchupi,* 1972; *Rand,* 1976], no radiometric ages are available for the crucial 500-km part of the inferred lineament between the northwesternmost Kelvin seamounts at the base of the continental margin (Figure 18) and the southeasternmost Mesozoic rocks in New Hampshire.

Ballard and Uchupi [1972, 1975] also infer that a series of Carboniferous and Triassic grabens extend southwest across the Gulf of Maine and appear to be offset by two major NNW striking fault zones (Figure 18). One of these lies along the lineament inferred to connect the New England seamounts with the younger Mesozoic rocks of New England. This lineament also appears to delimit the southwestern end of a major sedimentary basin found beneath Georges bank on the outer continental shelf off Massachusetts (Figure 21). At the present time, however, it appears to me to be premature to attempt to associate the shocks of 1638 and 1755 with this lineament definitively, as some have attempted to do in legal proceedings involving the seismic safety of nuclear reactors. As in the case of the Charleston, South Carolina, region, where a buried Triassic basin has been found [*Marine and Siple,* 1974], it is also possible that the large historic shocks off Massachusetts may be associated with Triassic structures striking northeast. Marine geophysical data from the outer continental shelf of Massachusetts, where the basement is buried beneath up to 10 km of sediment [*Sheridan,* 1974] should be crucial in understanding the possible connection among historic earthquakes, young igneous rocks, and the New England seamount chain.

Grand Banks and South Carolina Seismic Zones

Since a small circle drawn about the center of rotation for the early opening of the North Atlantic appears to define a lineament that includes the New England seamount chain, the White Mountain magma series, the Monteregian Hills, and regions of higher seismic activity, *Fletcher et al.* [1978] drew two similar small circles (transform directions) through the other major breaks in the magnetic anomaly pattern of Mesozoic age in the western Atlantic. The next major discontinuity to the north of the New England seamount chain is found at the Newfoundland fracture zone [*Auzende et al.,* 1971; *Pitman and Talwani,* 1972], while the next major break to the south occurs at the Blake fracture zone [*Vogt et al.,* 1971], where the Mesozoic magnetic lineation pattern is offset left laterally about 100 km [*Pitman and Talwani,* 1972]. By using the Mesozoic center of rotation, transform directions were drawn through the Newfoundland and Blake fracture zones and were extended into eastern North America (Figures 12, 15, and 16).

Newfoundland fracture zone. The large Grand Banks earthquake of magnitude 7.2 in 1929 is located very close to the end of the large Newfoundland fracture zone. Several historic and instrumentally located earthquakes define a northwest trending zone near the Grand Banks (Figures 13, 14, 16, and 20). The abrupt change in the strike of the continental margin southeast of Newfoundland (Figure 16), the fact that the greatest offset of the entire Atlantic margin of North America occurs there, and other geological and geophysical data [*Auzende et al.*, 1971; *Keen and Keen*, 1973] indicate that this southeast trending margin was the site of one of the largest transform faults involved in the early opening of the North Atlantic. As in the case of the New England seamount chain, this transform fault also developed seaward of a major bend in the Appalachian orogenic belt [*Rankin*, 1976]. Logan's line, which marks the northwesterly limit of thrusting in the northern Appalachians, and an Ordovician belt of ultramafic and other ophiolitic rocks (Figure 13) on the Gaspé Peninsula and in western Newfoundland [*St. Julien and Hubert*, 1975] both have an apparent right lateral offset of several hundred kilometers in the Gulf of St. Lawrence landward of the Newfoundland fracture zone. Thus the Newfoundland fracture zone seems to have developed along a major preexisting tectonic feature oriented transverse to the Appalachians.

Several workers have proposed a structural origin for the Laurentian channel, which lies between Newfoundland and Nova Scotia, while others conclude that it is of glacial origin. *Sheridan and Drake* [1968] found little evidence to support a southeast trending structure in the channel. On seismic reflection profiles, however, *K. L. King and B. MacLean* [1970] find numerous instances of folding of Cretaceous (but not Tertiary) strata at the mouth of the Lauentian channel along the Orpheus gravity anomaly. They infer that a fault system (OA in Figure 13) of Cretaceous and possibly early Tertiary age strikes northwest near the epicenter of the Grand Banks earthquake of 1929, bends to the west to follow the western portion of the Orpheus gravity anomaly, and then joins the Cobequid-Chedabucto fault system of Nova Scotia. No evidence of post-Triassic faulting, however, has been found for the landward extension of this proposed fault system. There is no support from the epicenters in Figures 13, 14, 16, and 20 for continuing the zone of Cretaceous or younger deformation onto land in Nova Scotia. Likewise, there are very few historic reports of earthquakes in the Gulf of St. Lawrence (Figure 20). Since coverage by seismograph stations for these areas was poor until the St. John's (Newfoundland) station was installed in the 1960's, the Grand Banks region is probably more active than the historic and instrumental locations might suggest. Thus earthquakes and Cretaceous to Tertiary deformation are indicative of intraplate tectonism near the western end of the Newfoundland fracture zone. Whether this deformation extends farther into the continent is problematical.

Lower St. Lawrence Valley. A number of historic and instrumentally located shocks are evident in Figures 13 and 20 along the lower St. Lawrence Valley northeast of Quebec City. Some, but by no means all, of these events, including the large shock of 1925 and its aftershocks, are located within or close to the Charlevoix impact structure near La Malbaie, Quebec, which is of Ordovician to Devonian age [*Le Blanc et al.*, 1973]. Le Blanc et al. located a number of microearthquakes within the impact structure, which is about 30 km in radius. They conclude that earthquakes tend to be concentrated within the weakened crust of the impact structure. Here again, preexisting faults, albeit of different origin, appear to control the release of seismic energy.

The other shocks along the lower St. Lawrence do not have a ready tectonic explanation. They follow Logan's line, the northwest limit of Ordovician thrust faulting, and are located about 50–100 km northwest of a major Ordovician suture zone in the Canadian Appalachians [*St. Julien and Hubert*, 1975].

Kumarapeli and Saull [1966] propose that a major rift system, which they believe developed in the Mesozoic, follows the St. Lawrence Valley, the Ottawa-Bonnechere graben, and the Lake Champlain lowland of Vermont and northeastern New York State. They hypothesize that earthquakes and alkaline rocks in those areas are associated with the rift system. Focal mechanism solutions in northern New York State and in adjacent parts of Canada near Ottawa, however, indicate thrust faulting along planes striking northwest to NNW [*Yang and Aggarwal*, 1977]. While these solutions do not support the idea of contemporary rifting, it is possible that a small amount of extension occurred during the early opening of the Atlantic along the lower St. Lawrence Valley near the Ordovician suture zone. As was mentioned before, however, earthquakes do not appear to be continuous along the entire St. Lawrence Valley. Activity near Ottawa does not appear to be continuous with that northeast of Quebec City.

Blake fracture zone and the Charleston earthquake. The transform direction drawn through the Blake fracture zone in Figures 12 and 16 passes very close to the epicenter of the Charleston earthquake of 1886 and is nearly coincident with the South Carolina–Georgia seismic zone of *Bollinger* [1973] (Figure 15), which contains the Charleston shock. The transform direction drawn through the Blake fracture zone passes through the Blake spur, a major structural and bathymetric feature of the continental margin at the edge of the Blake plateau (Figures 16 and 21). Using marine geophysical data, *Sheridan* [1974] and *Schouten and Klitgord* [1977] propose that a fracture zone of northwesterly trend passes through the Blake spur (Figure 21). One of the largest offsets of the Blake spur magnetic anomaly occurs at that fracture zone. Also, the M sequence of magnetic anomalies is offset farther southeast. Numerous dikes believed to be of Jurassic age strike northwest in South Carolina and Georgia [*P. B. King*, 1961; *de Boer*, 1967].

Rocks similar to those of the White Mountain magma series and the Monteregian Hills do not appear to have a surface expression along the lines drawn through the Newfoundland and Blake fracture zones. K-Ar ages of 95 and 109 m.y. were obtained from basalts encountered in a deep borehole in the Charleston area [*Rankin*, 1977]. These probably represent minimum ages, however. The chemistry of the rocks most nearly resembles widespread tholeiites of Triassic-Jurassic age that are found in other parts of eastern North America. The minimum ages may reflect post-magmatic processes associated with tectonic activity of Cretaceous age [*Rankin*, 1977]. *Popenoe and Zietz* [1977] identify a number of prominent magnetic anomalies in South Carolina and adjacent parts of Georgia that they infer are indicative of mafic and ultramafic rocks that are buried beneath the sediments of the coastal plain. It is possible that rocks similar in age and composition to those in New England and in the Monteregian province are prominent at depth beneath the coastal plain and continental margin of South Carolina. Another possibility is that the area underlain by this distinctive geophysical basement, the Charleston block, may represent a broad zone of Triassic and (or) Jurassic crustal extension formed during the early stages of the opening

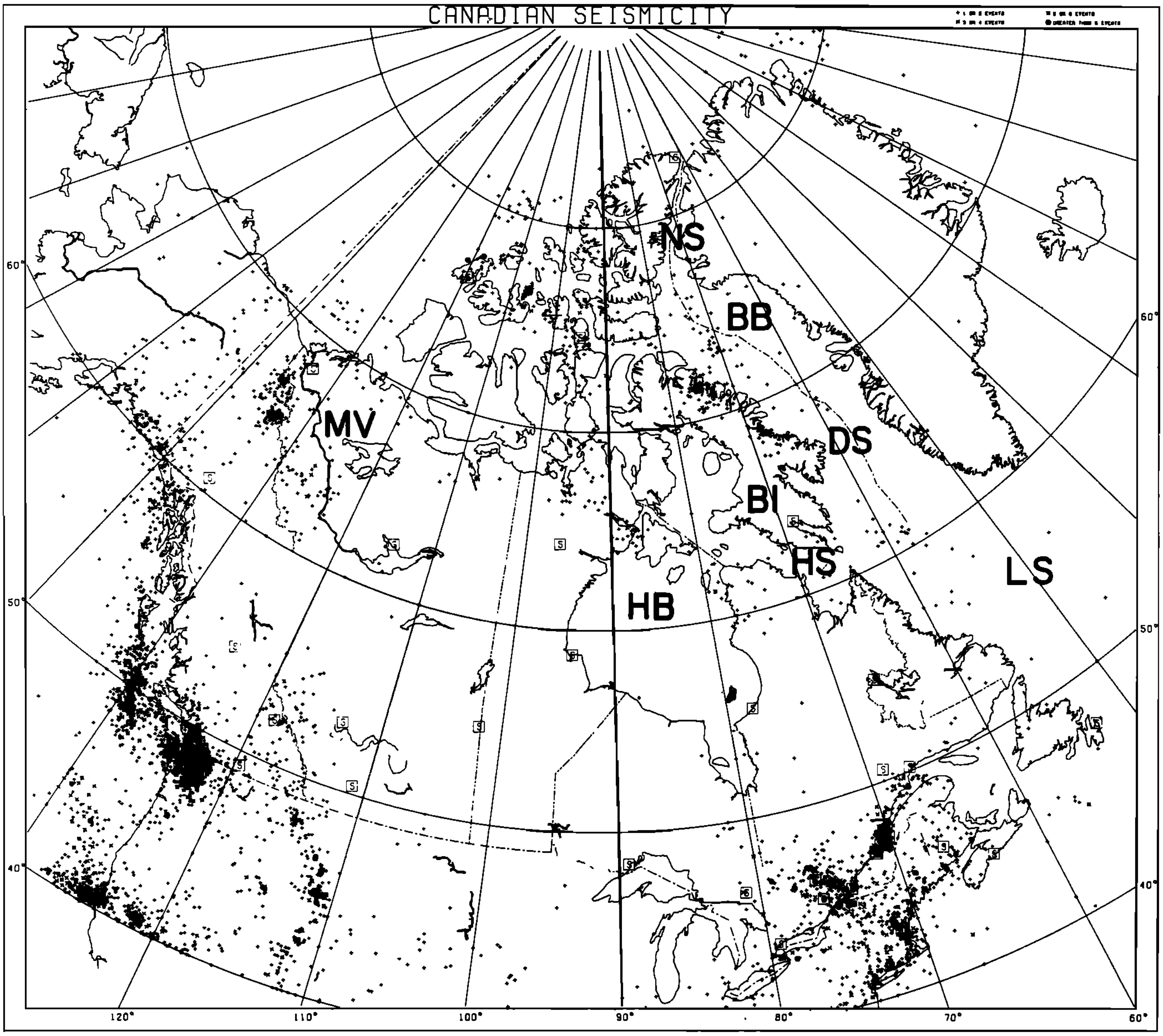

Fig. 20. Epicenters of about 6000 earthquakes in Canada and adjacent areas to 1974 (from files of the Earth Physics Branch, Department of Energy, Mines and Resources, Canada; P. W. Basham, written communication, 1977). Coverage was arbitrarily cut off south of 40°N in the United States and in other areas outside Canada. Note the very low level of activity in much of the Canadian shield and the higher levels on almost all sides of the shield. BB is Baffin Bay, BI Baffin Island, HB Hudson Bay, HS Hudson Strait, LS Labrador Sea, DS Davis Strait, MV Mackenzie Valley, and NS Nares Strait. Pluses are for one or two events, crosses for three or four events, asterisks for five or six events, and solid circles for more than six events.

of the Atlantic Ocean [*Rankin,* 1977; *Popenoe and Zietz,* 1977].

On marine reflection profiles taken parallel to the continental margin off the coast of South Carolina, *Dillon* [1974] finds faults that appear to cut the Tertiary section and that are located along the trend of the seismic activity that passes northwest through the state (Figure 15). *Colquhoun and Comer* [1973] made a series of seismic reflection profiles in the rivers near Charleston. These surveys reveal structural deformation in near-surface sediments along a feature which they called the Stono arch. The deformed section includes Tertiary sediments, but they could not determine if folding affects Pleistocene or Holocene sediments. The Stono arch trends west to northwest in the Charleston area. *Tarr* [1977] obtained a focal mechanism solution for an earthquake near Charleston in 1974, which involved thrust faulting along a plane striking N43°W, i.e., nearly parallel to the South Carolina–Georgia earthquake zone of Figure 15. A number of earthquakes located using data from a network of stations in South Carolina are confined to a northwest trending zone about 25 km long between Charleston and Summerville, South Carolina [*Tarr,* 1977]. Thus various types of geophysical evidence from the Atlantic Ocean seaward of Charleston, drilling and magnetic surveys on land, a focal mechanism, and the distribution of earthquakes suggest that a major lineation extending from the Blake fracture zone includes the epicenter of the Charleston earthquake and the South Carolina–Georgia seismic zone.

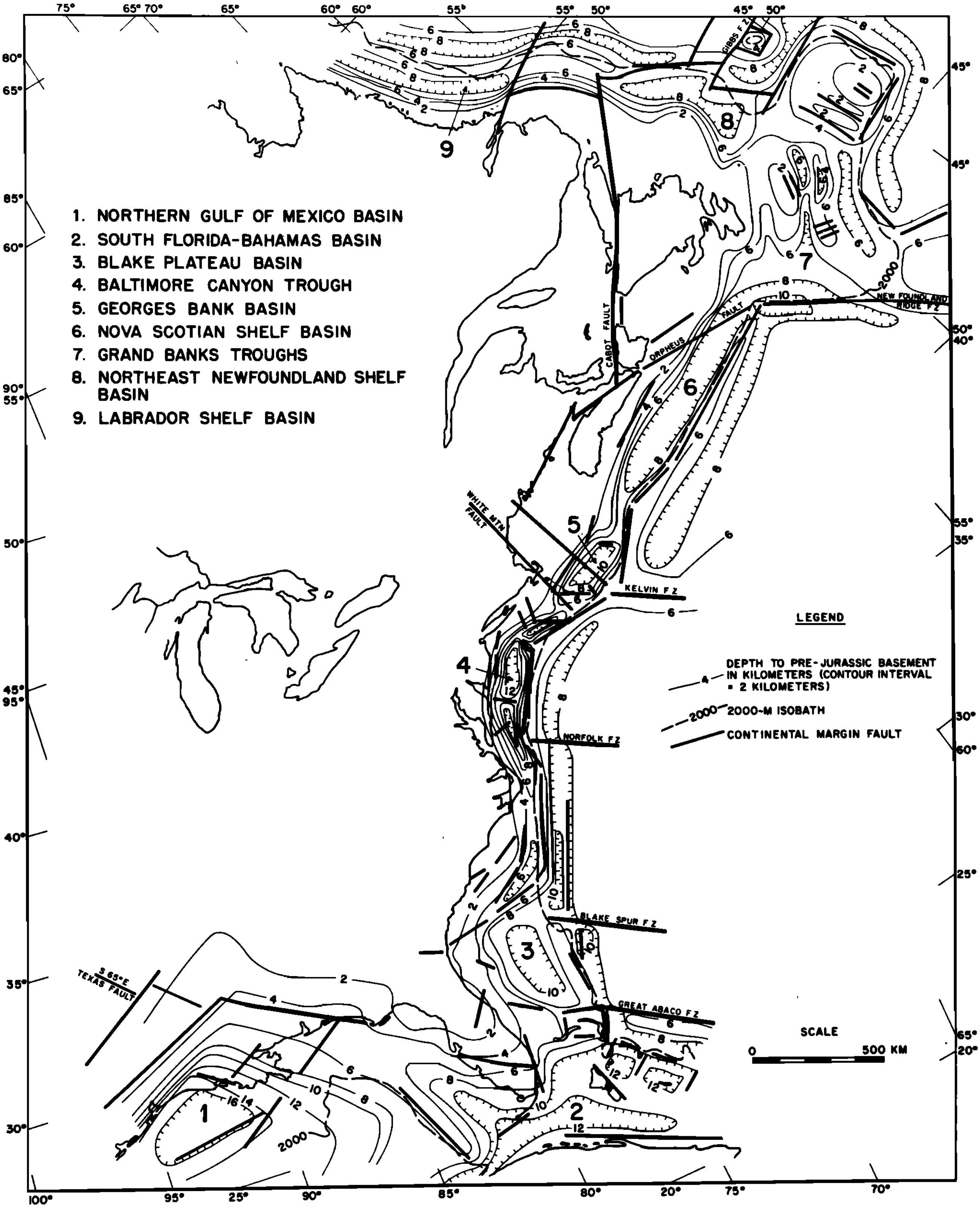

Fig. 21. Generalized basement map on the pre-Jurassic for the east coast of North America, after *Sheridan* [1974]. Large numerals correspond to major sedimentary depocenters along the margin.

As will be discussed in a later section, northeast trending faults parallel to the Appalachian orogen also appear to be active in the southeastern United States. At the present time it is not possible to ascertain how far inland the northwesterly zone of activity extends. We do not know, for example, if much of the activity in central and western South Carolina represents faulting along planes striking northwest or along planes parallel to the Appalachian fold belt. Similarly, the strike of the fault that moved during the Charleston shock of 1886 is also unknown. *Taber* [1914] concludes that it was northeasterly from the elongated nature of the isoseismals. Clearly, additional mechanism solutions would resolve this question.

Although a transform direction drawn through the Blake fracture zone and extended into North America passes close to the zone of high seismic activity in the New Madrid area of southeastern Missouri (Figures 12 and 16), there is no evidence from historical compilations of earthquakes or from the occurrence of younger rocks that the South Carolina–Georgia zone continues across central and western Tennessee to the New Madrid region. Much of the activity in the New Madrid region appears to be associated with the northeast trending Mississippi Embayment. *Bucher* [1963] placed considerable emphasis on northwest trending faults of Cretaceous age in western Tennessee, western Kentucky, and southern Illinois, since they appear to connect several cryptovolcanic features that he believed were of internal origin to the earth. Several of these cryptovolcanic features are now regarded as being of extraterrestrial origin and should not be considered to define a tectonic zone of northwesterly trend.

Other Fracture Zones and Segmented Nature of Continental Margin

In addition to the three fracture zones described thus far, *Sheridan* [1974, 1976] also infers (Figure 21) two other fracture zones, the Norfolk and Great Abaco, off the east coast of the United States. He concludes that a series of sedimentary basins found along the east coast of North America are separated from one another by transversely striking oceanic fracture zones (Figure 21). A major fracture zone may also be present along the Bahamas escarpment off southern Florida, which is conjugate to the Guinea fracture zone (Figure 5) of western Africa [*Pitman and Talwani*, 1972]. Very little seismic activity is known to have occurred, however, near the Guinea or Great Abaco fracture zones or in the Bahamas and southern Florida.

