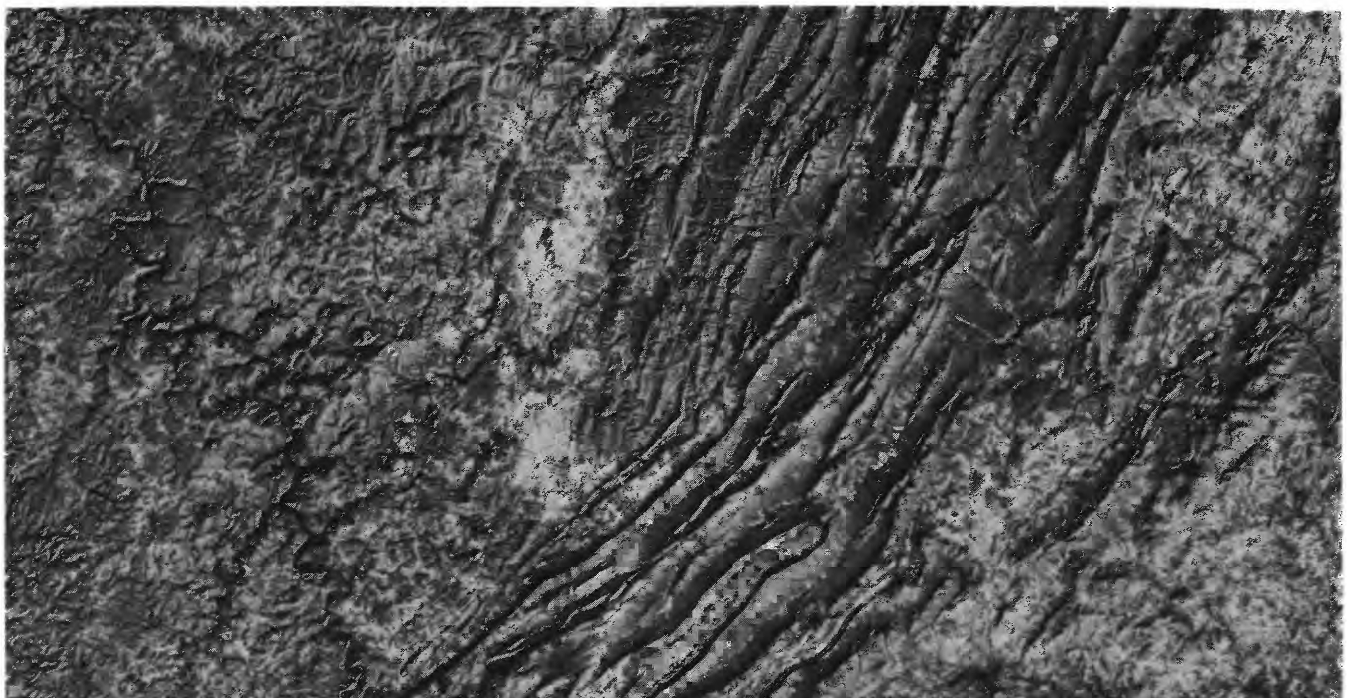
THE GILES COUNTY, VIRGINIA, SEISMIC ZONE—
SEISMOLOGICAL RESULTS AND GEOLOGICAL INTERPRETATIONS



U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1355

Landsat multispectral image showing Giles County, Virginia, and surroundings. Image number 2654-15090-7, taken November 6, 1976. The solid circle locates Pearisburg, Virginia, seat of Giles County and the presumed epicenter of the earthquake of May 31, 1897.

Cover designed by Norma J. Maes

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Angel's Rest promontory (center foreground, elevation 3633 ft [1107 m] above sea level) of Pearis Mountain with the town of Pearisburg and the New River (elevation 1581 ft [482 m] above sea level) at its base. View is toward the southwest. This mountain was erroneously reported to have cracked during the May 31, 1897, earthquake centered in the vicinity. The Narrows seismograph station (code: NAV) is located in Mill Creek Valley, this side of Sentinel Point promontory extending to the right (north). Refer to Narrows, Virginia-West Virginia 7.5 minute topographic map for details.