Dillon et al. [1976] report the interruption of a series of subbottom horizons on a multichannel reflection profile east of the Blake escarpment off central Florida. They interpret these interruptions, which offset horizon A but not reflectors above, as evidence of early Tertiary or possibly late Cretaceous faulting along a fracture zone that was reactivated by subsidence of the Blake plateau.

The pattern of sedimentary basins separated by topographic highs appears to be a persistent feature of the continental margins of eastern North America and the Gulf coast of the southern United States from the Mesozoic into the Cenozoic [*Durham and Murray*, 1967; *Owens*, 1970; *Sheridan*, 1974, 1976]. The warping of Pleistocene shorelines along the coastal plain from North Carolina to Florida indicates that the earlier pattern of topographic highs and lows persists into the Quaternary [*Winker and Howard*, 1977]. Thus it would not be surprising if present seismicity is related in some way to the block or segmented character of the continental margin. Of course, an important parameter is how far into the continent this pattern extends.

The Cape Fear arch forms a sill that separates sedimentary basins on the outer continental shelf off Virginia from those off South Carolina (Figure 21). It may represent a block of the margin that did not subside as much as those to the north and south. The adjacent area of North Carolina has a history of very low seismic activity (Figures 14, 15, and 16). The area of low activity is approximately bounded by transform directions extended inland from the Blake and Norfolk fracture zones.

Central Virginia Seismic Zone, Norfolk Fracture Zone, and Alkaline Magmatism

Schouten and Klitgord [1977] recently mapped a number of fracture zones that interrupt the Mesozoic M sequence of magnetic anomalies off the east coast of the United States between the New England seamount chain and the Blake fracture zone. Most of these fracture zones have only a small apparent offset. The east coast magnetic anomaly, however, experiences one of its greatest offsets, about 50 km, near the Norfolk fracture zone (Figure 21). *Bollinger*'s [1973] central Virginia seismic zone (Figure 15) is located inland from that fracture zone. It extends from the fall line near Richmond across the Piedmont to the Blue Ridge mountains [*Bollinger*, 1975]. It is difficult to ascertain from the historical earthquakes in Figure 15, however, how far this seismic zone extends eastward beneath the coastal plain or to determine its strike very precisely.

Eocene intrusives in west central Virginia (Eo in Figure 15) are the youngest known igneous bodies in the eastern United States [*Dennison and Johnson*, 1971]. They are situated to the west of the central Virginia seismic zone. This area also has the most pronounced development of thermal springs in the eastern United States. It is situated near a dome on the Schooley erosion surface of early Tertiary–Late Cretaceous age. *Zartman et al.* [1967] obtained radiometric dates of about 150 m.y. for a suite of alkalic rocks located just to the east of these Eocene intrusives (Ju in Figure 15). *Dennison and Johnson* [1971] observe that this area is one of three places in the Appalachians where diabasic dikes of Triassic-Jurassic age cross the Blue Ridge anticlinorium and extend into the western flank of the Appalachians.

Zartman et al. [1967] and *Dennison and Johnson* [1971] associate these younger rocks with a major structural lineament extending from Kansas to Virginia along 38°N. Many of the igneous rocks and faults found farther west along this proposed lineament in Kentucky and Missouri are dated as late Paleozoic [*Zartman et al.*, 1967; *Heyl*, 1972]. Since the igneous rocks in Virginia and the central Virginia seismic zone fall nearly along a transform direction extended inland from the Norfolk fracture zone, they may be related instead to a preexisting zone of weakness striking WNW to northwest that was reactivated during the development of the western Atlantic.

Cederstrom [1945] found a marked change in the thickness of Upper Cretaceous and Eocene sediments near the James River in southeastern Virginia. To account for this change, he proposed that a WNW striking fault, the Hampton Roads fault, is present along the James River. He proposed that it extends as far west as the fall line (the eastern boundary of the Piedmont as well as the Appalachian fold belt in Figure 14). He also found that the Eocene-Miocene contact is gently warped along that line. *Rogers and Spencer* [1971] find a change in piezometric surfaces and in the chloride content of groundwater across the fault. They state that displacement continues through Pleistocene sediments and cite Bollinger's central Virginia seismic zone as additional evidence for the fault. They infer that periodic movements along the Hampton Roads fault appear to have allowed marine waters to move into pre-Miocene sediments. Thus young igneous rocks, earthquakes, and Cenozoic faulting and folding are present along a zone extending several hundred kilometers inland from the end of the Norfolk fracture zone.

Marine magnetic anomalies of Mesozoic age appear to change strike drastically both at the Norfolk fracture zone [*Schouten and Klitgord*, 1977] and at its conjugate point in the eastern Atlantic near 23°N off Cape Blanc [*Hayes and Rabinowitz*, 1975]. Only the older Mesozoic magnetic anomalies [*Schouten and Klitgord*, 1977] appear to be offset at the Norfolk fracture zone. Thus that fracture zone may have developed initially adjacent to an older line of weakness in the continent but did not persist as a major offset on the ocean floor after the Jurassic.

Cretaceous Magmatism off New Jersey

A large igneous body of middle to Late Cretaceous age near the edge of the continental shelf of New Jersey [*Mattick*, 1977] has been studied extensively by the petroleum industry. It is situated near the end of one of the northwest trending fracture zones mapped by *Schouten and Klitgord* [1977] in the westernmost Atlantic. Thus igneous bodies postdating continental separation of North America and Africa are found near the ends of oceanic fracture zones in at least four places in eastern North America: New Hampshire, South Carolina, Virginia, and offshore New Jersey.

Earthquakes Along the Appalachian Fold Belt

Clearly, a number of the earthquakes in eastern North America (Figures 13, 14, 15, 16, 17, and 20) are not located along the northwest striking cross trends described thus far in New England, South Carolina, and Virginia. Other activity appears to follow the Appalachian fold belt [*Woollard*, 1958, 1965; *Fox*, 1970; *Oliver et al.*, 1974; *Brown and Oliver*, 1976]. Most of the earthquakes from New York City to Alabama tend to be concentrated along or to the west of the fall line. In that area, relatively few shocks are found along the coastal plain or along the continental margin except in South Carolina and Virginia (Figures 14 and 15).

Reactivation of Paleozoic and Mesozoic Faults

There is now mounting evidence that older faults oriented subparallel to the Appalachians have experienced late Mesozoic or Cenozoic movement and that they probably are the loci of many of the earthquakes of the region. *Woollard* [1958] concludes that epeirogenic uplift is in progress in the Appalachian region, which gives rise to earthquakes along older fractures and along the boundary faults of major tectonic units. *Fox* [1970] notes that seismic activity is high in the highly faulted portions of the Piedmont, Blue Ridge, and Valley and Ridge provinces of the Appalachians and is generally low in the less faulted Appalachian plateau and central interior region to the west of the Appalachian orogen. *Berkey and Rice* [1919], *Stose* [1927], and *White* [1950, 1952] cite evidence for post-Late-Cretaceous faulting near the boundary of the Piedmont and Blue Ridge provinces. *Conley and Drummond* [1965] describe faults that displace alluvial and colluvial deposits of probable Pleistocene age along the Blue Ridge front near Saluda, North Carolina. *Dryden* [1932], *Oliver et al.* [1970], *Sbar and Sykes* [1973, 1977], and *York and Oliver* [1976] mention a number of examples of Cretaceous and Cenozoic faulting in eastern North America; *Stephenson* [1928] cites similar examples for the coastal plain of the Gulf of Mexico.

The lack of post-Triassic faulting on many geologic maps of eastern North America can be attributed not only to the relatively low-to-moderate level of tectonic activity compared to, say, California but also to the fact that most geologists have been interested in Paleozoic or older deformation and were not actively looking for evidence of more recent faulting. In fact, in a recent research program aimed at understanding the geologic basis of the Charleston earthquake, several examples of deformed strata of post-Late-Cretaceous age were found in the western part of the coastal plain of South Carolina and Georgia [*Owens et al.*, 1976; *O'Connor and Prowell*, 1976]. In that area the Belair fault displaces strata 50 m.y. old, and considerable debate has centered on whether it also has offset Quaternary sediments [*U.S. Army Corp of Engineers*, 1977].

Aggarwal and Sykes [1978] find that most of the better located earthquakes (Figures 17 and 22) in northern New Jersey and southeastern New York are located on or near major northeast striking faults. About half of the shocks in that area occur along the Ramapo fault system, which forms the northwestern margin of the Triassic-Jurassic Newark graben in that area. The fault system appears to have a long history of movement including deformation in the Precambrian, Paleozoic, Triassic, and Jurassic [*Ratcliffe*, 1971; *Dames and Moore*, 1977]. Dames and Moore obtained radiometric ages of 92 m.y. on an apophyllite and a minimum age of 73 m.y. on zeolites from the Ramapo fault system in Rockland County, New York. *Offield* [1967] describes repeated movements in the Hudson Highlands along some of the other northeast striking faults of Figure 22. The Cortland igneous complex of Ordovician age is located within a few kilometers of the Ramapo fault [*Ratcliffe*, 1971]. The petrology of the complex suggests a mantle origin. Thus the Ramapo fault system appears to be a major preexisting feature, at least a part of which extended through the lithosphere during the Ordovician.

The hypocenters (Figure 23) of recent, well-located earthquakes define a plane extending from the surface trace of the Ramapo fault to a depth of about 10 km with a dip of about 60°SE. Focal mechanism solutions of several shocks along the fault yield strikes and dips that agree very closely with those inferred from the distribution of hypocenters and from surface geologic measurements [*Aggarwal and Sykes*, 1978]. Several additional mechanism solutions for shocks along nearby faults indicate movement along planes striking NNE to northeast. Thus faults with a history of repeated movements appear to be the locus of present seismicity in that area.

The study by *Aggarwal and Sykes* [1978] is one of the first in the eastern or central United States in which a sufficiently dense network of seismographs was used so that epicenters could be located with enough precision (better than 1–2 km) to demonstrate a correlation with specific faults. Frequently, the question is raised, Would shocks of that size be detected in any area when a net like the above one is placed in operation? The answer is clearly no. Since 1970 the network of stations in the northeastern United States detected very few events (Figure 17) in a broad area of central New York and central and western Pennsylvania. The pattern of activity is very similar to that of the much longer historic record of felt shocks (Figures 1, 13, and 14). Activity in southern New York State dies off abruptly near the northwestern boundary of the Precambrian Hudson Highlands (Reading prong). Thus the seismicity appears to be closely related to major geologic and tectonic units that predate the Mesozoic opening of the Atlantic.

Seismic activity in Connecticut (Figure 17) is concentrated near the eastern side of the Triassic-Jurassic graben of the Connecticut Valley. In New York and New Jersey the main depocenters and faults of Triassic-Jurassic age are found near the northwestern side of the Newark basin, while those in the Connecticut Valley are found along the eastern side of that basin. In both cases, seismic activity is concentrated on the side of the half graben that experienced large normal faulting in the Triassic and Jurassic.

Aggarwal and Sykes [1978] find that shocks in southern New York and northern New Jersey are characterized by reverse faulting. Hence the present sense of movement is nearly opposite to that which occurred during the early fragmentation of the continent in the Mesozoic. Paleozoic, Precambrian, and Jurassic movements along the Ramapo fault system also were of a variety of types [*Ratcliffe*, 1971; *Dames and Moore*, 1977]. Thus this deep-seated zone of weakness appears to have been reactivated a number of times in response to the orientation of the particular stress system acting at a given time.

Aggarwal and Sykes [1978] used the rate of occurrence of small shocks within 10 km of the Ramapo fault as well as the occurrence of historic shocks thought to be associated with the fault to estimate return periods for earthquakes of various magnitudes. They calculate that an event of magnitude 5.1 (intensity VII) occurs about once per 100 years along the entire 120-km length of the fault. While that rate is much smaller than that of very active faults near many plate boundaries, it appears to be significantly high to be of concern in designing critical structures.

On land the recognition of possible Cenozoic fault movement is difficult, since the youngest sediments are of Triassic-Jurassic age with the exception of minor amounts of postglacial material. From bathymetric and seismic studies, however, *Sheridan and Knebel* [1976] report post-Pleistocene fault movement along several faults striking N70°E on the continental shelf south of New York City. Historic and instrumentally located earthquakes have occurred in that area.

The activity in southern New York and northern New Jersey may be part of a more extensive belt of earthquakes extending farther southwest that generally follows either grabens and their associated faults of Triassic-Jurassic age or older Paleozoic faults that may have been reactivated in the Triassic. Several major faults have been reported, or their presence inferred along the eastern side of the Piedmont near the fall line from New Jersey to Alabama. *Sbar et al.* [1975] obtained a focal mechanism solution for an earthquake near Wilmington, Delaware, that involved dip slip movement on a nearly vertical fault that strikes northeast such that the seaward block moved down in relation to the landward block. This activity occurred near the fall line and about 30 km northeast of a graben in the basement rocks, which is filled with Cretaceous sediments [*Spoljaric*, 1973]. A detailed study of one fault near the graben system suggests movement during the Late Cretaceous [*Spoljaric et al.*, 1976]. A number of earthquakes have occurred historically along the fall line between Wilmington, Delaware, and Trenton, New Jersey.

The graben mapped by *Spoljaric* [1973] in northern Delaware and the Wilmington earthquake are located nearly along the strike of a major fault system that *McGee* [1888], *Higgins et al.* [1974], and others speculate may extend from New York City to the northern Chesapeake Bay and thence to Fredericksburg, Virginia. They propose that this fault system, which is identified from the pattern of aeromagnetic anomalies in the upper Chesapeake Bay [*Higgins et al.*, 1974], could explain why several major rivers of the mid-Atlantic region suddenly change course from southeast to southwest at or near where they enter the coastal plain. Referring to work by *Jacobeen* [1972] and a report by Dames and Moore, *Hansen* [1976] cites evidence for coastal plain faulting of Tertiary age southeast of Washington, D. C., along the strike of this inferred fault system. From an integrated program of surface mapping, drilling, and seismic exploration, *Jacobeen* [1974] suggests that post-Eocene faulting has occurred in that area along the Brandywine fault system. This faulting appears to be high-angle reverse, up on the east, and has a throw of over 70 m. Using ERTS satellite photographs, *Withington* [1973] shows that the fault system originally mapped in Maryland by *Jacobeen* [1972] is at least 50 km long. Both Withington and *Jacobeen* [1974] conclude that movement may have occurred in the Holocene along this lineament and others recognized on ERTS images.

Mixon and Newell [1977] delineate a zone of northeast trending faults and flexures at least 56 km long, which they call the Stafford fault zone near the fall line in northeastern Virginia. Outcrops show northwest dipping, high-angle reverse faults along which crystalline rocks of the Piedmont are thrust over Cretaceous sediments. They conclude that faulting was active during deposition of Cretaceous sediments, that additional deformation occurred in late Eocene to early Miocene time, and that these movements reflect late Mesozoic and Cenozoic compressional stresses with movements of opposite sense to the normal faulting that characterized the area during the Triassic. As Triassic rocks have been encountered in several test borings, Mixon and Newell proposed that the Stafford and Brandywine fault system may lie along a zone of weakness marked by a buried Triassic fault system. They also cite evidence for minor flexuring of Miocene strata along the Brandywine fault system of Maryland. Small reverse faults offsetting stream terrace deposits near Rock Creek in Washington, D. C., are located nearly along the strike of the Stafford fault system and indicate compressional deformation as recent as Pliocene or Pleistocene [*Darton*, 1950; *York and Oliver*, 1976; *Mixon and Newell*, 1977].

Prowell [1976] finds that faults of northeast strike in the coastal region of the southeastern United States commonly show reverse displacement. *Talwani and Howell* [1976] find that a number of recent and historic earthquakes in South Carolina and adjacent parts of Georgia are located along the Goat Rock fault, which is nearly coincident with the fall line. That fault appears to be part of a much larger system of faults identified from field mapping and aeromagnetic data that extends along the eastern Piedmont from Alabama to Virginia [*Hatcher et al.*, 1977]. Like the Ramapo fault this fault system appears to have experienced several periods of deformation from the Paleozoic into at least the Mesozoic.

Fletcher and Sykes [1977] and *Sbar and Sykes* [1977] located several instrumentally rcorded earthquakes in western New York State along the Clarendon-Linden fault system. The 1929 Attica earthquake of maximum intensity VIII appears to have been located along the fault system, which experienced major displacement during the Paleozoic. Three fault plane solutions [*Fletcher and Sykes*, 1977; *Herrmann*, 1978] for earthquakes in the Attica-Dale area each indicate one nodal plane nearly parallel to the Clarendon-Linden fault. Thus at least a portion of that Paleozoic fault system is active today, judged from the distribution of earthquakes. Other shocks in westernmost New York and in Lake Ontario (Figure 17) are located west of the fault system.

Thus it seems likely that additional mapping aimed at detecting Cretaceous and Cenozoic faulting and other types of deformation will likely uncover many other examples than

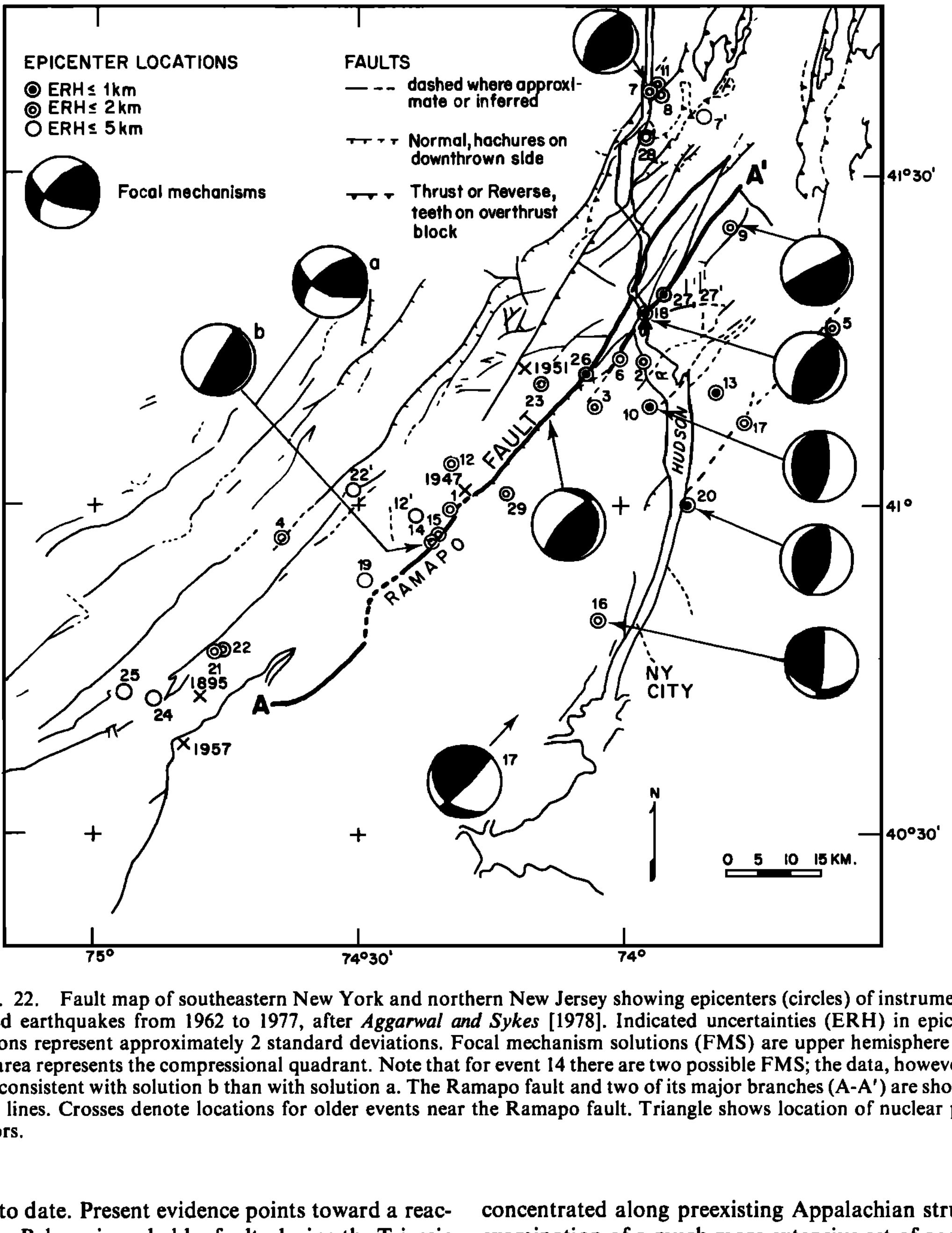

Fig. 22. Fault map of southeastern New York and northern New Jersey showing epicenters (circles) of instrumentally located earthquakes from 1962 to 1977, after *Aggarwal and Sykes* [1978]. Indicated uncertainties (ERH) in epicentral locations represent approximately 2 standard deviations. Focal mechanism solutions (FMS) are upper hemisphere plots; dark area represents the compressional quadrant. Note that for event 14 there are two possible FMS; the data, however, are more consistent with solution b than with solution a. The Ramapo fault and two of its major branches (A-A') are shown by heavy lines. Crosses denote locations for older events near the Ramapo fault. Triangle shows location of nuclear power reactors.

those reported to date. Present evidence points toward a reactivation of major Paleozoic and older faults during the Triassic as continental fragmentation began. Several of the classical explanations of the Triassic-Jurassic grabens of eastern North America portray them as a relatively minor disturbance involving the last dying gasp of Paleozoic orogeny and a relaxation of Paleozoic compressive stress. In a plate tectonic framework, however, it seems more reasonable to interpret these features as being related to the development of the Atlantic Ocean and to the Mesozoic evolution of the continental margins with time. Near the margin, deformation may have continued long after the initiation of continental separation, although it probably was at a maximum in the early stages of rifting. Thus some of the so-called Triassic-Jurassic faults in eastern North América still appear to be active today.

Cenozoic Uplift in Eastern North America and Relationship to Earthquakes

Bollinger [1973] infers that seismic activity in the southern Appalachians may be caused by crustal uplift, strains being concentrated along preexisting Appalachian structures. In an examination of a much more extensive set of geodetic leveling measurements for the eastern United States, *Brown and Oliver* [1976] conclude that the Appalachians are rising at rates of up to 6 mm/yr and that the Atlantic and Gulf coastal plains are tilting downward away from the continental interior. The sense of the modern movements is generally the same as that of movements during the last 130 m.y. The modern rate of vertical movement, however, is much higher than the rate for the longer interval, which suggests that the recent movements are either episodic or oscillatory. Although Brown and Oliver find instances in which modern seismicity appears to correlate with extremes of vertical movements in the Appalachians, they conclude that with present data it is impossible to demonstrate that the relationship is more than coincidental.

From repeated geodetic surveys about 25 years apart, *Isachsen* [1975, 1976] finds that the Adirondack region of northern New York State (Figure 24) is being uplifted at rates as high as 3.7 mm/yr. The region of uplift as inferred from the two geodetic lines that he analyzed and from the topography and geomorphology approximately coincides with the region of

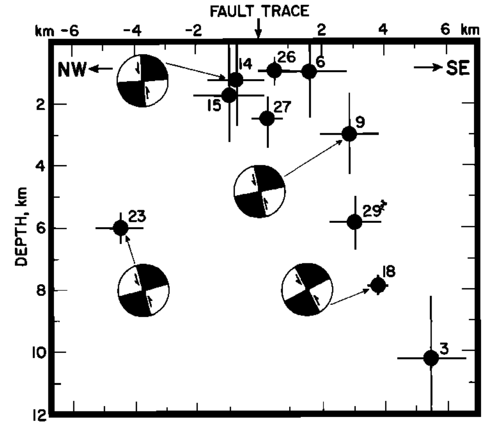

Fig. 23. Composite (stacked) vertical cross section showing focal depths and focal mechanism solutions (FMS) (dark area is the compressional quadrant) for events within 10 km of the Ramapo fault trace, after *Aggarwal and Sykes* [1978]. Event number is keyed to epicenter number in Figure 22; only those events are plotted for which reliable focal depths could be determined. Bars represent 1 standard deviation. Northeast of epicenter 26 (Figure 22), horizontal distance is measured from one of two major branches of fault on the basis of FMS.

high seismic activity in the Adirondacks (Figure 17). Nevertheless, high activity extends to the northwest beyond the region of crustal uplift and into a region of low topography along the St. Lawrence Valley. Also, it does not seem likely that this localized region of uplift, which is about 200 km in diameter, is related to the removal of Pleistocene ice.

Fairbridge [1976] mapped the present elevation of a prominent Late Cretaceous planation (peneplain) surface in the eastern and central United States (Figure 25). Some of the largest elevations are found along the Appalachians, with maxima centered in New England and in the southern Appalachians. The planation surface has been tilted seaward and downdropped more than 1000 m below present sea level off the east coast and along the Gulf coast. This pattern is very similar to that obtained by leveling over the last 100 years [*Brown and Oliver,* 1976].

Fairbridge [1976] concludes that modern seismicity appears to correlate with inflections in the elevation of the Late Cretaceous surface, which flank the Appalachians. With the exception of seismic activity near Charleston, South Carolina, most of the earthquakes in Figure 14 are located west of either the 0- or 250-m contours of Figure 25. The zone of seismic activity that appears to extend from the Mississippi Embayment across Illinois, Indiana, and Ohio to Lake Erie (Figures 1 and 14) is located near the 250- and 500-m contours on the less well defined western flank of the Appalachian high.

The pattern of seismicity in Figure 14, however, does not seem to be related in a simple way to uplift or to the gradient of vertical motion. The Blue Ridge dome of Figure 25 is roughly delimited on the northeast and southwest by the central Virginia and South Carolina seismic zones, respectively (and by transform directions extended into the continent from the ends of the Norfolk and Blake fracture zones, respectively (Figure 15)). It is located to the west of a zone of low historic seismicity in eastern and central North Carolina (Figure 15). Thus the dome may be related to the block structure of the continental margin. Nevertheless, the presence of seismic activity in the Appalachians and in eastern Australia, both of which have undergone recent uplift, suggests some relationship between the occurrence of earthquakes and either uplift, differential vertical motion, compression of the youngest orogenic belt in an area, proximity to a continental margin that has opened by continental drift, or abrupt changes in the thickness of the lithosphere. These possible relationships will be explored further in a later section.

A number of workers have suggested that stresses induced by the removal of Pleistocene ice cause earthquakes in eastern North America. The pattern of seismicity, however, is not well correlated with either the centers of Pleistocene glaciation or their peripheral bulges. The maximum compressive stress as inferred from focal mechanism solutions, hydrofracture measurements (Figure 26), and postglacial geologic features is oriented east to ENE and is nearly horizontal over a broad area including northern and western New York State, Ohio, Illinois, adjacent parts of Canada, and perhaps other parts of the central United States [*Sbar and Sykes,* 1973, 1977; *Yang and Aggarwal,* 1977]. These directions appear to bear no relationship to the pattern of glacial rebound. Also, the similarities in the pattern of seismicity between eastern North America and eastern Australia, which was only glaciated in a few locations during the Pleistocene [*David,* 1950], indicate that crustal stresses related to ice removal are not the principal cause of earthquakes in eastern North America.

New Madrid Zone and Other Earthquakes in Central United States

New Madrid zone. The most seismically active region in the central and eastern United States is located along the Mississippi River in southeastern Missouri and in the adjoining parts of Arkansas, Illinois, Kentucky, and Tennessee (Figure 1). This region contains the epicenters of the large New Madrid earthquakes of 1811 and 1812, has exhibited continuing activity [*Nuttli,* 1973*a, b*] throughout the historic and instrumental record (Figures 12, 14, 16, and 27), and accounts for about a third of the earthquakes in the central United States [*Nuttli,* 1978]. This activity is situated along the northern end of the Mississippi Embayment (Figures 25 and 28), a NNE to northeast tectonic feature extending from the coastal plain of the Gulf of Mexico into the midcontinent region. The embayment has undergone extensive subsidence during the Cretaceous and Cenozoic [*Stearns and Wilson,* 1972; *Ervin and McGinnis,* 1975]. The term New Madrid fault system is used to describe the system of faults trending northeast to NNE of Cretaceous and Cenozoic age within and surrounding the Mississippi Embayment (Figure 27). Many of the epicenters of earthquakes in the region from 1961 to 1974, which are shown in Figure 27, define a northeast trending zone along the northern portion of the embayment. Other earthquakes in Figure 27, however, appear to be associated with northwest striking faults such as the Ste. Genevieve fault.

The presence of thick sediments in the Mississippi Embayment has made it difficult to map faults in the basement and to ascertain the relationship between Cenozoic faulting and earthquakes. Epicenters located by a network of seismographs installed around the New Madrid region in 1974 show a tighter spatial clustering than the shocks in Figure 27. The newer data indicate narrow zones of activity A-B and C-D (Figure 27) of northeasterly strike parallel to the Mississippi Embayment connected by a zone B-C striking northwest [*Stauder et al.,* 1976]. The segment C-D is nearly coincident with *M. L. Fuller*'s [1912] inferred zone of faulting for the shocks of 1811 and 1812 (Figure 27). From subsurface geophysical and geological data, *Mateker* [1968] infers a pattern of northeast striking

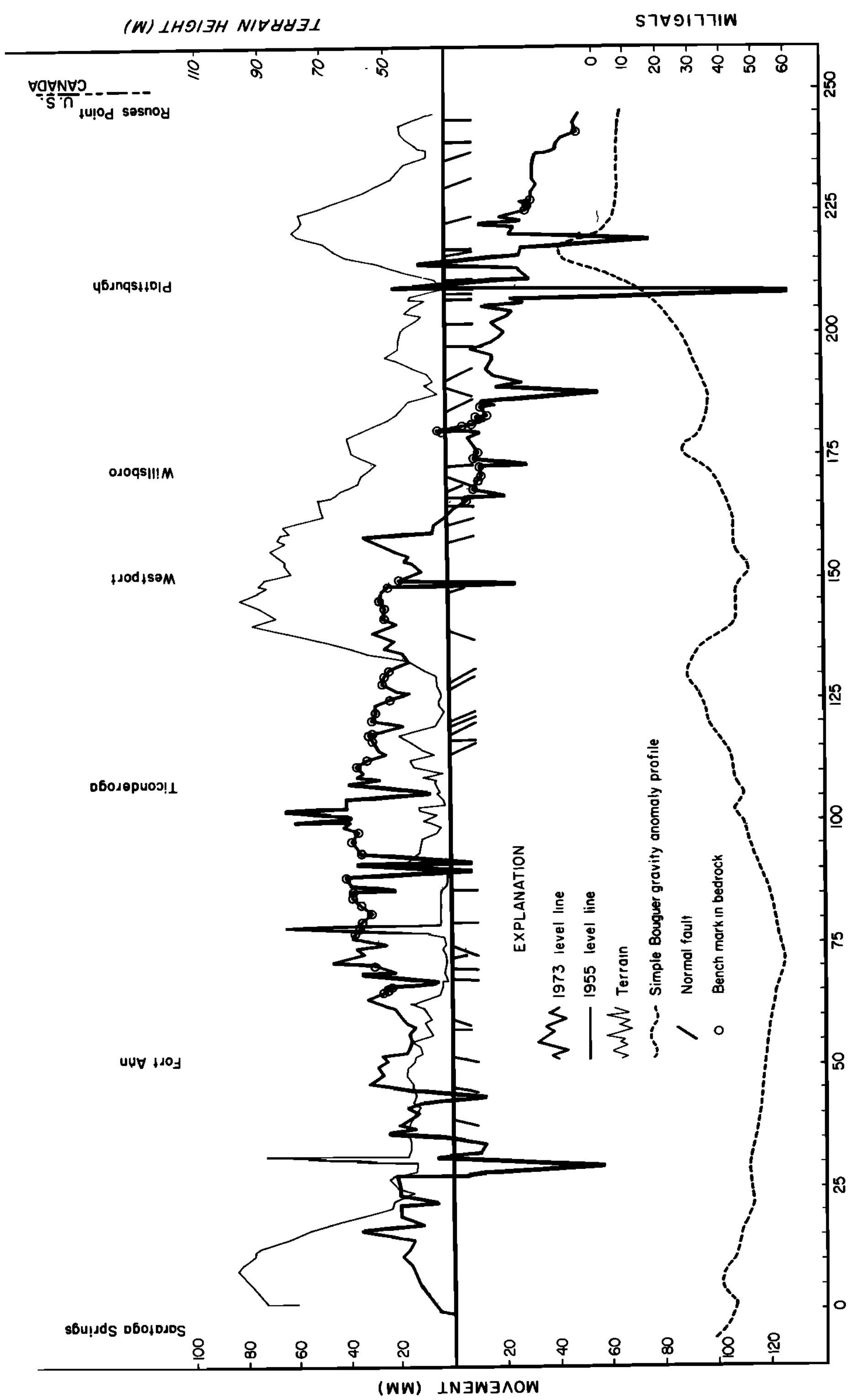

Fig. 24. Geodetic leveling profile along the eastern flank of the Adirondack dome for the period 1955–1973, after *Isachsen* [1975]. Note that the region of maximum uplift is rising at a rate of about 2 mm/yr relative to Saratoga Springs (and to sea level at New York City) and 4 mm/yr relative to the United States–Canada border. Another profile across the center of the Adirondacks indicates uplift of about 3.7 mm/yr relative to Utica, New York, in the Mohawk Valley [*Isachsen*, 1976].

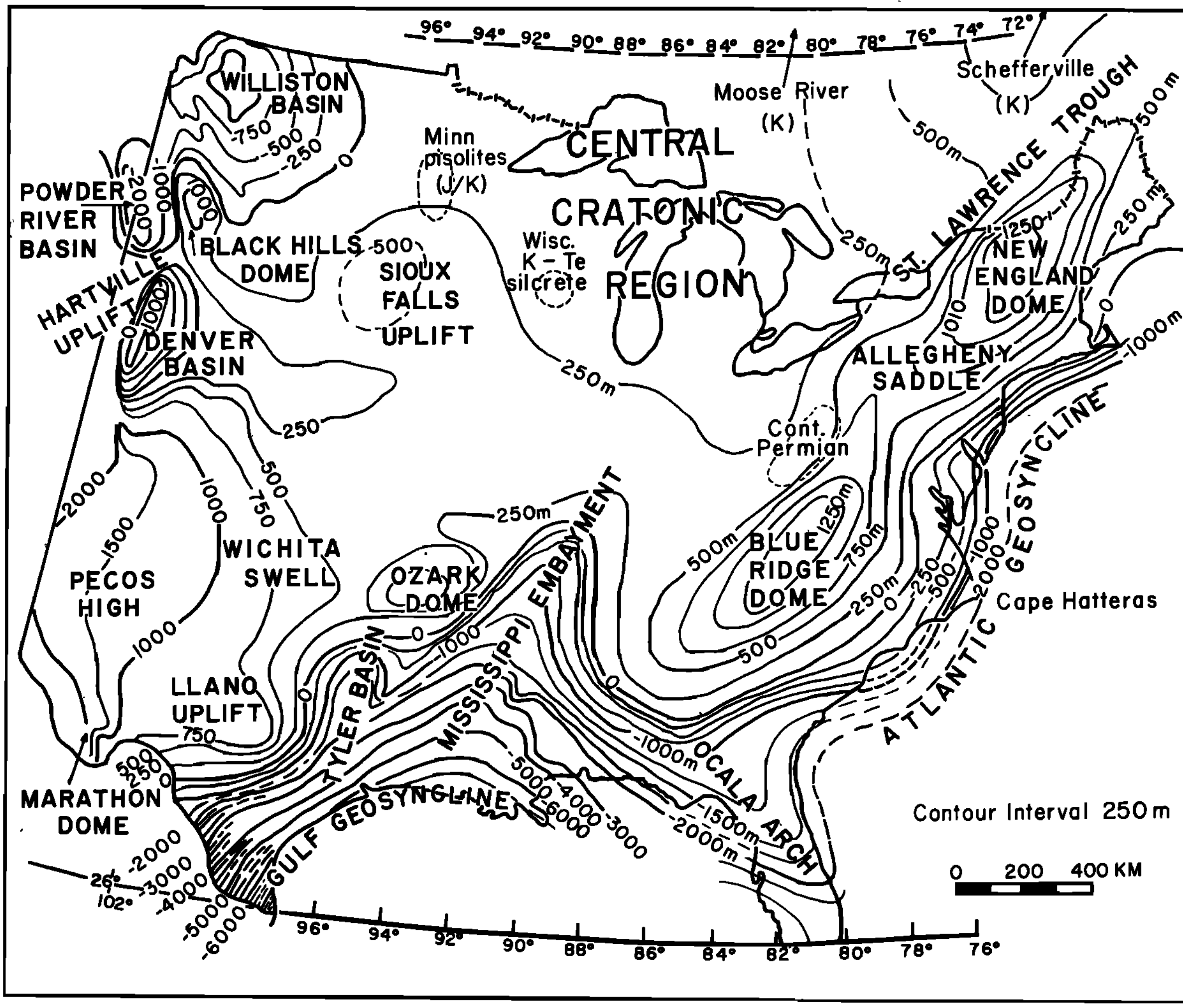

Fig. 25. Present elevation of Late Cretaceous planation surface in eastern and central North America, after R. W. Fairbridge (personal communication, 1977). Note regions of domal uplift in the Appalachian region.

faults within the northern Mississippi Embayment. Focal depths for recent earthquakes range from about 5 to 20 km [*Nuttli,* 1978].

Movement along several of the faults around the head of the Mississippi Embayment (Figure 27) is dated quite precisely by standard stratigraphic methods [*Stearns and Wilson,* 1972]. Faults offset loess of Pleistocene age and alluvium along the Mississippi River from southern Illinois to central Arkansas. Faulting along the southeastern edge of the embayment becomes progressively older proceeding southeast. Although mapping to date shows fewer than 25 faults than offset Cretaceous and younger strata in the northern part of the Mississippi Embayment, *Stearns and Zurawski* [1976] interpreted 70 hypothetical faults in the area by using topographic lineations and well data. They conclude that this number may be even greater if known faults that cut the Cretaceous along the northern perimeter of the embayment persist southwestward. They find a pattern of dominant northeast trending faults and subordinate northwest trending faults much like the pattern in Figure 27. Thus it is reasonable to conclude that earthquakes in this region, including the large shocks of 1811 and 1812, are associated with movements along faults, many of which strike northeast to NNE, within and near the Mississippi Embayment.

Young igneous rocks. A number of lamprophyres, syenites, and other alkalic rocks encountered in boreholes within the Mississippi Embayment (Figure 28) are of Late Cretaceous age [*Kidwell,* 1951; *Stearns and Wilson,* 1972; *Ervin and McGinnis,* 1975; *Sundeen and Cook,* 1977]. *Mateker et al.* [1965] report three large magnetic and gravity anomalies aligned north-south within the Mississippi Embayment near New Madrid. They deduce that these represent basic intrusions into the upper part of the crust. Much of the Mississippi Embayment is characterized by positive gravity anomalies (when a correction is made to the Bouguer anomaly for the effects of low-density sediments) [*Ervin and McGinnis,* 1975; *Cordell,* 1977]. This anomaly is indicative of generally higher densities in the crust and uppermost mantle of that region.