The Giles County, Virginia, Seismic Zone— Seismological Results and Geological Interpretations

By G. A. BOLLINGER *and* RUSSELL L. WHEELER

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1355

A description of a newly recognized seismogenic zone, with contributions towards evaluation of its seismic hazard

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1988

DEPARTMENT OF THE INTERIOR

DONALD PAUL HODEL, ***Secretary***

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, ***Director***

Library of Congress Cataloging-in-Publication Data

Bollinger, G. A.
The Giles County, Virginia seismic zone.

(U.S. Geological Survey professional paper ; 1355)
"A description of a newly recognized seismogenic zone, with contributions towards evaluation of its seismic hazard."
Bibliography: p.
Supt. of Docs no.: I 19.16:1355
1. Seismology—Virginia—Giles County. 2. Faults (Geology)—Virginia—Giles County. I. Wheeler, Russell L. II. Title. III. Series.
QE535.2.U6B63 1989 551.2'2'09755782 88-600018

For sale by the Books and Open-File Reports Section, U.S. Geological Survey,
Federal Center, Box 25425, Denver, CO 80225

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THE GILES COUNTY, VIRGINIA, SEISMIC ZONE—SEISMOLOGICAL RESULTS AND GEOLOGICAL INTERPRETATIONS

By G. A. BOLLINGER[1] and RUSSELL L. WHEELER[2]

ABSTRACT

A newly recognized 40-km-long seismic zone is described and interpreted. The zone is inferred to have been the locus of a damaging earthquake in 1897. That shock was the second largest known to have occurred in the Southeastern United States (MMI=VIII, m_b estimated at 5.8, felt over 725,000 km^2). It struck Giles County in southwestern Virginia, and a recurrence would affect populous regions on and near the central Atlantic seaboard. The seismic zone presents a hazard. We attempt to aid in evaluating the hazard by presenting and synthesizing new seismological data with geological inferences and deductions.

A five-station, 60-km aperture seismic network has been in operation in the Giles County locale since early 1978. For the subsequent 3-year monitoring period, 12 microearthquakes (M<2) have been detected. Hypocenters of eight of those 12 events, plus an additional four older, relocated, felt earthquakes ($3.2 \leq M \leq 4.1$; 1959–76), have a tabular distribution centered at Pearisburg, Va. That distribution is about 40 km long, 10 km wide, strikes N. 44° E., and has a nearly vertical extent of 5–25 km in depth. Thus, the Giles County seismic zone is defined presently by 12 earthquakes that span four orders of earthquake magnitude ($0 \leq M \leq 4$) and two decades of time (1959–80). We conclude that the 1897 earthquake occurred on that seismic zone. From the orientation of the zone, from evidence that greatest horizontal compressive stress trends east-northeast at seismogenic depths in and near Giles County, and from sparse *P*-wave first-motion data, we infer that the monitored microseismicity probably occurs by right-reverse motion on steep faults in the seismic zone, with the southeast side dropping down with respect to the northwest side.

In the Giles County locale, the upper 3–6 km of the crust are Paleozoic sedimentary rocks that have moved some tens of kilometers northwest on nearly horizontal thrust faults. The previously mentioned hypocenters for the region lie below the deepest likely thrust fault, indicating that Giles County seismicity probably has no simple relationship to surface geology.

Since Precambrian time, three deformational episodes could have formed steep faults under the present surface structures, at the observed hypocentral depths. These episodes were as follows: (1) As the Iapetus Ocean (Atlantic's predecessor) opened in late Precambrian or early Paleozoic time, northeast-striking normal faults formed, probably at the inferred Iapetan continental edge in central Virginia and at least as far northwest of that locus as Giles County. (2) In late Paleozoic time, thrust faults loaded the crust with several kilometers of overthrust sedimentary rocks, perhaps forming northeast-striking thrust-load faults in a brittle analog of isostatic depression caused by thrust masses and much lighter continental glaciers. (3) As the Atlantic Ocean opened in Mesozoic time, other northeast-striking normal faults formed on the present continental margin and inland of it.

The seismic zone seems most likely to have resulted from compressional reactivation of an Iapetan normal fault, which also may have been reactivated by late Paleozoic compression and Mesozoic extension. Two arguments support this conclusion. First, the seismic zone probably does not occur on a thrust-load fault. The zone underlies the thrust structures of southern Appalachian orientations (east-northeast), but those structures are not known to be displaced where they cross the zone. Thus, if the zone occurs on a thrust-load fault, the fault and its coeval causative central Appalachian thrusts would predate the southern Appalachian structures. That deduction contradicts stratigraphic and structural estimates of relative ages of southern and central Appalachian thrusting. Second, the zone probably does not result from a Mesozoic normal fault, because known locations of Mesozoic normal faults and grabens are well to the southeast of Giles County.

Not yet known is where else in the East reactivated Iapetan normal faults might generate shocks similar to that of 1897. However, our analysis enables us to suggest specific geological and geophysical investigations that may produce results useful in answering that question. Such investigations can concentrate on defining the area of probable occurrence of other Iapetan normal faults, and on determining whether the one inferred to underlie Giles County is uniquely active or is typical of others that might exist elsewhere.

INTRODUCTION

On May 31, 1897, a damaging earthquake struck Pearisburg, the seat of Giles County in southwestern Virginia (fig. 1). The shock was erroneously reported to have cracked Pearis Mountain, whose Angel's Rest promontory rises more than 600 m above Pearisburg and the New River. (See frontispiece.) The earthquake is especially important in the seismic history of the Southeastern United States, for the following reasons:

1. It is the largest shock known to have occurred in Virginia, and the second largest earthquake known in the entire Southeastern United States (Modified Mercalli Intensity (MMI)=VIII, body-wave magnitude (m_b) = 5.8, felt area = 725,000 km^2; Bollinger and Hopper, 1971; Nuttli and others, 1979; Street, 1979; see fig. 1).
2. It serves as the design earthquake for critical facilities sited in the Valley and Ridge and Blue Ridge provinces of the southeastern United States.

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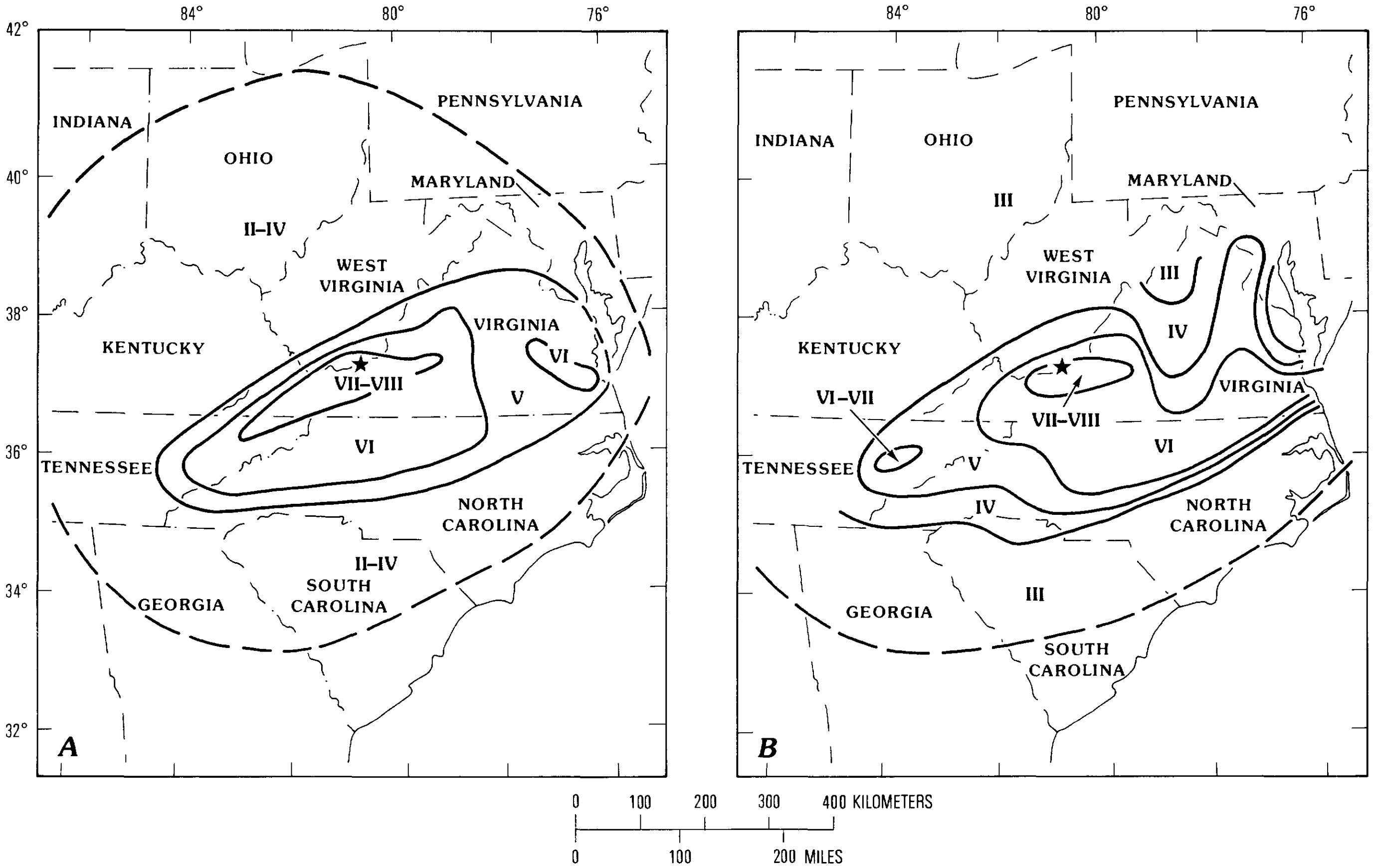