Zartman et al. [1967] obtained radiometric ages of 88–97 m.y. on nepheline syenites near Little Rock, Arkansas, and on alkalic igneous rocks from Magnet Cove, Arkansas, respectively (Figure 28). These and other igneous rocks of Late Cretaceous age in central and southern Arkansas are located near the western side of the Mississippi Embayment. Diamond-bearing peridotites from Murfeesboro, Arkansas, and water-laid volcanic debris in southwestern Arkansas and adjacent parts of Oklahoma and Texas are also indicative of igneous activity of Late Cretaceous age [*Miser and Ross,* 1925; *Zartman et al.,* 1967]. Both the gravity data and the known occurrences of intrusive rocks indicate that most or all of the large Cretaceous bodies were emplaced south of the head of the embayment [*Ervin and McGinnis,* 1975]. As in New Eng-

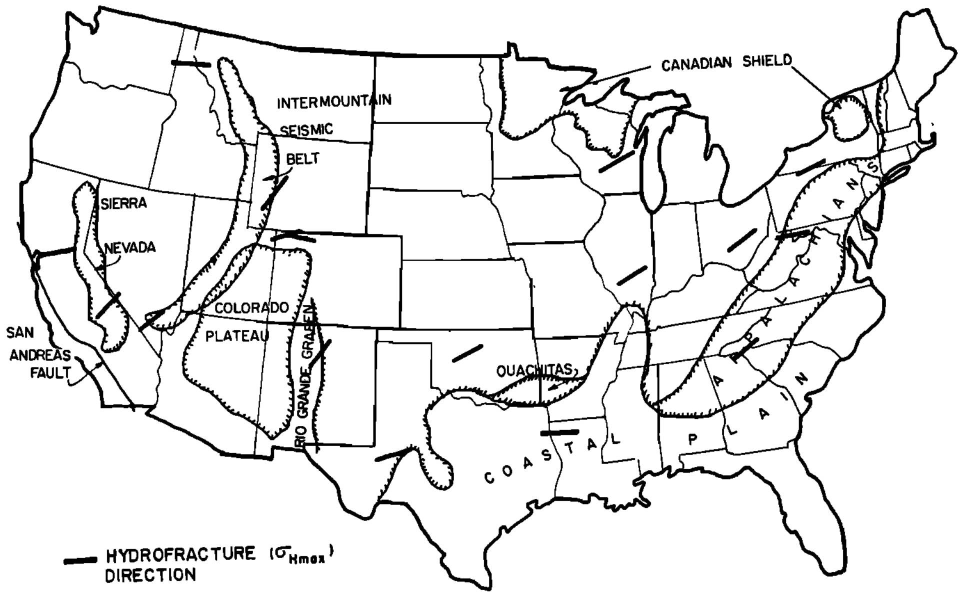

Fig. 26. Orientation of maximum horizontal compressive stress in the United States as determined from hydrofracturing measurements, after *Haimson* [1976].

land, Virginia, and South Carolina, alkalic igneous rocks of late Mesozoic age are commonly found along the Mississippi Embayment. *Kidwell* [1951] concludes that in detail the alkaline rocks of the embayment tend to be concentrated near the intersection of conjugate structural systems (such as the northeast trending embayment structures and the southeast trending continuation of the Ouachita fold belt).

Alkaline rocks and diatremes occur at a number of localities in the midcontinent region of the United States. Mica peridotites from the Silver City and Rose domes (Figure 28) in eastern Kansas (about 500 km west of the Mississippi Embayment) were dated by *Zartman et al.* [1967] as 90 m.y., i.e., about the same age as the Cretaceous igneous rocks in Arkansas. Several of the occurrences of alkaline rocks outside the embayment, however, have much older radiometric ages than Mesozoic. For example, diatremes and mafic dikes found in southeastern Missouri near Avon [*Rust,* 1937] and at the Hicks dome (HD in Figure 27) in southern Illinois [*Heyl et al.,* 1965] are dated by *Zartman et al.* [1967] as Devonian and Permian, respectively. These two structures, however, may be associated with a Precambrian rift zone that was the site of deep-seated igneous activity during the Paleozoic and that was reactivated as the Mississippi Embayment in the Cretaceous [*Ervin and McGinnis,* 1975]. A number of cryptovolcanic features in the midcontinent region, which *Bucher* [1963] and *Snyder* [1968] believed to be of internal origin, are now thought to be of impact (external) origin.

Focal mechanism solutions and the origin of the Mississippi Embayment. *Street et al.* [1974] obtained a number of focal mechanism solutions for earthquakes in and near the New Madrid seismic zone. In their study, shocks within the embayment (south of 36.3°N) are generally characterized by normal faulting, which is in agreement with the geological evidence for subsidence during the Cretaceous and Cenozoic. Shocks surrounding the embayment tend to exhibit thrust faulting, although scatter is evident in the types of mechanisms. R. B. Herrmann (written communication, 1977), however, has revised their solutions for shocks within the embayment and finds right lateral strike slip motion with a component of thrust faulting on planes striking N40°E and dipping 70°NW.

Historic earthquakes are not found along the entire Mississippi Embayment but are concentrated mostly in a 200-km-long zone near the head of the embayment. A possible explanation for this concentration of activity is that the embayment may be modeled as a slowly propagating crack (or rift zone) surrounded by relatively thick and old lithosphere, which is under horizontal compression. Other explanations for the concentration of activity near New Madrid generally relate the earthquakes spatially to the Pascola arch, a northwest trending Paleozoic structure connecting the Nashville and Ozark domes (Figure 25). This arch appears to have subsided as the Mississippi Embayment developed in the Cretaceous [*Ervin and McGinnis,* 1975] and may serve to concentrate tectonic stress. Another possible explanation is that the short historic record is not typical of long-term movements, which may still be in progress farther to the south along the embayment.

Of all the regions of Mesozoic or more recent alkaline magmatism in eastern and central North America, the Mississippi Embayment appears to be most similar to the East African rift valleys and other rift valleys in that it has undergone subsidence, it was the locus of alkaline magmatism, and it is the site of high seismic activity. *Ervin and McGinnis* [1975] mention that a thick lower crustal layer of velocity 7.4 km/s detected on a refraction profile by *McCamy and Meyer* [1966] near the western edge of the embayment is similar to that found in rift zones elsewhere in the world. From a study of travel time residuals, *Mitchell et al.* [1977] conclude that the northern end of the Mississippi Embayment is characterized by anomalously low seismic velocities in either the lower crust, upper mantle, or both.

Burke and Dewey [1973] interpret the embayment as a rift or failed arm about an RRR triple junction near Jackson, Mississippi. In their hypothesis the other two rifts continued to spread and developed into the Gulf of Mexico. One other possibility is that the embayment is located along an old zone of weakness that controlled the location of a major fracture zone farther to the south that was active in the opening of the Gulf of Mexico [*Fletcher et al.,* 1978]. The latter idea may not be as improbable as it may appear, since oceanic crust of the

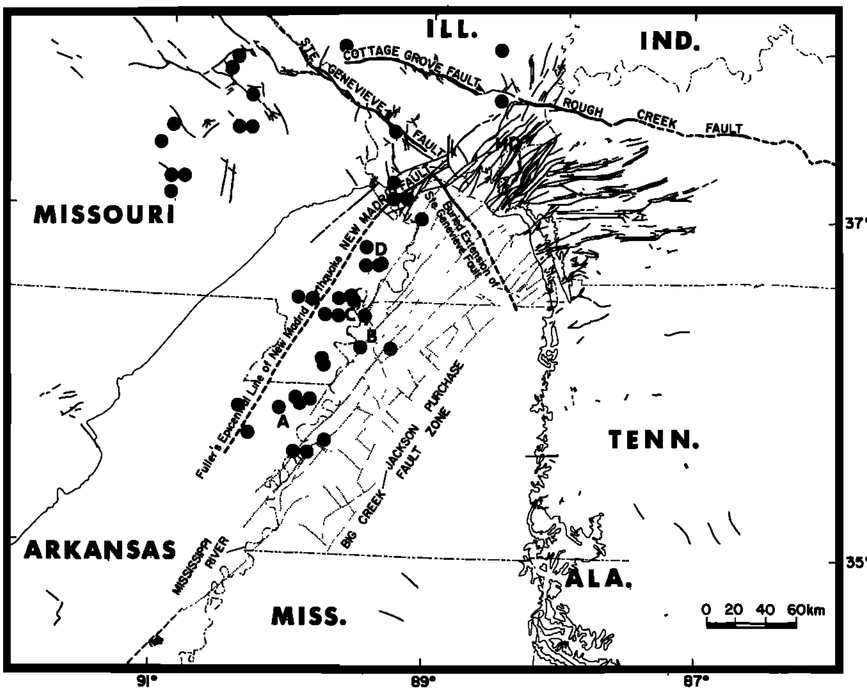

Fig. 27. Epicenters and faults along the New Madrid seismic zone in the central United States, after *Fletcher et al.* [1978]. Faults are from *Stearns and Wilson* [1972]. Many of the faults in the Mississippi Embayment are inferred and are shown as dotted lines. Solid circles are earthquakes from the earthquake data report of the U.S. Geological Survey, 1961–1974. Note the northeast trending zone of earthquakes along the Mississippi Embayment near *Fuller*'s [1912] epicentral line of New Madrid earthquakes of 1811–1812. Segments A-B, B-C, and C-D are zones of high activity defined by *Stauder et al.* [1976] using a recently installed seismic network. HD is Hicks Dome.

Gulf of Mexico probably extends much farther north than the present coastline. A preexisting rift zone, possibly an aulacogen, appears to have been reactivated to form the present embayment [*Ervin and McGinnis,* 1975]. It is difficult, however, to choose between the above two hypotheses. Nevertheless, seismic activity near the head of the embayment clearly does not occur at random in space but is somehow related to the development of the embayment and to the emplacement of alkaline igneous rocks.

Northern extension of the New Madrid zone. Faults of similar strike and age to those in the Mississippi Embayment appear to continue northeast into the Wabash fault system of southern Illinois and Indiana [*Heyl et al.,* 1965], where a number of historic shocks have occurred (Figures 1, 12, 14, and 16). Nevertheless, although seismic activity in the two areas seems to be related, the narrow zone of activity A-B-C-D (Figure 27) delineated by using data from a newly installed network of seismic stations [*Stauder et al.,* 1976] does not appear to continue north of the embayment. Since that data set covers a time period of only a few years, however, it is difficult to tell if the two tectonic and seismic zones are, in fact, distinct. *Nuttli* [1978] believes that the Wabash zone is about 200 km long and has the greatest potential for a future large shock, second only to the New Madrid seismic zone. Shocks as deep as 20 km have been reported from this zone. *McGinnis and Ervin* [1974] conclude that earthquakes in southern Illinois tend to be concentrated along the edges of blocks of dimensions tens of kilometers to about 100 km which are relatively undeformed internally.

It is possible that the activity in the Wabash River valley may be continuous with that farther northeast near Anna, Ohio, Lake Erie, and western New York (Figures 1, 13, and 14), although special studies are needed to ascertain if activity in those areas is, in fact, continuous or if it occurs in isolated spots. A boundary between basement of Grenville age (800–1100 m.y.) and that of Ozarkian age (1200–1450 m.y.) appears to pass near regions of historic earthquakes in southeastern Missouri, southern Illinois, and western Ohio [*Kanasewich,* 1965; *P. B. King,* 1970]. Hence seismic activity in those areas may be located on or near a major tectonic boundary of Precambrian age. That boundary may have been reactivated during the opening of the Gulf of Mexico and the North Atlantic.

Other parts of the central United States. As *Docekal* [1970] and *Nuttli* [1973*b*, 1978] emphasize, several large historic shocks have occurred in the midcontinent region of the United States outside of the New Madrid seismic zone (Figures 1, 12, 14, and 16). *Nuttli* [1973*b*, 1978] defines an Ouachita-Wichita mountain seismic zone about 150 km wide and over 1000 km long that extends from the Texas panhandle through southern Oklahoma and central Arkansas into northern Mississippi. He states that fault plane solutions indicate motion along the strike of the mountain front. Thus, as in the Appalachians, earthquakes seem to be occurring along a Paleozoic orogenic belt. This belt apparently was reactivated during the development of the Gulf of Mexico and the Mississippi Embayment in the Triassic and Jurassic. As in the Appalachians, the Late Cretaceous planation surface is domed upward (the Ozark dome of Figure 25) along a part of that Paleozoic fold belt.

Docekal [1970] and *Nuttli* [1978] conclude that a belt of earthquakes extending from Oklahoma to eastern Nebraska is associated with the Nemaha uplift and the midcontinent gravity high. It is possible that some activity may extend northward from the Nemaha uplift along the midcontinent gravity high

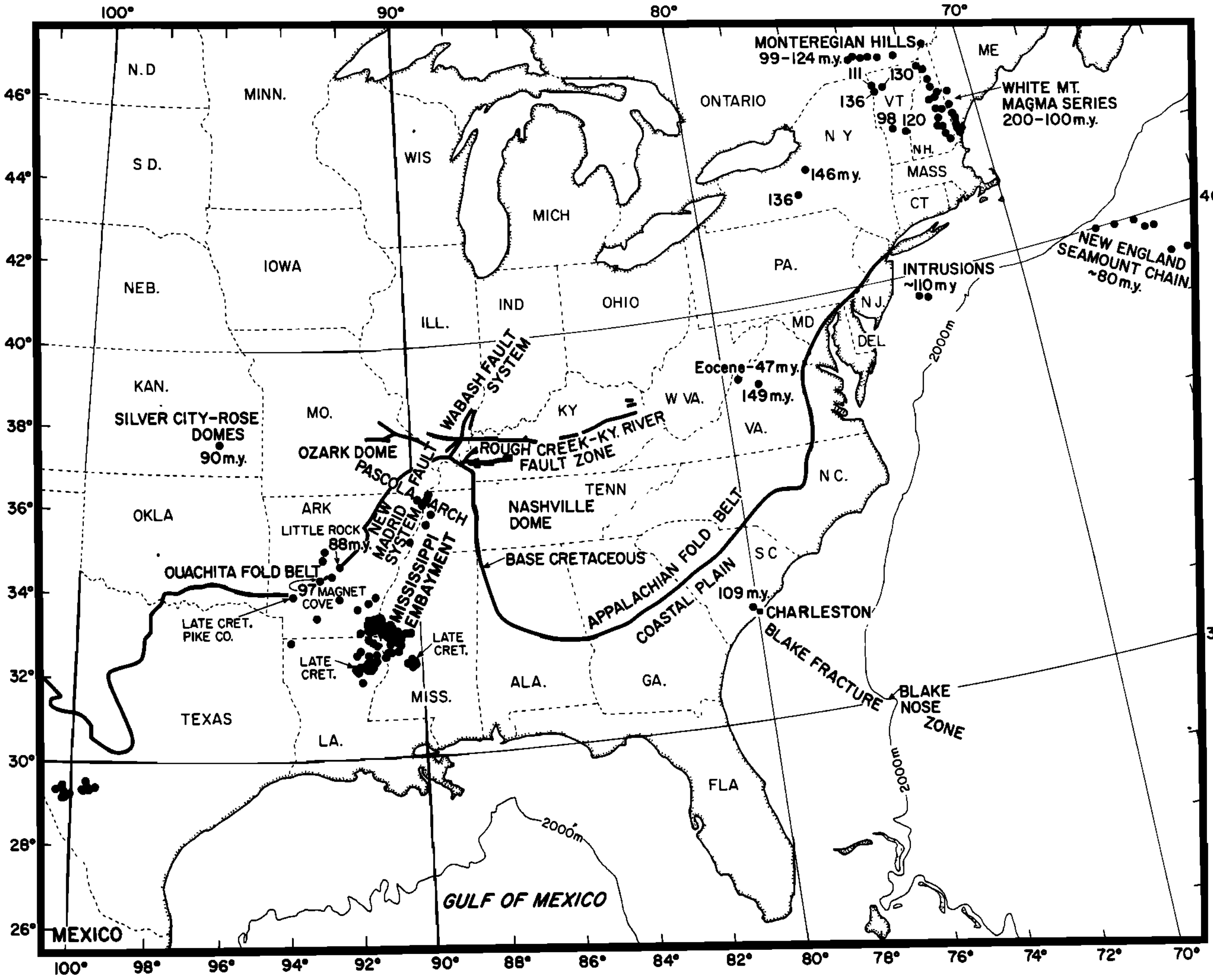

Fig. 28. Alkaline igneous rocks of Jurassic and younger age in the central and eastern United States and in southeastern Canada. Data and ages in millions of years are from *Kidwell* [1951], *Zartman et al.* [1967], *Dennison and Johnson* [1971], *Ervin and McGinnis* [1975], *Deep Sea Drilling Project* [1975], *McHone et al.* [1976], *Fletcher et al.* [1978], *Foland and Faul* [1977], and *Mattick* [1977].

(which appears to mark a late Precambrian rift zone) into Minnesota. At present, however, very little is known about the relationship of seismic activity in the Great Plains to tectonic features. This situation may improve as better epicentral locations and focal mechanism solutions become available and as old suture zones and other zones of deformation become better delineated.

Other Intraplate Regions

Canada and Greenland

Since seismograph stations were installed in northern Canada in the early 1960's, a large number of small and moderate size shocks have been detected near the Arctic continental margin of northern Canada, in the Mackenzie Valley of northwestern Canada, in Baffin Bay, and in the Labrador Sea (Figure 20). A number of larger shocks from these areas were also recorded teleseismically prior to 1960 [*Gutenberg and Richter*, 1954; *Sykes*, 1965; *Rothé* 1969; *Basham et al.*, 1977]. Since Figure 20 is a compilation of all shocks in the Canadian files up to 1974, the detectability of events varies considerably in both space and time. Hence care should be taken in comparing the levels of activity in different regions. Nevertheless, large parts of the Canadian shield, which is now well covered by stations, exhibit extremely low activity, while many of the areas marginal to the shield are moderately active.

Basham et al. [1977] conclude that activity in northern Canada tends to be associated with Paleozoic or older structures that have been reactivated in response to the present stress field. High activity in the Mackenzie Valley of northwestern Canada coincides with the most severely faulted part of that area; several zones of earthquakes in the Canadian Arctic coincide with Paleozoic structures. Bathymetric data indicate a series of horst and graben structures extending from northwest of Southampton Island (northwestern Hudson Bay) to the eastern end of Hudson Strait (Figure 20). These structures indicate tectonic activity at least during the mid-Paleozoic [*Basham et al.*, 1977]. Seismic activity in Figure 20 is also concentrated along those structures and is much higher than that of the surrounding areas of the Canadian shield. Since seismic activity in Hudson Strait may be continuous with that

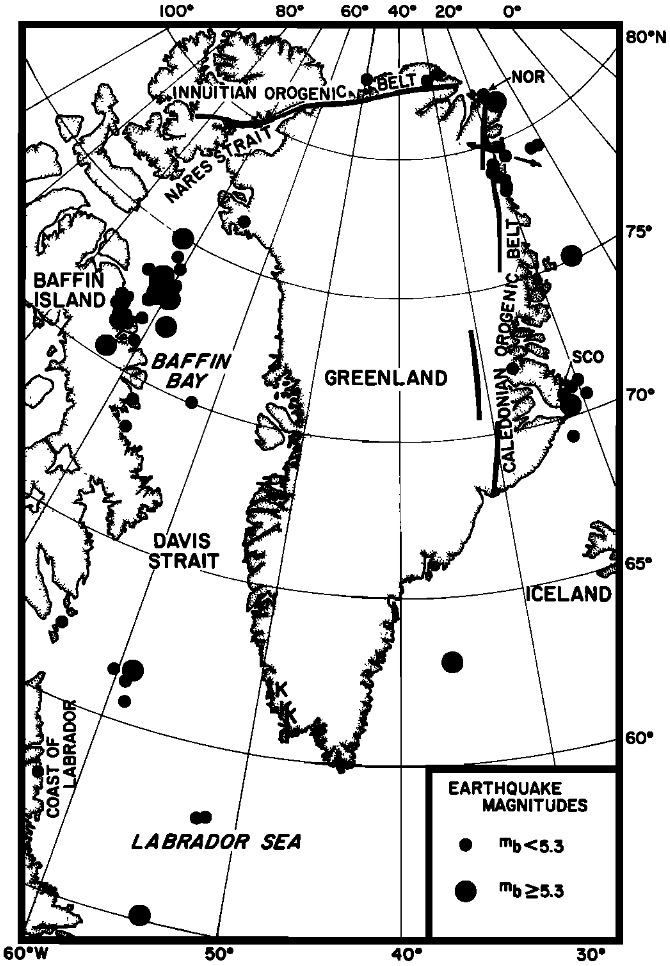

Fig. 29. Earthquakes in and near Greenland, 1904–1975. Data are from *Gutenberg and Richter* [1954], 1904–1952 ($m_b \geq 5.3$); *Sykes* [1965], 1955–1964; *Rothé* [1969], 1953–1965; and earthquake data report of the U.S. Geological Survey, 1962–1975. Note that earthquakes are concentrated in Paloezoic fold belts along the northeast and north coasts. Arrows indicate horizontal component of axis of least compressive stress (T axis) as inferred from the focal mechanism of shock near the northeast coast of Greenland by *Sykes and Sbar* [1974]. SCO and NOR are seismograph stations Scoresby Sund and Nord. Boundaries of fold belts are from *Haller and Kulp* [1962]. Note that the data set probably is not complete, especially for smaller shocks, before 1955.

in the Labrador Sea, the former areas may have been reactivated during the opening of the Labrador Sea in the Cretaceous.

A number of epicenters in Figure 20 are situated off the Arctic coast of northern Canada along the continental margin. *Herron et al.* [1974] identify this margin as an active plate boundary along which sea floor spreading occurred during the Jurassic and strike slip motion took place from about 81 to 63 m.y. ago. Earthquake activity indicates that this margin and several of the Paleozoic and Mesozoic terrains adjacent to it are active sites of intraplate tectonism.

One of the most active regions of intraplate activity in North America is situated in Baffin Bay and along the east coast of Baffin Island (Figures 20 and 29). Baffin Bay and the Labrador Sea (the latter of which is a region of moderate activity in Figures 20 and 29), were sites of active sea floor spreading from about 60 to 40 m.y. ago, when Greenland drifted away from North America. Although areas near former spreading centers in the Labrador Sea and in Baffin Bay are still seismically active, very few earthquakes are found along what appear to have been old transform faults along the Nares and Davis straits.

Although the interior of Greenland, which is underlain by Precambrian rocks and is covered by an ice sheet, is nearly aseismic (Figures 20 and 29), a number of earthquakes have been detected along the northeastern and northern coasts. Early Paleozoic (Caledonian) structures along the east coast of Greenland north of 70°N apparently were reactivated during the Cenozoic opening of the northernmost Atlantic [*Haller and Kulp,* 1962] and appear to be moderately active today judging from the distribution of epicenters. Earthquake epicenters in Figures 20 and 29 and the presence of many small local earthquakes on the records of the seismograph station NOR in northeastern Greenland [*Husebye et al.,* 1975] indicate that the northeastern tip of Greenland is one of the most active areas in Greenland. Northeast Greenland is situated very close to the intersection of the large Spitsbergen (Nansen) fracture zone and the Nansen ridge, the latter of which is the active spreading center in the Eurasian basin of the Arctic Ocean. Hence activity in northeastern Greenland appears to be related either to the reactivation of Caledonian structures or to the proximity of the area to the end of the Spitsbergen fracture zone.

Many small local earthquakes are also present on the seismograms for the station SCO (later KTG) near Scoresby Sund in eastern Greenland [*Husebye et al.,* 1975]. Scoresby Sund appears to mark a major transverse boundary in the present-day topography, in the Caledonia fold belt, and in the distribution of Tertiary flood basalts (most of which are found south of that latitude). Hot springs are also found in the region [*Wager,* 1939]. A nearly 200-km offset of the continental margin near Scoresby Sund appears to match a similar offset where the Jan Mayen fracture zone intersects the Norwegian margin along the south side of the Vøring plateau [*Talwani and Eldholm,* 1974]. The highest activity in Norway is found near the end of the Jan Mayen fracture zone [*Husebeye et al.,* 1975].

Seismic activity along the north coast of Greenland and in Baffin Bay (Figures 20 and 29) may also represent continued movement on faults that were active during either the opening of Baffin Bay and the Labrador Sea or the opening of the Arctic basins. *Herron et al.* [1974] propose that active plate boundaries passed through both areas during the Tertiary. *Andrews and Emeleus* [1975] and *Emeleus and Andrews* [1975] describe three kimberlite intrusions (Figure 29) in southwest Greenland that appear to have been emplaced during the early stages of rifting between Canada and Greenland.

Antarctica

Antarctica is clearly the most aseismic continent. No earthquakes have been detected from that continent except for a few very small events related to the movement of ice and to volcanism. Although seismic coverage of Antarctica was poor prior to about 1957, since then a number of very sensitive stations have provided detectability down to magnitude 4.5–5.0 for the entire continent. Hence the lack of detectable shocks since 1957 cannot be ascribed to poor coverage. The absence of earthquakes appears to hold not only for Antarctica itself but also for its continental margins and for much of the surrounding sea floor.

Nevertheless, parts of Antarctica were affected by Mesozoic tectonism, and all or most of its continental margins developed in either the Mesozoic or the Cenozoic as other blocks moved away by continental drift and sea floor spreading. The aseismic nature of Antarctica and its continental margins indicate that the presence of relatively young continental margins (which presumably are still sinking) or the presence of older zones of weakness in the continent is not sufficient in itself to localize detectable earthquakes.

It seems unlikely that the load of the antarctic ice sheet would impede the occurrence of earthquakes unless the state of stress was already low or nearly hydrostatic. In Greenland, unlike Antarctica, earthquakes do occur along the continental margins and around the margins of the large ice sheet. It is not unreasonable that tectonic stresses are not transmitted into Antarctica from surrounding plates, since nearly all of the plate boundaries of the antarctic plate are of the spreading type. Conceivably, Antarctica may be moving very slowly with respect to the deep mantle. Thus it is possible that significant stresses are not being transmitted into the antarctic plate either by tractions imposed from below or by forces along the plate boundaries. This has important implications for other intraplate regions that are seismically active, since contemporary stresses generated by forces either on the edges or on the bottom of the plate appear to be a necessary ingredient for intraplate seismicity.

Forsyth [1973] obtained a focal mechanism solution that involves thrust faulting for an earthquake of magnitude 6.2 in the South Pacific during 1971. The shock was located in the antarctic plate about 500 km east of the East Pacific rise and about 500 km south of the Chile ridge. Clearly, stress differences in that part of the oceanic lithosphere are high. These stresses, however, may have been generated as a thermoelastic stress by the cooling of the oceanic lithosphere [*Turcotte*, 1974] or may be related to the proximity of the shock to two active ridge systems. The lack of earthquakes throughout most of the antarctic plate argues that large stress differences sufficient to generate earthquakes probably are not present in much of that plate.

Western and Northern Europe

Earthquakes in western and northern Europe from 1901 to 1955 are shown in Figure 30. A broad zone of seismic activity in southern Europe and northern Africa (largely omitted from Figure 30) appears to mark the boundary between the Eurasian and the African plate. Seismic activity in the region just to the north of the Alps probably is related to compressional stresses along the plate boundary, since measurements of stress in situ show a systematic decrease in the size of the maximum horizontal compressive stress in passing from the Alps northward into Germany [*Greiner and Illies*, 1977]. *Ahorner* [1975] finds that the direction of maximum horizontal compression as inferred from focal mechanism solutions is oriented consistently northwest in the western Alps, the Rhine graben, and in the lower Rhine Valley. Thus earthquakes in Europe south of about 50°N probably should be regarded as plate-boundary-related rather than strictly as intraplate events. Ahorner's focal mechanism solutions also indicate that the Rhine graben is presently the site of strike slip faulting. Thus it is not undergoing simple extension at the present time.

Focal mechanism solutions are not available to my knowledge for any earthquakes in Britain or Fennoscandia. Hence it is difficult to ascertain if events in those areas represent intraplate or plate-boundary-related earthquakes. Many of the historic and instrumentally recorded shocks in Scotland (Figures 30 and 31) appear to be related to the Great Glen and Highland Boundary faults [*Kennedy*, 1946; *Gutenberg and Richter*, 1954; *Crampin et al.*, 1972; *Lilwall*, 1976]. These and other major northeast striking faults, which had major movement during the Paleozoic, appear to have been reactivated in the early Cenozoic during the separation of Greenland from Europe. Earthquake activity indicates that those faults are still moderately active today. In Figure 31 a northeast trending lineation, possibly a fault zone that extends from Wales across central England and thence to the area of two large shocks in the North Sea, appears to be defined by the presence of numerous shocks on its northwest side and by a near absence of reported activity on its southeast side.

The distribution of earthquakes in Fennoscandia from 1901 to 1955 in Figure 30 is very similar to that deduced by *Bungum and Husebye* [1978] mainly from felt reports of earthquakes from 1497 to 1973. The highest activity in both maps is found along the western and southern coasts of Norway. Seismic activity is low throughout much of the broad continental shelf north of Europe in the Barents Sea except in Spitsbergen, where a number of epicenters are found [*Husebye et al.*, 1975]. These shocks are definitely located well to the east of the plate boundary along the mid-Atlantic ridge. As in Britain and eastern Greenland, activity along the west coast of Norway and in Spitsbergen coincides with a Caledonian fold belt. Seismic activity in southern Norway is also high in the vicinity of the Olso graben, a Permian structure which is also noted for its ring dikes [*Ramberg*, 1976]. Subsidence of the graben may have continued well into the Mesozoic and perhaps into the Tertiary [*Whiteman et al.*, 1975]. This zone of subsidence can be traced by marine surveys southwest into the area of epicenters (Figure 30) in the Skagerrak trough between Denmark and Norway [*Whiteman et al.*, 1975; *Ramberg*, 1976]. These structures, as well as the Caledonian zones of weakness along the west coast of Norway and Spitsbergen, were reactivated during the early opening of the North Atlantic in the early Cenozoic.

As was mentioned in the discussion about seismic actvity near Scoresby Sund, Greenland, the highest historical activity in Norway is concentrated between 61° and 63°N near the end of the Jan Mayen fracture zone [*Husebye et al.*, 1975; *Bungum and Husebye*, 1978]. In a predrift reconstruction of Norway and Greenland, that Norwegian zone of activity appears to be conjugate to the relatively high activity near Scoresby Sund.

I have not attempted to relate the occurrence of earthquakes in western Europe to either the distribution of alkaline rocks of Cenozoic age or the distribution of major fracture zones that were active in the early spreading history of the North Atlantic (with the exception of the Jan Mayen fracture zone); the early opening is complicated by movements along the Bay of Biscay–Labrador Sea zone and by the development of a number of troughs in the area between Rockall bank and the North Sea [*Whiteman et al.*, 1975]. Also, the continental margins are generally thickly sedimented, and the distribution of older fracture zones is still being worked out.

Northern USSR and the Urals

Very few earthquakes have been detected from those parts of the USSR that are well north of the zone of Alpine deformation or west of the landward continuation of the plate boundary that extends across the Arctic Ocean into the northeastern USSR. A few small events have been detected from the Urals [*Academy of Science of the USSR*, 1962], but their number and magnitude are low in comparison to those for

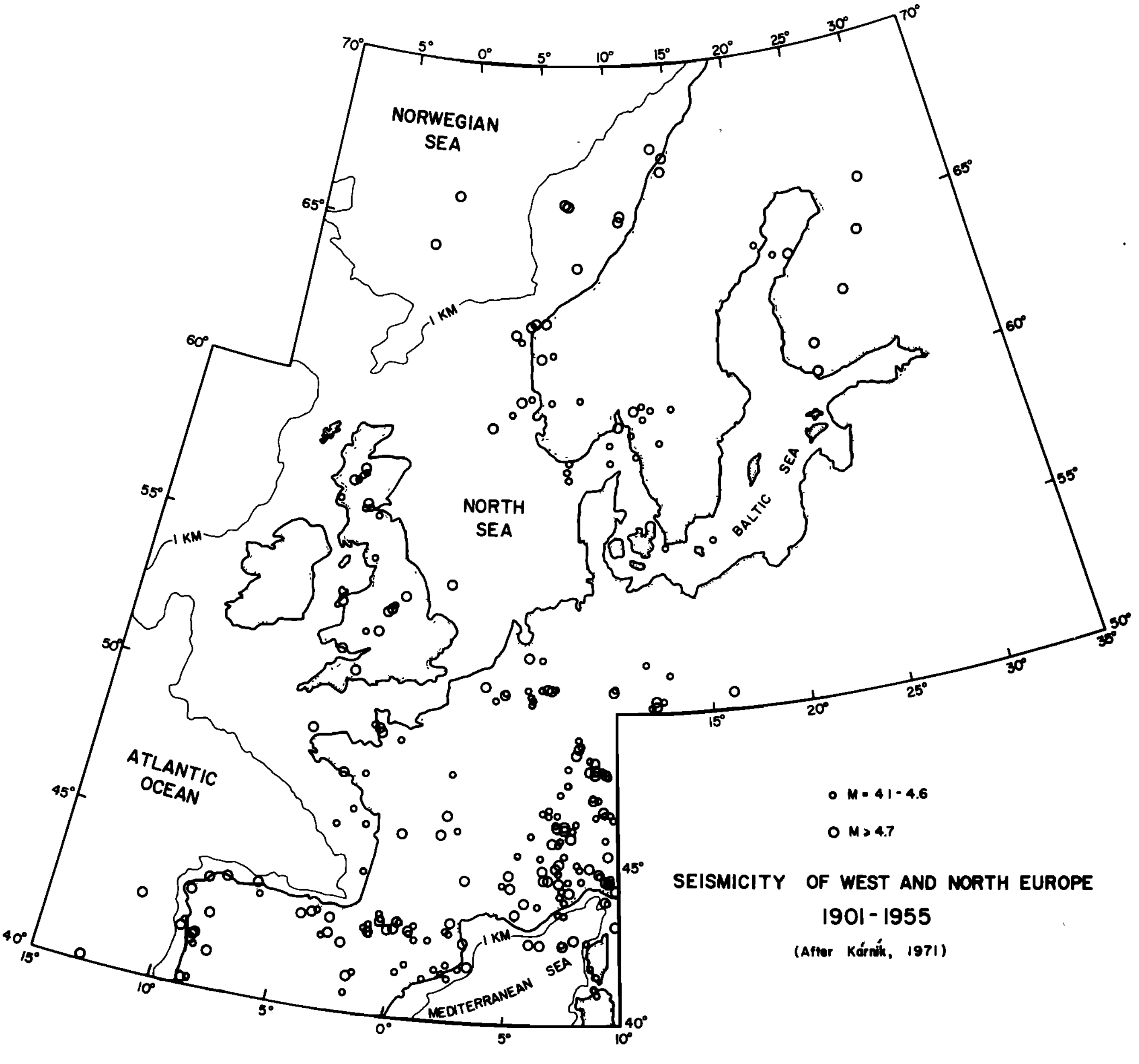

Fig. 30. Earthquakes in northern and western Europe, 1901–1955 [after *Kárník*, 1971; *Oliver et al.*, 1974].

eastern North America, northern Canada, or the west coast of Norway. The last major deformation in the Uralian fold belt appears to have involved the suturing of the Russian and Siberian platforms at the end of the Paleozoic [*Hamilton*, 1970]. The low level of seismic activity in the Urals indicates either that the stresses responsible for the higher levels of activity in, say, the Appalachians are not found in every Paleozoic fold belt or that reactivation of older faults by continental rifting and distention is needed before slippage can occur. Nevertheless, the recent uplift of the Urals resembles that of eastern Australia and the Appalachians.

Peninsular India

A number of historic and instrumentally recorded earthquakes have occurred in peninsular India well to the south of the main plate boundary along the Himalayas [*Balasundaram*, 1971; *Chandra*, 1977]. The continental margins of India appear to have developed by sea floor spreading as India split away from adjacent parts of Gondwanaland. Focal mechanism solutions for several shocks in peninsular India are characterized by strike slip motion or thrust faulting such that the axes of maximum compression are each oriented north [*Sykes and Sbar*, 1974; *Chandra*, 1977], that is, nearly perpendicular to the plate boundary. High compressive stresses may be transmitted into the Indian plate from the zone of continental collision farther north along the plate boundary between Eurasia and India. It is not unreasonable that high stresses may be associated with continental collision, since the low average density of the continental lithosphere (i.e., its greater buoyancy) resists the subduction of continents. Thus stresses in peninsular India, much like those north of the Alps, may be a plate-boundary-related phenomenon.

Peninsular India has experienced five shocks of magnitude greater than 5.2 and a number of historic shocks of intensity

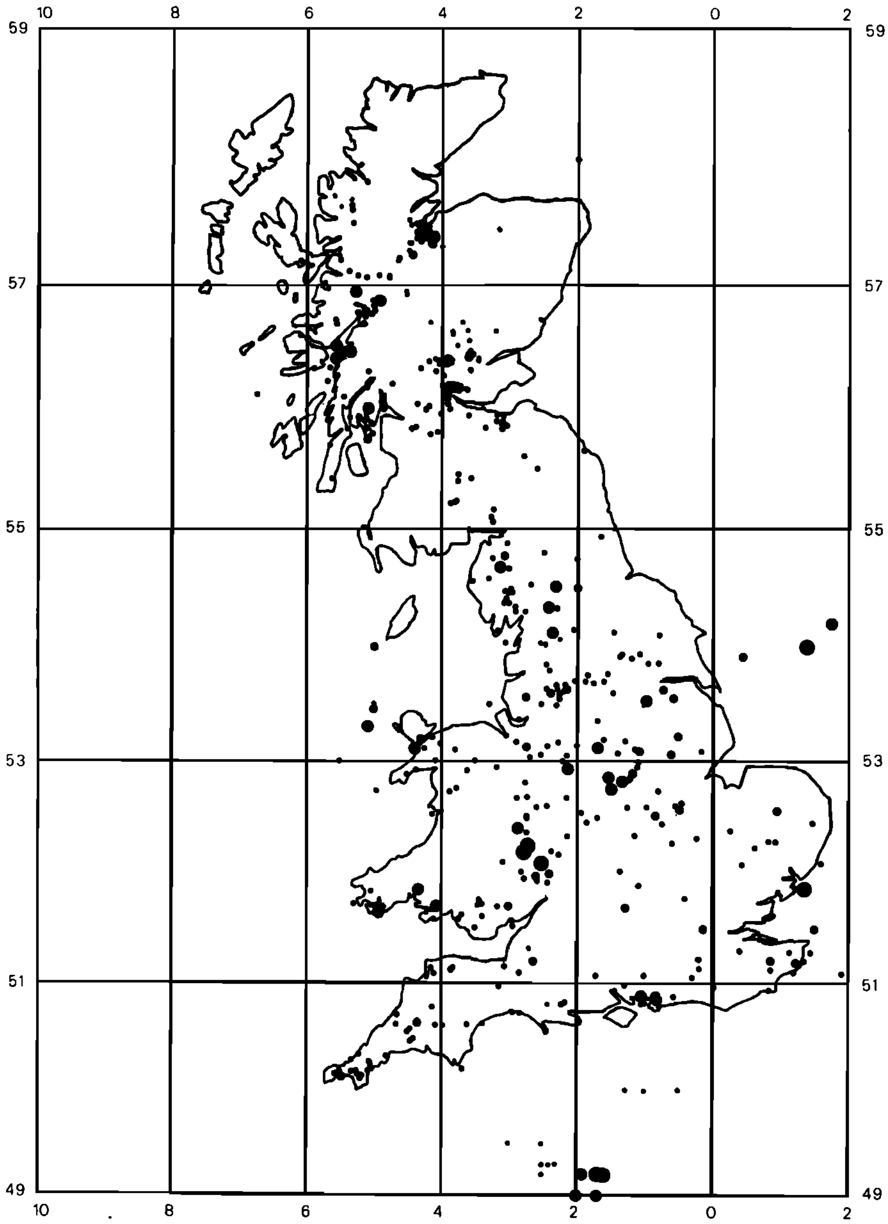

Fig. 31. Earthquakes in Britain from 1800 to 1970 [after *Lilwall,* 1976]. Natural Environmental Research Council copyright, reproduced by permission of the director, Institute of Geological Sciences, London SW7 England.

VII or greater [*Chandra,* 1977]. Hence it is clearly not aseismic. These shocks are concentrated either near the coastlines, which appear to be fault controlled, or along older fault zones striking northeast [*Grady,* 1971; *Chandra,* 1977]. Carbonatites and kimberlites are found along some of the latter zones in south India [*Grady,* 1971].

Margins of Red Sea and Gulf of Aden

Since the Gulf of Aden and the Red Sea developed by sea floor spreading during the Cenozoic, their continental margins and the continents adjacent to them are likely places to seek evidence for intraplate tectonism that may be associated with faults parallel to these features as well as with reactivated faults near the ends of transform faults. Using a variety of geological and geophysical methods, *Garson and Krs* [1976] infer that a number of transverse fractures are oriented perpendicular to the Red Sea in southeastern Egypt and in Saudi Arabia. The continental fractures appear to join transform faults within the Red Sea, as is inferred from topographic offsets, metalliferous deposits, and the occurrences of mafic and ultramafic rocks. They postulate that the direction of spreading in the Red Sea (and hence the trend of its transform faults) was guided by continental fractures of Precambrian age trending ENE. The presence of several much younger alkalic ring structures and carbonatites along these continental fractures argues that they were reactivated during the opening of the Red Sea. Garson and Krs also mapped a number of other fractures in Egypt and Saudi Arabia that strike parallel to the Red Sea.