FIGURE 1.—Intensity maps for the May 31, 1897, Giles County, Va., earthquake. *A*, modified from Law Engineering Testing Company (1975); *B*, modified from Bollinger and Hopper (1971). Differences between the two maps reflect difference in data bases (Law Engineering Testing Company's was the larger) and in the interpreters. Star indicates the location of Pearisburg, Va., the presumed epicenter of the shock. Contours are drawn on values of intensity reported from various places. Typical intensity values for areas between or within contours are shown as Roman numerals. Dashed contours show approximate limits of data: earthquake was felt at least that far from the epicenter. Reproduced from Bollinger (1981a) with permission.

3. No earthquake activity prior to 1897 has been definitely assigned to the Giles County locale (Hopper and Bollinger, 1971; Reagor and others, 1980a, b). However, a foreshock-aftershock sequence did occur in conjunction with the May 31, 1897, main shock from May 3 to at least June 6 (Bollinger and Hopper, 1971). A local resident estimated that there were at least 250 distinct shocks observed at Pearisburg subsequent to May 3, 1897 (Campbell, 1898).

The felt aftershocks apparently ended in 1902 (MMI=V shock on May 18 near Pearisburg, the presumed epicenter of the main shock of 1897; table 1) or perhaps in 1917 (southwestern Virginia earthquake reported on April 19, MMI=II; Reagor and others, 1980a). There followed a quiescent period of 4–6 decades that ended in 1959 with the occurrence that year of three felt shocks (MMI=VI, IV, IV). In the following 2 decades (1960–79), six additional felt earthquakes (MMI≤VI) were reported from within 50 km of Pearisburg. The largest of those six shocks was the m_b=4.6, MMI=VI, Elgood, W. Va., earthquake of November 20, 1969 (felt area=324,000 km^2). Elgood is a small community just north of the Virginia–West Virginia border north of Giles County. Thus, there has been an apparent modern renewal of seismic activity (nine felt earthquakes in 1959–76) in or near Giles County (table 1).

This report has three purposes:

1. It presents and interprets results of a recent seismic monitoring program in the Giles County locale. (We define the Giles County locale as that area within 50 km of Pearisburg.) The first section of the report achieves this purpose.
2. The report attempts to integrate those results of the seismic monitoring program with what is known or reasonably inferred about local and regional geologic structure at seismic focal depths, which is accomplished in the second and third sections

TABLE 1.—*Chronological listing of earthquakes that occurred prior to 1978 in the Giles County, Va., locale (within 50 km of Pearisburg, Va.)*

[Data sources: Reagor and others (1980a, b). Letter code for "Quality" column is defined as follows: Determination of instrumental hypocenters is estimated to be accurate within the ranges of latitude and longitude listed; each range is letter coded as indicated--A, 0.0°-0.1°; B, 0.1°-0.2°; C, 0.2°-0.5°; D, 0.5°-1.0°; E, 1.0° or larger. Determination of noninstrumental epicenters from felt data is estimated to be accurate within the ranges of latitude and longitude listed below; each range is letter coded as indicated--F, 0.0°-0.5°; G, 0.5°-1.0°; H, 1.0°-2.0°; I, 2.0° or larger. Body-wave magnitude (m_{bLg}) (Nuttli, 1973; Bollinger, 1979); MM, Modified Mercalli intensity rating in Roman numerals (Wood and Newmann, 1931). Leaders (---) indicate no value available]

Date			Locality	Origin time (UTC)			Latitude (north)	Longitude (west)	Depth (km)	Quality	Magnitude (m_{bLg})	Intensity (MM)
Year	Month	Day		Hours	Minutes	Seconds						
1876	Dec.	21	Wytheville, Va.	15	30	--	36.9°	81.1°	--	G	----	II
1879	Sept.	1	----do----------	12	--	--	36.9°	81.1°	--	G	----	II
1885	Feb.	2	----do----------	12	10	--	36.9°	81.1°	--	G	----	IV
1897	May	3	Pulaski, Va.----	17	18	--	37.1°	80.7°	--	G	[1]4.3	VII
1897	May	3	----do----------	19	--	--	37.1°	80.7°	--	G	----	III
1897	May	3	----do----------	21	10	--	37.1°	80.7°	--	G	----	III
1897	May	3	----do----------	23	--	--	37.1°	80.7°	--	G	----	III
1897	May	31	Pearisburg, Va.	18	58	--	37.3°	80.7°	--	G	[1]5.8	VIII
1897	June	29	----do----------	03	--	--	37.3°	80.7°	--	G	[1]4.0	IV
1897	Sept.	4	Wytheville, Va.	11	--	--	36.9°	81.1°	--	G	----	III
1897	Oct.	22	----do----------	03	20	--	36.9°	81.1°	--	G	----	V
1898	Feb.	5	----do----------	20	--	--	37.0°	81.0°	--	G	[1]4.3	VI
1898	Feb.	6	----do----------	02	--	--	37.0°	81.0°	--	G	----	II
1898	Nov.	25	----do----------	20	--	--	37.0°	81.0°	--	G	[1]4.6	V
1899	Feb.	13	----do----------	09	30	--	37.0°	81.0°	--	G	[1]4.7	V
1902	May	18	Blacksburg, Va.	04	--	--	37.3°	80.4°	--	G	----	V
1917	Apr.	19	Wytheville, Va.	--	--	--	37.0°	81.0°	--	I	----	II
1959	Apr.	23	Virginia-West Virginia border.	20	58	40.2	37.40°	80.68°	1	A	[2]3.8	VI
1959	July	7	Pearisburg, Va.	23	17	--	37.3°	80.7°	--	F	----	IV
1959	Aug.	21	----do----------	17	20	--	37.3°	80.7°	--	F	----	IV
1968	Mar.	8	Narrows, Va.	05	38	15.7	37.28°	80.77°	8	A	4.1	IV
1969	Nov.	20	Elgood, W.Va.	01	00	09.3	37.45°	80.93°	3	A	4.6	VI
1974	May	30	Virginia-West Virginia border.	21	28	35.3	37.46°	80.54°	5	A	3.6	V
1975	Mar.	7	Blacksburg, Va.	12	45	13.5	37.32°	80.48°	5	A	3.0	II
1975	Nov.	11	Giles-Bland Counties, Virginia, border.	08	10	37.6	37.22°	80.89°	1	A	3.2	VI
1976	July	3	Virginia-West Virginia border.	20	53	45.8	37.32°	81.13°	1	A	[3]2.1	---

[1] Nuttli and others (1979).
[2] J. W. Dewey and D. W. Gordon (written commun., 1980).
[3] G. A. Bollinger (unpub. data, 1976).

of the report. Our goal of integrating results from diverse portions of seismology and geology has required us to write for two audiences. Thus, we have included some material that may seem unnecessary to members of one audience or the other. As geologists and seismologists reviewed drafts of this report, some specialists in each discipline questioned inclusion of some of the details. So, we have relegated highly specialized material to appendixes, but in general we have preferred to risk too much detail rather than to chance omitting something of interest or importance.