In the Afar depression at the southern end of the Red Sea, *Barberi et al.* [1974] also find that volcanic features parallel to the rift are offset by transverse features. The latter appear to have developed along older zones of weakness. The margins of the Ethiopian plateau at the western edge of the rift are offset in a similar fashion, much like the offsets of the continental margin of equatorial Africa.

Seismic activity (Figure 10) is high along the plate boundary in the center of the Gulf of Aden. Only one of the events in Figure 10 appears to lie off the active plate boundary of the gulf. A few epicenters, however, have been reported by the National Oceanic and Atmospheric Administration (NOAA) and the U.S. Geological Survey (USGS) along the margins of the Gulf of Aden in Somalia and in Arabia since 1966, the date of the most recent earthquakes plotted in Figure 10. Nevertheless, in the Red Sea, many epicenters lie well off the central axis. These shocks may be associated with faults reactivated during the opening of the Red Sea. Also, several historically active volcanoes (indicated by V in Figure 10) are present near the west coast of Arabia. Thus continental areas adjacent to the Red Sea are the sites of earthquakes, alkaline magmatism, and renewed movement along preexisting faults. It is not clear, however, whether these processes should be considered strictly intraplate phenomena, since they occur close to an active plate boundary.

FOCAL MECHANISMS AND STATE OF STRESS IN INTRAPLATE REGIONS

Very few focal mechanism solutions of earthquakes or measurements of stress in situ are available for intraplate environments. This situation, coupled with the small amount of geologic evidence bearing on the nature of recent movements, is a great drawback to understanding the state of stress within plates. Hopefully, this situation will change as more mechanism solutions become available for smaller earthquakes and more intensive efforts are made to search for recent deformation.

Eastern North America

Using focal mechanisms, hydrofracture, and strain relief measurements of stress and geologic evidence for postglacial deformation, *Sbar and Sykes* [1973] conclude that a broad area in eastern North America is characterized by large horizontal compressive stress with the maximum compressive stress oriented eastward. Large compression is here taken to be much larger than the horizontal compressive stress (about $\frac{1}{3}$ of the overburden stress) created by gravitational attraction alone. Several new focal mechanism solutions by *Yang and Aggarwal* [1977] for northern New York State and southern Quebec show a high degree of consistency with one another. They indicate thrust faulting along NNW striking planes, i.e., the trend of the seismic zone in Figure 17 that extends from northern New York into southern Quebec.

Fletcher and Sykes [1977] obtained a composite focal mechanism solution for a series of shocks near Dale in western New York State. It indicates thrusting along the NNE trending Clarendon-Linden fault. This fault experienced major offset during the mid-Paleozoic, and at least a portion of it appears to have been reactivated, judged from the distribution of earthquakes. *Herrmann* [1978] obtained strike slip mechanisms for two nearby shocks, wherein one of the nodal planes is also nearly parallel to the fault system. In his solutions the maximum horizontal compressive stress is oriented northeast to ENE. This direction agrees with that measured nearby by using the hydrofracturing technique [*Fletcher and Sykes*, 1977]. The proximity of thrust and strike slip faulting in these three mechanisms is not unexpected, since stress measurements in the area indicate that the minimum and intermediate principal stresses are nearly identical in size.

If earthquakes like those in eastern North America occur along old planes of weakness, the directions of principal stress that are usually calculated, assuming that new fractures are created [*Anderson*, 1951], may differ substantially from the actual stress tensor. Using focal mechanisms alone, the only conclusion that can be drawn about the stress field is that principal stresses must be oriented so as to cause the observed sense of motion on the preexisting fault. For example, present-day movements along the NNE trending Clarendon-Linden fault may be in response to a maximum compression oriented about N60°E as inferred from nearby hydrofracturing experiments [*Sbar and Sykes*, 1973; *Fletcher and Sykes*, 1977]. Thus in situ measurements of stress are required to ascertain more than a qualitative understanding of the present stress field.

As I mentioned earlier, focal mechanism solutions for nine earthquakes in southern New York and northern New Jersey [*Aggarwal and Sykes*, 1978] show reverse faulting along steeply dipping faults that strike NNE to northeast. In a study of focal mechanisms of other shocks in the Appalachians near the fall line, *Aggarwal* [1977] finds that one of the nodal planes for each shock is nearly parallel to older NNE to northeast striking faults and that the inferred axis of compression tends to be more nearly horizontal than vertical. These results, as well as geologic evidence for horizontal compression along the Stafford and Brandywine fault zones of Virginia and Maryland [*Mixon and Newell*, 1977], suggest that the contemporary stress field is different from that which prevailed in the Triassic and Jurassic during continental rifting. Hydrofracture measurements (Figure 26) in the eastern and central United States also indicate a high horizontal compressive stress which is oriented ENE to east [*Haimson*, 1976]. The stress field deduced from the orientation of dikes of Jurassic age in eastern North America also differs from that inferred for the initial stages of continental rifting in the Triassic-Jurassic [*P. B. King*, 1961; *de Boer*, 1967]. Also, dikes of Jurassic and Cretaceous age in the Champlain valley of Vermont and New York and in the Monteregian province of Quebec show strong east-west and northwest-southeast trends and fill extension fractures (G. McHone, written communication, 1976). A northwesterly strike, however, would be among the least likely directions for dikes to be emplaced if the present-day stress field were active at that time. Thus the stress field in eastern North America appears to have changed since the initiation of continental rifting.

Hydrofracturing measurements (Figure 26) and focal mechanism solutions in eastern North America [*Sbar and Sykes*, 1973, 1977; *Aggarwal*, 1977; *Yang and Aggarwal*, 1977; *Aggarwal and Sykes*, 1978] give nearly identical orientations of the maximum compressive stress at nearby sites. That principal axis is nearly horizontal as determined by those two techniques, and its orientation appears to be quite uniform over broad areas. Likewise, both techniques indicate that the least compressive stress is nearly horizontal and is oriented west to WNW in the Basin and Range province of the western United States [*Sbar and Sykes*, 1973; *Haimson*, 1976].

Zoback et al. [1978] performed hydrofracture measurements in sediments of early Tertiary age in wells near Charleston, South Carolina. They conclude that the maximum horizontal compressive stress trends northwest. While many of the fracture azimuths they measured were northwesterly, they state that the single most distinct fracture impression was N51°E. Since they did not record a breakdown event in their pumping experiments, they could not calculate the maximum stress directly. Also, it is not clear if fluids created new fractures or merely opened existing fractures. The state of stress in that area may be one favoring either strike slip or normal faulting. From auger drilling in the Charleston area, Zoback et al. conclude that normal faulting of Cenozoic age has occurred along a fault zone striking northwest. They cite reports of northwest striking normal faults that displace Eocene sediments near Langley in west central South Carolina as well as evidence of northwest directed horizontal compression of Cenozoic age in the north central and central parts of the state. Thus a northwest trending maximum compressive stress is consistent with most of the observations that they cite, with a focal mechanism solution of a shock in Virginia near the fall line [*Aggarwal*, 1977] and with geologic evidence for reverse faulting along the Stafford fault system of Virginia. It does not agree, however, with the hydrofracture measurement (Figure 26) by *Haimson* [1976] in westernmost South Carolina or with the focal mechanism solution reported by *Tarr* [1977] for a shock near Charleston. Obviously, more determinations of the directions and magnitudes of the principal stresses are needed in the area, both in the Mesozoic-Cenozoic sedimentary section and in more competent rocks. Perhaps, the most that can be said at this time about the stress field in the southeastern United States is that it does appear to be of tectonic origin, it differs from the stress field active during continental rifting in the Triassic, and it is not generated solely by gravitationally induced forces.

In situ stress measurements made near the surface of the earth by the strain relief (overcoring) technique tend to show much greater scatter even at nearby sites. Thus many strain relief measurements near the surface appear to be greatly perturbed by the presence of cracks and joints. Hence focal mechanisms and hydrofracturing measurements appear to give a more consistent and reliable qualitative measure of the orientation of the stress field. The hydrofracturing technique appears to be a superior method to obtain quantitative measurements of stress.

Western Europe

There is also considerable evidence from western Europe for large horizontal compressive stresses that are nearly uniform in orientation over large areas. Focal mechanism solutions for about 40 earthquakes either in or north of the Alps show either thrust or strike slip faulting such that the maximum horizontal stress is oriented northwest [*Ahorner*, 1975]. From in situ stress measurements made by the strain relief technique, *Hast* [1969] finds large horizontal compressive stresses in Norway, Sweden, Spitsbergen, Ireland, Iceland, and Portugal. The maximum compressive stress in those measurements is oriented west in Portugal and southern Norway and northwest to north in Ireland, northern Fennoscandia, and Spitsbergen. From the distribution of historic earthquakes and known faults, *Anderson* [1951] deduces that the maximum compressive stress in Britain is now oriented between NNW and north, which is close to that (northwest to NNW) measured by Hast in Ireland. The northerly direction of maximum compressive stress in Spitsbergen as determined by *Hast* [1969] does not agree, however, with the ENE direction obtained by *Bungum* [1977] from a focal mechanism solution.

In addition to problems related to making measurements near the surface of the earth, there is considerable debate whether strain relief measurements sample a combined effect of contemporary (applied) stress and older (residual) stress. Nevertheless, the measurements in western and northern Europe do agree with results from focal mechanisms and hydrofracturing measurements [*Ahorner*, 1975]. Double overcores taken in southern Germany indicate a negligible residual component to the stress field [*Greiner*, 1975; *Greiner and Illies*, 1977]. The strain relief measurements also support the conclusion reached by examining focal mechanisms that large horizontal compressive stresses are common in intraplate areas of continents.

There is also geologic evidence in western Europe for compressive stresses oriented northwest. Volcanic features of Quaternary age in the Eifel and Neuwied areas near the Rhine are oriented northwest, that is, nearly parallel to the directions of maximum compressive stress inferred from focal mechanisms and in situ measurements of stress [*Greiner and Illies*, 1977]. This is consistent with the emplacement of these igneous bodies by hydrofracturing of the crust such that their strike (northwest) is nearly perpendicular to the direction of least horizontal compression (northeast).

Other Intraplate Regions

Focal mechanism solutions for four earthquakes in Australia (Figure 11) also are characterized by thrust faulting and by large horizontal compressive stresses. Similar mechanisms were obtained for a shock in Gabon (Figures 5 and 7), which was located in the Congo craton, and for an earthquake in the Gulf of Guinea (Figure 4). The maximum compressive stress inferred for the latter shock trends northwest and agrees with the direction obtained for coastal Liberia by *Hast* [1969], using the strain relief technique. The northeasterly direction of maximum compression inferred for the shock in Gabon is nearly parallel to the strike of the Cameroon line. A northeasterly direction of maximum horizontal compressive stress is obtained for the Cameroon line if contemporary magmas are being emplaced by hydrofracturing as dikes trending northeast. Nevertheless, the state of stress in the western portion of the African plate is quite different from the extensional environment deduced from both focal mechanisms and geologic evidence for the East African rift system [*McConnell*, 1972; *Maasha and Molnar*, 1972; *Gay*, 1977].

Mendiguren and Richter [1978] obtained focal mechanism solutions for five shocks and for a swarm sequence in various parts of the South American plate to the east of the Andes. They consistently find thrust mechanisms for these shocks. Likewise, *Sykes and Sbar* [1974] and *Chandra* [1977] report either thrust or strike slip faulting with the maximum compressive stress oriented north in peninsular India. The few mechanism solutions available for Canada indicate thrust faulting near Hudson Bay, in the northern Yukon, and along the Ottawa-Bonnechere graben northwest of Ottawa; strike slip faulting near the Arctic margin of northern Canada; and strike slip and high-angle reverse faulting along the east coast of Baffin Bay [*Sykes*, 1970*a*; *Hashizume*, 1973, 1974; *Sykes and Sbar*, 1974; *Qamar*, 1974; *Hasegawa*, 1977].

Some confusion may appear to exist about the focal mechanisms for shocks near the east coast of Baffin Island, since *Hashizume* [1973] and *Sykes* [1970*a*] report tension perpendicular to the coast of the island. The mechanism of a shock in 1963 reported by Sykes contains one nodal plane of nearly vertical dip and another that is nearly horizontal. One of Hashizume's solutions is also of this type. If the nearly vertical nodal plane is chosen as the fault plane, the block on the northeast side (the sea side) moves up in relation to the rest of North America. Hence the motion is opposite to that which is thought to have occurred during the early rifting history. The other mechanism reported by *Hashizume* [1973] involves strike slip faulting. *Qamar* [1974] obtained a strike slip mechanism for the 1963 shock by placing the focus in low-velocity crustal material. His velocity model is probably better than that used by *Sykes* [1970*a*] for the same shock. Hence focal mechanisms in that area do not reflect horizontal extension perpendicular to a former spreading center in Baffin Bay. This reversal in the sense of movement with time and the change in the stress regime are similar to those found for the eastern United States and Australia.

The mechanism of a shock in 1971 [*Sykes and Sbar,* 1974] on the continental margin of northeast Greenland (Figure 29) is one of the few intraplate solutions involving normal faulting. One possible reason for this is that the shock is located close to the axis of the mid-Atlantic ridge even though it occurred within continental crust. *Sbar et al.* [1970] report normal faulting for a series of small shocks in New Jersey in 1969. *Aggarwal and Sykes* [1978], however, find that a broad area including that shock is not characterized by normal faulting. Thus evidence for either thrust or strike slip faulting is growing for intraplate regions of continents, while that for normal faulting is limited to about three mechanism solutions. From focal mechanisms, *Sykes and Sbar* [1974] also find large horizontal compressive stresses in oceanic lithosphere older than about 10–20 m.y.

Thus focal mechanisms, in situ stress measurements, and geologic evidence indicate that several late Precambrian and Paleozoic fold belts that have been reactivated by more recent continental separation are now characterized by large horizontal compressive stresses that are fairly uniform in direction. This stress field is different from that deduced from the formation of grabens of the Newark type during the early stages of continental rifting and separation. Possible mechanisms for generating these stress fields within continents are discussed in the next section.

DISCUSSION

Plate Tectonic Framework of Intraplate Earthquakes

A number of inferences about intraplate tectonism can now be drawn by comparing the occurrences of earthquakes, alkaline rocks, Cretaceous and Cenozoic faulting, vertical movements, and the type of faulting as inferred from focal mechanisms of earthquakes. This study has concentrated on examining regions that are remote from plate boundaries and hence are clearly intraplate in character. Most of the investigation was aimed at intraplate areas of continents in the vicinity of continental margins that developed by rifting and sea floor spreading in either the Mesozoic or the Cenozoic. These areas include eastern North America, Brazil, the Atlantic margins of Africa, Greenland, Australia, Norway, Britain, Spitsbergen, peninsular India, Antarctica, continental areas adjacent to the Red Sea and the Gulf of Aden, northern Canada, and Baffin Bay.

With the exception of Antarctica, each of these intraplate areas has a record of historic and instrumentally recorded earthquakes. While the seismic activity in these regions is low in comparison to that at most plate boundaries, a number of damaging shocks, some of magnitude as large as 7–7.5, have occurred in some of these areas. In several of these regions the felt areas for intraplate shocks tend to be much larger than those for shocks of similar energy release along plate boundaries. Hence since several of these regions have a high population density, the seismic risk is clearly not negligible.

Within these intraplate regions, seismic activity does not appear to be distributed at random in space but tends to be concentrated either near the ends of major oceanic transform faults along preexisting zones of deformation or along faults of old fold belts within the thicker lithosphere of the continents. These preexisting faults generally trend either parallel or transverse to modern continental margins and were reactivated during the early stages of continental fragmentation and drift. Also, seismic activity is relatively high along belts of young alkaline magmatism, such as the Mississippi Embayment, the Cameroon line, and the Monteregian Hills, which appear to be either rifts that did not develop into full ocean basins (failed arms) or tectonism extending inland from the ends of major oceanic transform faults. It is understandable that earthquakes would tend to occur near continental margins along major preexisting tectonic features that were reactivated during continental fragmentation. These reactivated features probably have not been healed by metamorphism, and the pattern of faults has not been reset by continental collision.

Some of the best evidence for the concentration of seismic activity, particularly the occurrence of large shocks within continents near the ends of major transform faults, comes from the Accra area of Ghana near the end of the Romanche fracture zone; the Adelaide area of southern Australia, where late Cenozoic faulting is found landward of several major transform faults; offshore Massachusetts near the end of the New England seamount chain, Grand Banks near the end of the Newfoundland fracture zone; Charleston, South Carolina, near the end of the Blake fracture zone; the Scoresby Sund and Nord areas of eastern Greenland near the ends of major oceanic transforms; coastal Norway near the end of the Jan Mayen fracture zone; and in the Canning basin of Australia near the ends of major fracture zones on the northeastern side of the Exmouth plateau.

Major oceanic transform faults appear to have developed offshore from preexisting fault zones or other zones of deformation within the continents. These preexisting zones were inherited from the last major period of orogenesis and apparently were reactivated in the early stages of continental rifting during either the Mesozoic or the Cenozoic. In many cases, alkaline magmatism was initiated along these transverse features at the time of initial rifting and continued for tens of millions of years. One or more of the following are often found near the ends of major transform faults or, in a few cases, along zones extending inland from the ends of transform faults: kimberlites, carbonatites, syenites, ring dikes, diatremes, and mafic dykes. These rock types and some of these tectonic features appear to be indicative of a deep-seated source, probably sublithospheric. Magmas probably penetrated the entire lithosphere during their emplacement. It is understandable that in the initial rifting of a new ocean, these preexisting zones of weakness would be most likely to be reactivated close to the continental margin, and less frequently, zones of alkaline magmatism would extend a few hundred kilometers inland along these preexisting zones (Figure 2).

Some of the better examples of tectonic features of this type that extend well into continents from continental margins include the White Mountain magma series of New England, the Monteregian Hills of southern Quebec, the Benue trough of Nigeria, the younger Nigerian granites and ring dikes, the

Cameroon line, the Adelaide zone of Tertiary and Quaternary faulting in southern Australia, and the Mississippi Embayment. None of these features occur in isolation within a continent, but each appears to extend to a continental margin that developed by rifting and sea floor spreading (or to join some other feature that extends to such a margin). Thus these features seem to be intimately related to the early development of continental margins and ocean basins.

Zones of alkaline magmatism of Cretaceous and Cenozoic age appear to be closely related spatially either to nascent rift zones that did not develop into oceanic spreading centers or to major preexisting tectonic features oriented transverse to the present continental margin. The distribution of earthquakes, however, is not always clearly related to those tectonic features. For example, seismic activity is high along that portion of the Boston-Ottawa seismic zone of *Sbar and Sykes* [1973] extending northwest from northern New York into Canada but appears to be lower along the northern half of the area occupied by the White Mountain magma series. Most of the seismic activity along the Mississippi Embayment is concentrated along its northern half. In the Appalachians and other older fold belts, earthquakes tend to occur along major preexisting faults that trend nearly parallel to present-day continental margins as well as along features transverse to the margins. Alkaline rocks, however, seem to be confined largely to the transverse features. The more complex distribution of earthquakes may be related to the fact that the contemporary stress field, to which the earthquakes respond, differs in many areas from that involved in the early stages of rifting and in the emplacement of alkaline rocks. Also, the contemporary stress field may have several sources.

Since alkaline magmatism appears to be indicative of a source in the asthenosphere, the transverse zones of weakness may extend through the entire lithosphere. Hence it is understandable that larger intraplate shocks tend to occur more readily along such deep features, where a greater rupture area is possible. From very meager evidence, *Sbar and Sykes* [1977] suggest that earthquakes in eastern and central North America only extend to depths greater than about 10 km in seismic zones that are located either along the inferred continental extensions of fracture zones or along the Mississippi Embayment, where young alkaline rocks are also found. If this is true, these greater focal depths would be in accord with the concept of deep fractures that cut through the entire lithosphere. It should also be noted that the association of earthquakes with alkaline rocks described here is quite different from mechanisms described by *McKeown* [1978] and *Kane* [1976], wherein mafic and ultramafic bodies are thought to act as stress concentrations.