3. To the extent that the second purpose is fulfilled, the report can contribute to improved evaluation of seismic hazard. Throughout much of the Western United States, many known or suspected seismic faults are exposed for study, together with the geologic evidence of their past activity. The best known example is the San Andreas fault zone. There, the geologic record forms an important

adjunct to records of historical and instrumental seismicity, and the resolution and reliability of hazard evaluation benefit markedly. In sharp contrast, throughout most of the East any seismic faults are buried beneath sediments, sedimentary rocks, or thrust sheets. Thus, evaluation of seismic hazard in the East often must be based mostly or entirely on the historical seismic record. However, in most Eastern areas, evaluation would benefit if seismicity could be associated with individual faults or classes of faults, whether buried or exposed. The geological characteristics of such faults could then be used to infer where else in the East similar faults might occur, and might also be subject to seismic reactivation like that which occurred in 1897. In this way an evaluation of the seismological and geological characteristics of Giles County may lead eventually to enhanced evaluation of seismic hazard for urban centers, critical facilities, and lifelines far removed from semirural Giles County itself.

Since 1962, a Worldwide Standard Seismograph Network (WWSSN) Observatory (call letters: BLA) has been in operation at Blacksburg, Va., some 35 km southeast of Pearisburg, the county seat of Giles County. Operation of a five-station network, centered in Giles County, began in April 1978. (See fig. 2 and frontispiece.) That network was designed to enclose the aforementioned concentration of historical and recent epicenters. Following discussions of terminology and local geology, we will present the results of that network monitoring.

A reviewed but mostly unedited draft of this report was published by Bollinger and Wheeler (1982) and summarized by Bollinger and Wheeler (1983). This report and that of Bollinger and Wheeler (1982) differ only in editorial details, the addition of explanatory material dealing mostly with regional geology and seismological matters, addition or completion of citations of references that were unpublished in 1982, changes of 1-2 km in earthquake locations and their locational uncertainties as more data have become available, and the revision and expansion of Appendix D. Monitoring and analysis of earthquakes after December 1980 (Munsey and Bollinger, 1984; Bollinger and others, 1985; Gresko, 1985; Viret and others, 1986) support and build on the conclusions of this report and those of Bollinger and Wheeler (1982, 1983) and suggest no reason to change those conclusions.

TERMINOLOGY

Some of the terms used in this report should be defined and their usage explained, because usage differs from one specialty to another, usage changes rapidly, and in some contexts the terms have economic and legal implications. Some abbreviations are also defined.

balanced cross section A cross section that has been drawn so that the structures depicted satisfy certain geometric rules. Balancing is usually applied to sections that are drawn through a thrust belt (Dahlstrom, 1969; Hossack, 1979). The goal is to produce a section that accurately depicts the geometries of buried structures, at depths where direct information about the geometries is sparse. Balancing involves comparison between two cross sections: (1) an undeformed section that represents the lengths and thicknesses of strata as they are inferred to have been before the deformation, and (2) a deformed section that shows structures as they are inferred to be now. Mass must be conserved between the two sections. This required conservation of mass, and appropriate assumptions, are expressed as geometric rules that the deformed section must satisfy. One rule that is commonly used is that the cross sectional area of each stratigraphic unit must be conserved so that the deformed section can be pulled apart into an undeformed section without producing gaps or overlaps. Another common rule specifies that bed length across the deformed section must be the same for all units that have buckled and faulted but have not flowed internally. A deformed section that matches the corresponding undeformed section, according to the geometric rules and within the limits of available data, is said to be balanced. Subsurface data are usually sparse, and different assumptions can produce different geometric rules, so several different deformed sections may balance a particular undeformed section. Thus, balancing cannot guarantee that a particular deformed section is correct. However, if a deformed section is unbalanced, and if the imbalance cannot be explained, then the deformed section must be wrong. Thus, at the least, balancing can narrow the range of possible geometries of subsurface structures. Balancing by hand is tedious, so some workers balance their sections only partly, for example by balancing bed lengths but not cross sectional areas, or by balancing shallow structures but not deep ones. The recent advent of interactive computer programs is a great aid to balancing.