Francheteau and Le Pichon [1972] pointed out that the Atlantic continental margins of Africa and South America are segmented into a series of blocks by major oceanic transform faults. Sedimentary basins along these and other passive continental margins (Figure 2) are normally confined to a particular block and are delimited by fracture zones oriented perpendicular to the margin. Since these blocks appear to behave as individual units, it would not be surprising that this block character would be reflected in the seismicity.

In practice, it may not always be easy to estimate future seismic risk for areas near continental margins, since marginal fracture zones have a spectrum of sizes and offsets from large to negligible. This paper has focused only on the most prominent offsets on the ocean floor adjacent to passive continental margins. As expected, more detailed mapping of fracture zones, such as off the eastern United States [*Schouten and Klitgord*, 1977], indicates many small offsets in addition to the major offsets discussed in this paper. Some of these probably affect the block structure and seismicity of the adjacent continental margin. It might be expected, however, that the largest transform faults would be associated with the largest shocks and that older faults landward of them would tend to be reactivated a greater distance into the continent. In several cases this appears to be true. The presence of a major preexisting tectonic boundary landward of a large oceanic transform fault, such as the suture zone near Accra, Ghana, appears to be indicative of a potential for future large shocks. The existence of a prominent belt of alkaline magmatism may also be indicative of deep faults and hence possibly of a greater potential for future shocks.

Earthquakes Along Old Orogenic Belts

As I mentioned earlier, seismic activity in a number of intraplate regions near continental margins, including the Appalachians, Britain, Greenland, Norway, and Australia, appears to be associated with preexisting faults of Paleozoic or Precambrian fold belts that trend subparallel to present continental margins. These older faults within continents also appear to have been reactivated during continental rifting in either the Mesozoic or Cenozoic. Geologic evidence is gradually accumulating for continued movement along some of them well after continental separation. As more precise locations have become available for earthquakes in these areas, many are falling on or very close to major faults that have had a history of repeated movement in the Precambrian or Paleozoic. Focal mechanism solutions are also indicative of recent movement along these older faults.

Wilson [1965] proposed that the present North Atlantic Ocean developed nearly along the line of former continental closure in the Appalachians. *Burke et al.* [1977] conclude that other portions of the present midoceanic rift system developed along older orogenic belts, which were also the sites of former oceans. Thus it is not surprising that preexisting faults within those orogenic belts were reactivated during the most recent phase of continental fragmentation, since those belts may be considered to be zones along which the continents were weak and were most readily split apart. The present Atlantic continental margin of Africa lies almost entirely within the zone of late Precambrian–early Paleozoic (Pan-African) orogenesis. The break that developed into the South Atlantic Ocean tended to follow the most recent belt of deformation in Africa and as much as possible to avoid passing through the older cratons. Also, Cenozoic rifting in East Africa generally follows Precambrian or Paleozoic orogenic belts and avoids passing through the older cratonic nuclei such as the Tanganyika and Rhodesian shields [*McConnell*, 1972; *Scholz et al.*, 1976]. Hence various types of intraplate tectonism appear to be contolled by the preexisting tectonic fabric. Seismic activity and many types of alkaline magmatism are largely confined to the youngest belts of deformation that predate the formation of present oceans.

Several observations about the absence or near absence of earthquakes also bear on the mechanism of intraplate tectonism. Antarctica and its continental margins appear to be totally aseismic, although Phanerozoic tectonism affected significant parts of that continent. The continental margins of Antarctica, like many others, developed by sea floor spreading in either the Mesozoic or the Cenozoic. The Urals, a late

Paleozoic orogenic belt now located within a continent, exhibit very little seismicity. Earthquakes in shield areas, such as the Canadian, Ukrainian, Baltic, and Angaran shield and individual blocks of the western Australian shield, are rare. Most events that might be thought to be located within these shields occur either in the younger (less than 1 b.y. old) orogenic belts overprinted on them, along failed arms or rifts extending into them, or along their margins where they have been ruptured by continental drift. Earthquakes tend to be concentrated either within the continents or on continental shelves but are nearly absent along the nearby continental slope, continental rise, or within the adjacent oceanic lithosphere [*Oliver et al.*, 1974]. For example, most of the earthquakes in the Appalachian region (Figure 14) are located west of the fall line even though older rocks extend well to the east beneath the coastal plain. In eastern North America, eastern Australia, Norway, and Greenland, regions of seismic activity tend to be areas of recent uplift as detected from geodetic leveling, changes in sea level, or the seaward tilting of Cenozoic sediments.

A common feature of earthquakes in these areas is that they occur along an old orogenic belt adjacent to a continental margin that developed during or since the Mesozoic. The near absence of earthquakes along the Urals seems to indicate that the development of a continental margin is generally needed to reactivate older fault zones and to serve as a source of earthquakes. Continental collision as is occurring along the northern edge of the Indian plate is another means of reactivating old faults in continents. The absence of earthquakes in Antarctica and the near absence of shocks seaward of the continental shelf (or seaward of the fall line in much of the eastern United States) indicate that these aseismic areas are characterized either by slow movements not associated with earthquakes, by faults that have been healed by metamorphism (and hence that sustain high stresses without being the site of earthquakes), or by stresses that are lower than those in the adjacent older rocks of the continents.

One possible explanation for the occurrence of earthquakes in older continental lithosphere and for their general absence within the adjacent oceanic lithosphere involves the great increase in the thickness of the lithosphere in passing from ocean to shield. Seismological and other evidence indicates that the low-velocity channel for seismic waves is weakly developed or is absent beneath shields in comparison to oceanic areas. Also, the relative rate of movement between plates that contain large shield areas is smaller than that between plates that are largely composed of oceans and young orogenic belts [*Minster et al.*, 1974]. Hence a significant source of the stress field within plates, especially in continental areas, may come from the drag exerted on the base of the lithosphere beneath shield areas. This resistance to movement may place shield areas under horizontal compression. The easterly trend of the maximum compressive stress in a large area of eastern North America is nearly the same as the direction of absolute movement of the North American plate as inferred from movement of the Yellowstone hot spot. Of course, this near coincidence may be a chance occurrence, since the inferred direction of absolute plate motion depends crucially on the validity of the hot spot reference frame.

The abrupt change in the thickness of the lithosphere near continental margins may concentrate stress there; the presence of faults that were reactivated by continental rifting in old fold belts would serve to release that stress in earthquakes. Stresses could be large in shield areas, but the seismicity could be low, since faults in shield areas usually would not be reactivated by continental fragmentation. The general aseismic nature of shields does not necessarily imply that stresses are low; reactivated faults may not be present. *Greiner and Illies* [1977], for example, show that stress as measured in situ is high in some areas of low seismicity, and vice versa.

The lack of earthquakes both in Antarctica and along its margins may be ascribed to the low rate of movement of the antarctic plate with respect to the deeper mantle. Hence the forces on the base of that plate may be negligible; also, stresses transmitted into Antarctica from its plate margins, which form almost a concentric ring of spreading ridges, may be small or radially symmetrical.

The impingement of India upon the rest of Eurasia in the Cenozoic has reactivated preexisting faults in a broad area to the north of the Himalayas [*Molnar and Tapponnier*, 1975]. Thus continental fragmentation and collision both appear to be methods of reactivating old faults. Deformation in central Asia and China is highly concentrated along the youngest orogenic belts that predate collision. Old stable blocks like the Tarim basin and the Ordos plateau are nearly ringed by seismic activity but are themselves nearly aseismic. These blocks are apparently strong enough to transmit the relatively high compressive stresses associated with continental collision. There, as in the Alps, stresses appear to be transmitted from a plate boundary far into the surrounding plates.

Another possible mechanism for generating earthquakes in old orogenic belts like the Appalachian belt is that they are relatively weak zones that are being compressed between two blocks of greater rigidity. This is easy to accept for an area like that of the Canning basin and Fitzroy trough of northwestern Australia (Figure 11), which is situated between older cratonic blocks. Older basement of Grenville age is found to the west of and along the western side of the Appalachian orogenic belt. Perhaps the older oceanic lithosphere of the western Atlantic, while presumably not as thick as that beneath the Appalachians, is relatively strong and capable of transmitting compressive stress. In that instance the uplift and seismicity of the Appalachians could be attributed to its being a relatively weak block that is being compressed between two stronger blocks. Some hydrofracture measurements of stress in the oceanic lithosphere would be of great value in ascertaining if high compressive stresses are present and if they are, what their orientation is. This mechanism of compression between two stronger blocks would probably require that faults beneath the continental rise and the slope become healed within about 50–100 m.y. of the formation of the oldest oceanic crust. It is clear that additional stress measurements and focal mechanisms are greatly needed to test these ideas about the source of stress within continental regions.

Many faults beneath the continental rise, slope, and outer continental shelf of several intraplate regions were evidently formed or reactivated during the early stages of continental rifting. Hence it is surprising that so few earthquakes are found in many of those areas. If seismic faulting is confined to more competent materials (that is, those with P wave velocities greater than about 6 km/s), then the depth range at which earthquakes could occur beneath most continental margins would be limited by the depth to basement and the depth of the brittle-ductile transition. In areas of very thick sediments it is possible that the material properties and temperatures may inhibit the occurrence of earthquakes in the more competent basement rock even if stresses are high. Nevertheless, earthquakes are absent or rare along parts of the continental shelf of eastern North America, where the thickness of sediments is

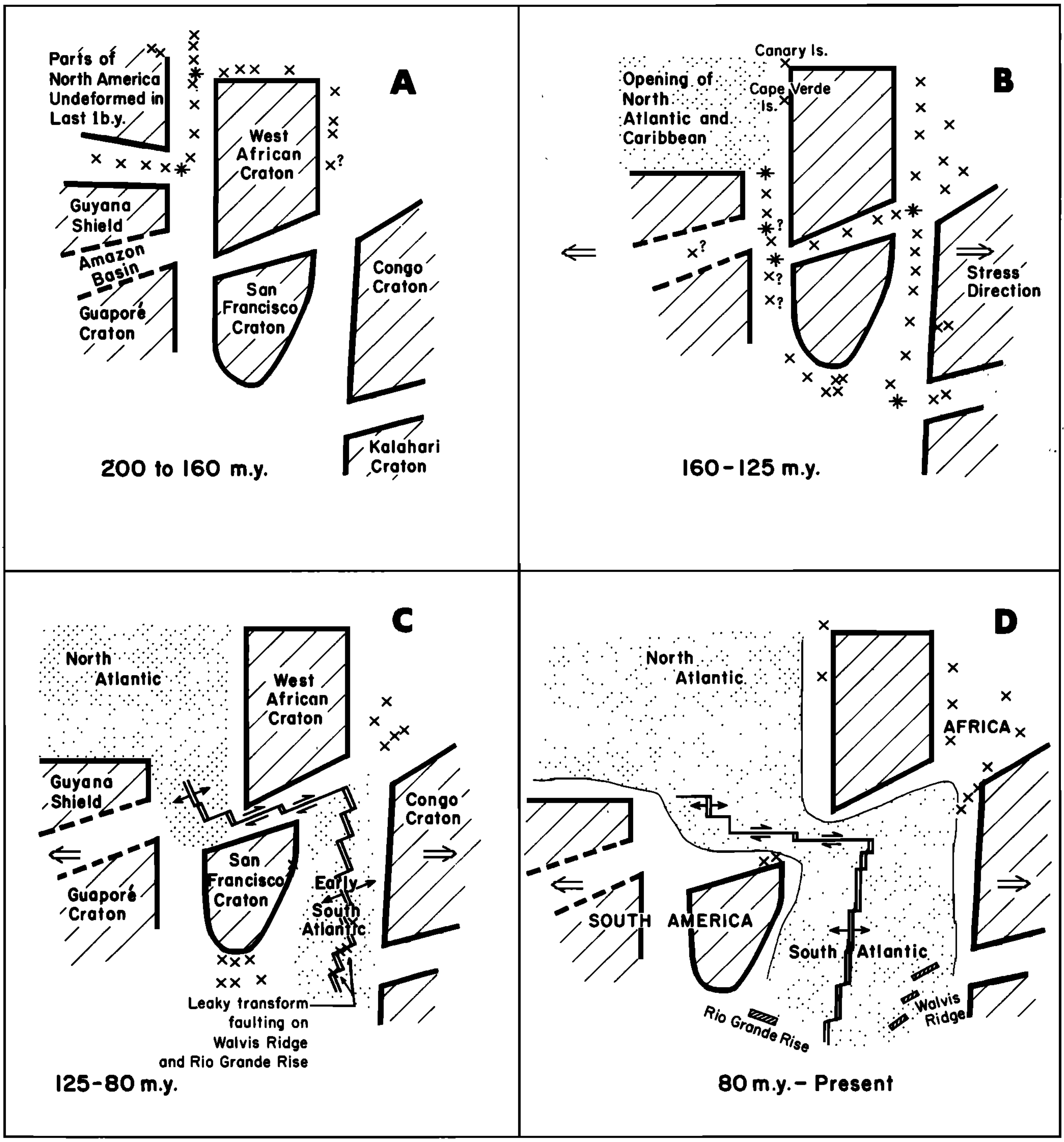

Fig. 32. Schematic representation of continental rifting between cratons, of magmatism (crosses), and of movements between old cratonic nuclei (hatched areas) of Africa and North and South America as a function of time. Note that continental rifting, new oceans, and magmatism develop along the youngest fold belts that predate rifting (blank areas) and avoid as much as possible passing through the older and thicker lithosphere of cratons. Large asterisks denote junctions of old fold belts that may become centers of unusually high rates of magmatism, i.e., hot spots. Magmatism during the early stages of continental fragmentation may be even more widespread along other mobile belts than is indicated here. A small piece of the West African craton appears to have been broken off during rifting and is now present in South America (San Luis craton, Figure 9). For discussion, see the text.

a few kilometers or less. Hence in those areas it is unlikely that the absence of earthquakes can be attributed to high temperatures in the basement rock.

Very little is understood about recent uplift in intraplate regions of continents. Although stresses and vertical movements in some of these regions may be related to the removal of Pleistocene ice, recent uplift and earthquakes are also present in eastern Australia, of which only small areas experienced Pleistocene glaciation. The easterly directions for the maximum compressive stress in a broad area of eastern North America also seem to bear no relationship to the distribution of Pleistocene ice. It is also unlikely that these high horizontal compressive stresses are caused mainly by the removal of a vertical load [*Voight,* 1966], i.e., that they are erosional in nature, since the same state of stress is found from focal mechanisms of shocks with depths of 0–25 km [*Sbar and Sykes*, 1977]. While many earthquakes in eastern North America extend far to the south of the limits of Pleistocene ice, the peripheral bulge surrounding the ice sheets may well include all of the seismic areas. Nevertheless, the present rates of movement inferred for the bulge are much smaller than those found from leveling data. While they may be a source of stress, thick wedges of sediments, such as those along the Gulf coast of the United States, are often not associated with appreciable seismic activity even though down-to-the sea faulting is often observed. Thus for these reactivated orogenic belts it seems reasonable to seek a source of stress related in some way to the development or structure of the continental margin, to the distribution of stresses within lithospheric plates, or to convective movements in the mantle.

The uplift of fold belts that are adjacent to present-day continental margins may be caused by the bending of the lithospheric plate as the oceanic portion of the lithosphere cools and sinks. That mechanism, however, does not explain the high horizontal compressive stresses present at shallow depth in most intraplate regions. Also, if such a bending stress is invoked as the only source of stress, it is difficult to explain the lack of earthquakes along the continental margins of Antarctica and the near absence of shocks along the margins of the Gulf of Mexico.

One other possible source of stress in plates is a thermoelastic one related to the cooling of the lithosphere [*Turcotte*, 1974]. Such a mechanism can place the upper portions of a plate (either oceanic or continental) in compression as hot material is welded onto the bottom of a plate and cools with time. Either this mechanism or one involving drag forces beneath shields may become dominant sometime after initial rifting. Thus the stress regime may change with time from horizontal extension to horizontal compression. *Mendiguren and Richter* [1978] and others emphasize 'ridge push' as a source of compressional stresses in plates. That mechanism could account for the change from extensional to compressive stress in the oceanic lithosphere as it moves away from a spreading center. If compressive stress can be transmitted across continental margins, that fact could account for compressive stresses and earthquakes in the adjacent older orogenic belts. Thus our lack of knowledge about the mechanism(s) generating contemporary compressive stresses and uplift in intraplate regions is a serious impediment to understanding the future occurrences of intraplate earthquakes.

Hot Spots, Alkaline Rocks, and Early Opening of an Ocean

Since many of the large oceanic fracture zones are located seaward of major preexisting faults, suture zones, failed arms (aulacogens), salients in orogenic belts, or other major tectonic boundaries, these features seem to have governed the locations of many major fracture zones that developed during the opening of the present-day oceans. In many areas the emplacement of alkaline rocks along these preexisting tectonic features appears to start at the time of initial rifting; in some cases it continues for as long as 100 m.y. The concentration of earthquakes within continents near the ends of fracture zones may be attributed to the relief of the contemporary stress field along these older zones of weakness that were reactivated during continental separation (provided, of course, that the feature is suitably oriented with respect to the present-day stress tensor).

Moving hot spot hypothesis. As a number of workers have pointed out, the moving hot spot (mantle plume) hypothesis fails to account for the long duration of magmatism (about 100 m.y.) and the lack of a clear space-time progression of magmatism for several belts of alkaline rocks.

If hot spots are taken to be active elements in the driving mechanism of plate tectonics and fixed to the lower mantle, it is indeed strange that the Cape Verde and Canary hot spots are connected by transform directions (i.e., the flow lines of *Pitman and Talwani* [1972]) to major tectonic zones in eastern North America. This association with flow lines would only be expected to occur if one plate was stationary with respect to a hot spot reference frame throughout the opening of the North Atlantic. This does not seem to have been the case, however. Rather than regard these two hot spots as being fixed with respect to the deep mantle and as the surface manifestation of active tectonic features that drive lithospheric plates, it seems more reasonable to envisage them as being passive features that owe their existence either to major zones of weakness on the sea floor, i.e., fracture zones, or to preexisting faults near the continental margin. Magma from the asthenosphere can rise to the surface along such faults as long as the stress tensor is oriented favorably with respect to the preexisting zone of weakness. In many ways this idea of an open crack along which magmas rise through the lithosphere is similar to the proposal of *Turcotte and Oxburgh* [1973] for the generation of the Hawaiian chain.

While many hot spots may be merely a passive phenomenon related either to the initial pattern of continental rifting or to the locations of major oceanic transform faults, some other hot spots may have a different origin. The Hawaiian-Emperor and Yellowstone hot spots, for example, each appear to have a fairly simple space-time migration pattern, which may reflect the movement of the lithosphere with respect to a hot spot in the asthenosphere. As yet, there is no clear evidence from geochemistry or geophysics that any of the igneous rocks thought to be associated with hot spots originate below the asthenosphere. The hot spot reference frame [*Minster et al.*, 1974] may not provide a measure of the movements of plates with respect to the deep mantle. Thus statements about the stationarity of, say, Antarctica or Africa with respect to the deep mantle should be treated with skepticism. Clearly, hot spots that appear to be fixed to the continents and that appear to be merely the passive upwellings of magmas at nodes in the preexisting tectonic fabric should not be used in deriving an absolute reference frame for plate tectonics. Likewise, reported movements between hot spots of 0.5–1 cm/yr may result from the inclusion of 'passive' centers of magmatism. Perhaps a more realistic reference frame can be developed by including only hot spots, like the Hawaii-Emperor chain, that show a clear space-time migration pattern and that are not obviously related to the preexisting pattern of tectonism.

Effect of cratons on the opening of the South Atlantic. Figure 32 illustrates schematically how the various cratons on either side of the Atlantic affected loci of rifting and magmatism, the development of fracture zones in the new Atlantic Ocean, and the formation of aseismic ridges.