confidence ellipsoid A measure of the locational uncertainty of a calculated earthquake hypocenter. The size, shape, and orientation of the ellipsoid are measures of the uncertainty of the latitude, longitude, and depth of the calculated hypocenter. The equation of the ellipsoid is calculated so that there is a specified probability, usually 0.68, 0.90, or 0.95, that the true hypocenter falls within the ellipsoid (J. W. Dewey and D. W. Gordon, written commun., 1980; Dewey and Gordon, 1984). Then the ellipsoid is referred to as, for example, the 0.90 confidence ellipsoid or the 90-percent confidence ellipsoid. Most confidence ellipsoids are tilted, so that their semimajor axes are not vertical or horizontal. The projection of the ellipsoid onto a plane, such as a horizontal map or a vertical cross section, is a confidence ellipse. Equivalent terms are error ellipsoid and error ellipse, and hypocenter ellipsoid and epicenter ellipse.

dB Decibels.

ERH, ERZ Measures of the locational uncertainty of a calculated earthquake hypocenter, in terms of the confidence ellipsoid (Lee and Lahr, 1975). ERZ is half the length of a vertical line that spans the ellipsoid and passes through its center. ERH is half the greatest width of a horizontal cross section through the center of the ellipsoid. Thus, ERZ measures depth uncertainty, and ERH, map or epicentral uncertainty. Because most confidence ellipsoids are tilted, ERZ and ERH generally do not correspond to semiaxes of the ellipsoids.

felt area The size of the area over which an earthquake was sensibly felt or otherwise noticed by humans.

hazard or **risk** These concepts have been expressed as verbal descriptions, numerical values, and probabilities that a specified value will be exceeded in a specified time at a specified site. However, one distinction is common and we shall follow it here: Hazard refers to the geologic effects of an earthquake; whereas, risk refers to its societal effects (Hays, 1979; Earthquake Engineering Research Institute Committee on Seismic Risk, 1981, 1984). Because we do not use the term "risk" herein, that distinction will suffice for our needs.

hypocenter The three-dimensional location of an earthquake within the Earth, usually specified by latitude, longitude, and depth below some datum.

intensity A standardized, qualitative measure of the effects of an earthquake at a particular place. Intensity is expressed as a Roman numeral and determined for that place by comparing the earthquake's effects there with written lists of effects on man-made structures, natural systems, and human behavior. The most commonly used list of effects gives Modified Mercalli Intensity (MMI) with 12 divisions, I-XII. For example, people usually feel earthquakes of MMI=II or III or greater, and structural damage to buildings usually does not occur below MMI=VI or VII (Wood and Newmann, 1931).

interference structure A complex structure formed when a younger structure deforms an older one. Examples include folded folds, folded faults, faulted folds, and faulted faults. Two interfering structures bear an interference relationship to each other.

magnitude A measure of the strength or size of an earthquake, usually expressed in Arabic numerals. Magnitude is a measure of the energy released by the earthquake as elastic waves. Two earthquakes whose magnitudes differ by 1 differ by a factor of 10 in the amplitude of their elastic waves and by a factor of about 30 in the energy released. Various ways of calculating magnitudes produce slightly different results, because different seismic waves of the seismograms are measured. The types of magnitudes used in this report are M (unspecified type of magnitude), M_L (Richter magnitude, measured on a specific type of seismograph), m_b (magnitude calculated from body waves of distant earthquakes), M_S (magnitude calculated from surface waves of distant earthquakes), and m_{bLg} (also written $m_b(L_g)$; similar to m_b but modified to use a short-period type of surface wave; for application to Eastern earthquakes).

microearthquake A small earthquake not felt by humans. In this report, an earthquake with a magnitude of less than or equal to 2.

seismogenic fault A fault that generates earthquakes.

seismic zone This term has two different meanings. One refers to engineered structures, and is "a generally large area within which seismic-design requirements for structures are constant" (Earthquake Engineering Research Institute Committee on Seismic Risk, 1984). The other meaning refers to a volume or area defined by a group of hypocenters or epicenters that are presumed to be related because they are considered to form a single spatial pattern. We will use seismic zone in its second meaning, as in "seismicity of the Giles County seismic zone."