In Figure 32*a*, multibranched rifting within a supercontinent, much like that which is occurring in East Africa today, develops in the early Mesozoic along several of the youngest preexisting orogenic belts that lay between the West African, North American, Guyana, and Baltic cratons. One of the rules of continental fragmentation appears to be for deformation to avoid as much as possible passing through the older, thicker lithosphere of the cratons. Instead, deformation tends to follow younger, narrow mobile belts located between cratons. Magmatism (crosses in the figure) develops from about 200 to 160 m.y. along several continental rift zones prior to the formation of the first oceanic crust in the North Atlantic about 160 m.y. ago. Junctions of the youngest preexisting orogenic belts are likely centers of greater than usual magmatism (asterisks) during these early stages of continental rifting. One of these becomes the Canary Islands hot spot. Other centers of increased magmatism (i.e., hot spots), such as those in the Cape Verde Islands and in the White Mountains of New England, develop at other zones of weakness in the preexisting tectonic fabric. They appear to be passive and to have no relationship to mantle plumes.

About 160 m.y. ago a single throughgoing plate boundary developed along some of these zones of multibranched rifting as oceanic lithosphere was generated by sea floor spreading in the nascent North Atlantic Ocean and in a nascent Gulf of Mexico and Caribbean Sea. Hence certain rift zones or regions of early Mesozoic magmatism, such as that along the east side of the West African craton, became stranded within one of the plates. These stranded zones may be sources of magmatism long after a throughgoing plate boundary becomes established if the stress tensor is oriented correctly.

In Figure 32*b* the North Atlantic has partly opened offshore from Newfoundland to the Caribbean by 125 m.y. ago. Magmatism continues in the Canary and Cape Verde islands near the intersection of the continental margin and some major preexisting zones of weakness. About 160 m.y. ago, multibranched rifting and magmatism commence along mobile belts located between the Guyana, Guaporé, San Francisco, West African, Congo and Kalahari cratons. Again, centers of abundant magmatism (asterisks) are situated near the intersections of three or more of the youngest preexisting mobile belts.

In Figure 32*c* a single throughgoing plate boundary develops by about 125 m.y. ago along the line of the present South Atlantic. Magmatism continues along some of the mobile belts of Pan-African age that did not develop into oceanic spreading centers. From about 125 to 80 m.y. ago the direction of sea floor spreading in the South Atlantic (and the orientation of transform faults) was northeasterly with respect to the present coordinates of Africa.

Le Pichon and Hayes [1971] and *Le Pichon and Fox* [1971] indicate that the directions of transform faults changed dramatically in the North and South Atlantic about 80 m.y. ago (Figures 32*c* and 32*d*). After that time, thick continental lithosphere was no longer present along the active parts of long transform faults such as the Newfoundland and St. Paul's fracture zones. Until that time the direction of transform faulting appears to have been constrained by the presence of thick lithosphere along the active portions of long transforms. There are three possible explanations for the change in the strikes of transform faults about 80 m.y. ago. One is that the driving mechanism changed in both oceans at that time. A second is that the driving mechanism propelling the plates changed prior to 80 m.y. ago, but transform faults could not readjust their directions to remain zones of pure shear until thick lithosphere was no longer present along the active portions of long transforms. A third possibility is that during the early opening the directions of transform faults were controlled or modified by the configuration of the cratons and mobile belts. Hence if the latter two hypotheses are correct, the early directions of opening (Figure 32*c*) may not have been exactly those that would have developed in response to the driving mechanism alone. In Figures 32*c* and 32*d* I suggest that the stress system related to the driving mechanism may have been oriented east with respect to contemporary Africa. By 80 m.y. ago, boundary forces on the edges of plates, particularly normal forces across long transforms in the equatorial Atlantic, may have become negligible as thin lithosphere came into contact on both sides of the active parts of those faults. Hence the direction of spreading in the South Atlantic during the last 80 m.y. may reflect mainly the forces driving the plates and not normal forces across long transforms.

During the early opening of the South Atlantic a series of long transform faults developed in the equatorial Atlantic along the southern margin of the West African craton. These transforms connected centers of spreading that formed along the Appalachian-Mauritanide fold belts of North America and West Africa with zones of Pan-African deformation situated between the Congo and the San Francisco craton. Hence the orientation of the series of long transforms may have been controlled or modified by the shape of the cratons to the north and to the south of those transforms.

Thus boundary forces across long transform faults may be important during the early opening of an ocean. On a smaller scale we can imagine pulling apart a mosaic that is akin to a jigsaw puzzle. The 'tabs' can only come apart in certain directions. Hence during continental rifting and during the earliest stages of development of an ocean the stress field may be quite variable in both time and space as the tabs clear one another. This may account for reported variations in the direction and style of deformation in the Triassic-Jurassic Newark rocks of eastern North America [*Sanders*, 1963, 1974] as well as for the different stress field under which Jurassic dikes were emplaced [*P. B. King*, 1961].

As I discussed earlier, the Benue trough of West Africa (Figure 5) developed as a tensional graben system from about 102 to 80 m.y. ago, i.e., prior to the realignment of fracture zones. The transition from extension to compression at the end of the Cretaceous appears to correlate almost exactly in time with the reorientation of fracture zones about 80 m.y. ago. The reactivation of old lines of weakness in the area of the trough can be attributed both to its location near a junction of pre-Mesozoic mobile belts and to a possible stress concentration related to its location near the end of long transforms across which a large component of normal stress may have existed prior to 80 m.y. ago. Thus intraplate stresses, particularly those in areas near continental margins and along old lines of weakness, may change as normal stresses are relaxed across major transform faults. If displacements across the Benue trough were small, a throughgoing plate boundary extending from the mid-Atlantic ridge to Egypt may not have existed as *Burke and Dewey* [1974] suggest. Instead, deformation may be more appropriately modeled as a response to intraplate stresses along an old zone of weakness that is confined to the continent and is not part of a plate boundary.

It is interesting that a number of aseismic ridges in the Atlantic, the Walvis ridge, Rio Grande rise, and New England seamount chain, are located entirely on oceanic crust that is older than 80 m.y. Although fracture zones are found along those features, they were not among the longest in the two oceans, and they were not the last transform faults along which thick lithosphere was finally cleared about 80 m.y. ago. Some transform faults of moderate length may have acted as leaky transforms (a small component of extension may have been present across them) during the period when some long transforms are thought to have had a component of normal stress across them (Figure 32*c*) Better dating of aseismic ridges could ascertain if magmatism occurred along them for tens of millions of years or if it occurred during a relatively short period of time.

The major segments of the Walvis ridge occur along a series of transform faults located seaward of the Damaride orogenic belt (Figure 5) of South-West Africa. That Pan-African belt, which is located between the Congo and the Kalahari cratons, formed a triple junction with other mobile belts that opened to form the South Atlantic. The triple junction may have had some control on the development of leaky transform faults in the young South Atlantic Ocean. Abundant magmatism along those leaky transforms prior to 80 m.y. ago may have generated the Walvis ridge and Rio Grande rise, which fit together in a reconstruction that closes the South Atlantic to the 80-m.y. isochron.

The cessation of magmatism along the New England seamount chain, in New England, and in southern Quebec about 80–100 m.y. ago also may be related to the reorganization in the trends of fracture zones and to possible change in stress that accompanied it. Perhaps a small component of extension may have existed for a considerable period of time before 80 m.y. ago along the New England seamount chain. One possible mechanism for magmatism in New England is that a crack propagated from the end of the oceanic transform that follows the New England seamount chain into the White Mountains and the Monteregian Hills. It is not clear, however, that a stress field near the end of a transform could propagate the necessary distance, about 1000 km, into the continent or that it existed at the end of the transform for the entire period from about 200–80 m.y. ago. Regardless of the origin of the stress field, a preexisting zone of weakness in the continent that was reactivated during the Triassic and that remained open to the emplacement of magmas for about 100 m.y. more satisfactorily explains the origin of rocks like those of the White Mountain magma series and the Monteregian Hills.

The North Atlantic north of 53°N opened much later than segments of that ocean to the south [*Pitman and Talwani,* 1972]. Thick lithosphere in Greenland, Spitsbergen, and Norway has either just been cleared or is about to be cleared along several of the long northwest striking transform faults in the Greenland Sea. Thus the plate tectonic development of the Atlantic north of 53°N during the last 65 m.y. may have been greatly influenced by large normal stress across the active parts of long transform faults. It is tempting to ascribe the Icelandic hot spot and the development of the Iceland-Faeroes ridge to leaky transform faulting rather than to the presence of a mantle plume. Thus the tectonics of the northern Atlantic may resemble the development of the South Atlantic prior to 80 m.y. ago. Changes in the orientation of fracture zones in the entire North Atlantic during the last 10 m.y. may result from the gradual (and sometimes sudden) release of normal stresses across long transforms in the Greenland Sea.

Triple junctions and failed arms. *Burke and Dewey* [1973] hypothesize that RRR triple junctions form above mantle plumes and are essential elements in the development of a new ocean. Usually, one of the three rifts about a triple junction is aborted and fails to develop into a full ocean as the other two arms often do. A number of their ideas about triple junctions also fit the model advanced here (Figure 32), whereby the junctions of rifts are situated at nodes in the preexisting tectonic mosaic but are not located above active plumes that extend deep into the mantle. Triple junctions appear to be a consequence of the fact that orogenic belts do not just end and of the topological fact that the intersection of three belts or rifts is much more likely than that of four or more. In fact, some fourfold junctions do exist, i.e., the Rungwe volcanics at the junction of four rifts in East Africa [*McConnell,* 1972] near 9°S, 33½°E (Figure 10).

In my model the activation of three intersecting zones of weakness during continental fragmentation and the early development of oceans is a consequence of multibranched deformation which typically appears to affect mobile belts on several sides of a craton. Once a throughgoing plate boundary comes into existence, multibranched tectonism tends to subside as some of the reactivated features are stranded in the interior of a plate. Thus two rifts that are situated along the new throughgoing plate boundary continue to develop into an ocean, while the third, which extends along another preexisting zone of weakness or mobile belt, fails to develop further.

If preexisting zones of weakness in continents exert a strong control over the locations of multibranched rifting, the initial arrangement of tectonic features in a supercontinent undergoing fragmentation may be much less regular than the 120° spacing of rifts envisaged by *Burke and Dewey* [1973]. They emphasize the active nature of mantle plumes in generating RRR triple junctions and point out that many large transform faults in the oceans appear to have originated above plumes and triple junctions. If preexisting zones of weakness in the continents also exert a marked control on the locations of oceanic transforms, however, many regions of unusually large amounts of magmatism, i.e., hot spots, may be a passive phenomenon.

A failed rift is also not a very satisfactory explanation for some alkaline rocks. For example, the structure of the White Mountains does not resemble that of a rift zone. Other features like the Benue trough and the Mississippi Embayment are more readily described as rifts. Also, the presence of ring dikes in some intraplate regions, such as New Hampshire, Nigeria, Scotland, and the Ahaggar, suggest that the two horizontal principal stresses were nearly equal at the time of emplacement. Those principal stresses usually differ in the vicinity of a well-developed rift zone, and dikes tend to be aligned along the rift. Features like the ring dikes of New Hampshire may develop in response to the reactivation of a broad zone of weakness without further development of the zone into a rift zone in the customary usage of the term. The emplacement of kimberlites in cratonic areas during the early opening of a nearby ocean is the most extreme example of reactivation of an old, deep zone of weakness that does not develop further or lead to the development of a rift zone.

Do Fracture Zones Develop Parallel to Old Zones of Weakness?

I have cited a number of examples in which seismic activity and alkaline magmatism develop along old zones of weakness in continents landward of large oceanic transform faults. The antiquity of the preexisting zones of weakness and the general lack of seismic activity in the oceanic lithosphere argue against the propagation of a new fault or crack into the continent from the ends of a transform. Instead, oceanic transform faults are likely to develop near a major preexisting zone of weakness that is oriented approximately transverse to the continental margin. While the development of oceanic transform faults may be caused by tensile stresses created by the cooling of the oceanic lithosphere [*Turcotte,* 1974] or by some other factor, the exact location and the amount of offset across oceanic transforms appear to be influenced greatly by preexisting zones of weakness.

Some oceanic fracture zones appear to have nearly the same strike as that of older adjacent zones of weakness in the continents. Nevertheless, the South Atlas fault, the suture zone at the southeastern side of the West African craton near Accra (Figure 5), the Takutu fault of northern South America (Figure 9), and the Darling fault of western Australia (Figure 11) differ in strike by at least 25° from that of transform faults located offshore from the ends of those faults. It might be expected that the mechanism driving continents apart would selectively reactivate old lines of weakness that have strikes either roughly parallel or perpendicular to the induced stresses. Probably, the near parallelism of some oceanic fracture zones and continental zones of weakness is a coincidence or the result of the selective reactivation of those zones of weakness that are oriented roughly parallel to the stress system set up by the driving mechanism.

Predrift Fit of Africa and North America

In a reconstruction of Africa and North America that closes the Atlantic, the Canary Islands nearly coincide with the western end of the New England seamount chain. Likewise, the Cape Verde Islands become nearly conjugate to Charleston, South Carolina, and to the western end of the Blake fracture zone. *Pitman and Talwani*'s [1972] flow lines, which they derived by successively closing the Atlantic to a given set of magnetic anomalies, also connect these points, which are now on opposite sides of the Atlantic. These flow lines, of course, have the same orientations as transform directions. *Le Pichon and Fox* [1971], however, could not obtain a good match of marginal ridges (located along major transform faults) on either side of the North Atlantic. Their failure probably resulted from attempting to match the Cape Verde Islands with the Cape Fear arch off North Carolina. A better match of various marginal ridges is obtained if the Cape Verde Islands are taken to be conjugate to the South Carolina-Georgia seismic zone. More recently, *Le Pichon et al.* [1977] reanalyzed the match of marginal ridges and pre-Mesozoic tectonic features on either side of the North Atlantic. Their reconstruction brings the following pairs into near coincidence: the Canary Islands and the western end of the New England seamount chain; the western end of the Norfolk fracture zone and Cape Blanc, West Africa; and the Cape Verde Islands and the western end of the Blake fracture zone.

Implications for Seismic Zoning and Need for Future Work

In this study I propose that earthquakes in intraplate areas are not distributed at random but that they occur mainly along faults and other zones of weakness associated with the last major orogeny in a region. A number of faults that experienced major movement during either the Paleozoic or the late Precambrian and that were reactivated during continental rifting in the Mesozoic or Cenozoic appear to be the sites of many intraplate shocks. Thus faults of this type, particularly throughgoing Paleozoic and Mesozoic faults, should not be assumed to be tectonically dead.

Preexisting zones of weakness that are now oriented nearly perpendicular to continental margins and that are situated near the ends of major oceanic transform faults appear to be the sites of some of the largest intraplate shocks. Since focal mechanisms are not available for many intraplate earthquakes, it is not clear if seismic faulting in continents near the ends of oceanic transform faults is oriented nearly perpendicular to or parallel to the continental margin. There is obviously a great need to obtain more focal mechanisms for intraplate shocks by analyzing surface waves for events in the magnitude range 4.5-5.5 and by installing local networks. As these areas are not as active as most plate margins and as the detection level for many of them has been low since the advent of instrumental seismology, much additional information on the relationship of earthquakes to tectonic features can be gained by studying historic reports of felt shocks and by installing local seismic networks.

In eastern North America many of the larger shocks, which have maximum intensities of IX-X, appear to be located near the ends of major transform faults or along structures extending inland from them. The largest known historic shocks in eastern North America that do not appear to be related to transverse features have maximum intensities of VIII [*Coffman and von Hake*, 1973]. Nevertheless, in other intraplate areas, shocks exceeding magnitude 6, such as those in western Australia during 1968 and in Gabon during 1974, have occurred along faults oriented parallel or subparallel to continental margins. Other large shocks, such as the New Madrid sequence of 1811-1812, are not obviously related to transform faults but are situated along tectonic structures extending inland from oceanic crust of Mesozoic or younger age. The occurrence of these various shocks indicates that it is probably premature to set an upper limit to the size of intraplate shocks for, say, the Appalachians. *McGuire* [1977] concludes that the available historic record of shocks in the eastern United States is adequate to establish activity rates for various seismic sources but not to calculate maximum possible sizes of events. Thus it is not justifiable on a statistical basis alone to assume that events larger than those observed historically will not occur in the future.

Although throughgoing faults in intraplate regions probably should not be considered to be totally inactive, they are obviously not as active as many faults along plate boundaries. Thus the major concern in seismic zoning is how active are they and is their repeat time 10^3 or 10^6 years. As *Evernden* [1975] points out, if future large shocks of the size of the Charleston, New Madrid, and Cape Ann earthquakes occur only in those specific areas, the repeat time at one of those locations is about 300 years (i.e., three large shocks in three areas in the last 300 years). If such events occur at random throughout eastern and central North America, however, the repeat time for experiencing, say, intensity X at a given location is much longer. Probably, reality is somewhere between these extremes. These great differences in calculated repeat times show the tremendous effect that a knowledge of the tectonics could have on better seismic zoning. Improved statistical methods probably cannot improve this situation significantly. In view of the great lack of knowledge about the tectonic setting of earthquakes in eastern North America, a conservative method of estimating seismic risk is in order for critical structures such as nuclear reactors, hospitals, and high dams.

This study supports the view that large intraplate shocks are situated along specific tectonic features, probably with repeat times of about 10^3-10^4 years; many other areas have lower seismic risk, i.e., longer repeat times for large shocks. Probably, the sites of future large earthquakes in eastern and central North America are not limited solely to Charleston, Cape Ann, and New Madrid but might include, for example, a shock near the end of the Norfolk fracture zone or one along the Wabash fault zone. The presence of major offshore fracture zones, major tectonic boundaries in the continents, and alkalic rocks in the subsurface may provide clues to estimating the potential of a region for a large future shock. The maximum depths of earthquakes may also be a guide to the presence of deep fault zones that extend through the lithosphere and which may have the potential for generating large shocks.

The orientation of the present stress tensor will also govern the seismic potential of faults and other zones of weakness that were reactivated during continental fragmentation. An active program of stress measurements using the hydrofracture technique could provide an assessment of the likelihood of stress release along those faults and of the uniformity (or lack thereof) of the stress tensor in space.

Several studies could help to increase greatly our knowledge of intraplate shocks and of seismic risk. Abundant Jurassic, Cretaceous, and Cenozoic thicknesses of sediments are found along many continental margins. Studies of these could provide a much needed record of the history of deformation both there and on the adjacent continents in time and space. Very little work has been done in the last 50 years in examining planation (peneplain) surfaces or in searching for deformation on land of Mesozoic or Cenozoic age. Renewed interest in these areas as well as in the analysis of lineaments as seen on satellite photographs may provide important insights about intraplate tectonism, and seismic risk.

Resource Implications

The occurrence of magmatism postdating continental separation near the ends of oceanic fracture zones may have considerable bearing upon the resource potential of various continental margins. Considerable thicknesses of sediments accumulated along the continental shelf and rise of eastern North America in the Mesozoic prior to the occurrence of Cretaceous magmatism near Charleston, along the New England seamount chain, in New England, and off New Jersey. This Cretaceous event may have created traps for petroleum and served as a heat source for converting organic material of Mesozoic age into a usable petroleum resource. Igneous rocks of this kind may have been buried subsequently by Cenozoic sediments.

Bonatti et al. [1976] argue that oceanic fracture zones are loci of metallogenesis and that important metal deposits appear to be aligned along the predrift landward extensions of some oceanic fracture zones. Deep faults extending through the lithosphere may provide a setting for increased circulation of mineralizing fluids. *Garson and Krs* [1976] conclude that some preexisting faults in Egypt and Arabia that where reactivated by the development of the Red Sea are important sites of mineralization. In cratonic areas where the continental lithosphere is thick, faults reactivated during the development of a nearby continental margin, such as off South Africa, may control the emplacement of kimberlites and diamonds. This hypothesis helps to explain the apparent enigma that kimberlites are usually emplaced in cratonic environments. A cratonic setting need not imply total stability since the Precambrian. Kimberlites and other rocks of deep-seated origin appear to owe their emplacement to a combination of the state of stress, presence of deep-seated faults, and tectonism in an adjacent area (such as the development of a nearby continental margin off South Africa or the elevation of the Colorado plateau). Perhaps, deep faults in some of the cratonic areas of western Australia, for example, may have been the loci of kimberlite emplacement during the development of adjacent oceans.

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