structure We will use this term in accordance with the usage common in structural geology: "1. The way in which a rock, a rock-mass, or a whole region of the earth's crust is composed of its component parts: the form and mutual relations of the parts of a rock * * * 2. Structural discontinuity of any kind occurring in rock bodies" (Dennis, 1967, p. 145). For example, the structure of the Valley and Ridge province of the Appalachians can be described as a complex of thrusting-related folds and mostly shallowly dipping faults. Also, a discontinuity in rock properties, such as an igneous or erosional contact that cuts across beds, is a structure. Such a discontinuity could concentrate enough stress to cause seismicity at a point, line, or surface, but the presence of such a structure is not in itself grounds for inferring the presence of a fault or the likelihood of seismicity. We will not use structure as a synonym for fault.

velocity model A mathematical representation of vertical changes in the velocity of seismic waves through the Earth. In the models of this report, the Earth is represented by several horizontal layers overlying an infinite half-space. Each layer or half-space of the model has its own constant velocity. The velocity layers usually bear little relation to the stratigraphic units of the geologist, because most of these units are too thin, differ too little in velocity to be resolvable by the seismic wavelengths of interest, or both. For example, most velocity models used in the Southeastern United States have the several kilometers of sedimentary rock represented by one to three velocity layers, and the entire velocity model typically has three to five layers to represent the crust and upper mantle.

GEOLOGIC SETTING

INTRODUCTION

This report uses the widely accepted division of the Appalachians into structural provinces, each characterized by a distinctive combination of exposed stratigraphy, structural styles, and degree of metamorphism. Geologic aspects of this report deal with structures and stratigraphy that are exposed at ground level, and with structures that are inferred to exist in the subsurface. Accordingly, the division into structural provinces is more appropriate for our purposes than is the older division into physiographic provinces, which we will not use. For more detailed descriptions of the structural provinces than are appropriate here, see Rodgers (1970).

From northwest to southeast in southwestern Virginia and adjacent States, the structural provinces are the Appalachian Plateau, Valley and Ridge, Blue Ridge, and Piedmont Plateau provinces. The LANDSAT image on the cover shows the characteristic topographic expressions of the rock types, structural styles, and levels of metamorphism that characterize each province.

The strongly defined, linear ridges that cross the center of the cover image from southwest to northeast identify the western Valley and Ridge province. The ridges are limbs of breached anticlines, some cut by southeast-dipping thrust faults. The ridges and intervening valleys expose unmetamorphosed sedimentary rocks. Most are sandstones, siltstones, shales, and some limestones, of middle Paleozoic age. Southeast of the ridges lies the Great Valley or eastern Valley and Ridge province. Structural style is similar to that in the western part of the province, but the rocks exposed in the Great Valley are mostly unmetamorphosed to slightly metamorphosed limestones and dolomites of early Paleozoic age. In the humid climate of this region, these rocks weather to form flat to rolling farmland, mostly light colored in the image. Here and there in the Great Valley are steep ridges, supported by middle Paleozoic sandstones preserved in synclinoria.

Because the topography of the Valley and Ridge province reflects structural style and orientation so clearly, the province best shows the junction of the central and southern Appalachians. In the north-central part of the image, north-northeast-trending structures of the central Appalachians interfinger with and are replaced southwestward by east-northeast-trending structures of the southern Appalachians.

The Blue Ridge and Piedmont Plateau provinces lie southeast of the Valley and Ridge province. Their topographic expressions are varied in the image, but the expressions are mostly finely lineated, with subdued topography and structural grain. Exposed in these two provinces are metamorphic rocks of diverse Precambrian and Paleozoic ages, origins, and degrees of metamorphism.

Northwest of the Valley and Ridge province lies the Appalachian Plateau province, a rugged, heavily dissected region that exposes mostly sandstones, shales, and coals of late Paleozoic age. Most beds lie flat in the northwest corner of the image, but bed dips increase southeastward toward the Valley and Ridge province. Here and there in the easternmost Appalachian Plateau province are anticlines whose amplitudes and limb dips give them a structural style between the tighter folding of the Valley and Ridge province and the gentler folding of the Appalachian Plateau province. Where such anticlines are present, topographic expression alone makes the Valley and Ridge province seem unusually wide. Examples occur in the northern third and at the western edge of the cover image. However, most workers place such structures in the Appalachian Plateau province because of their structural style and the age of the exposed rocks. We follow such usage here, using the boundary between the two provinces as that of Rodgers (1970, p. 5–8). At the latitude of Giles County, the province boundary is so sharp structurally, stratigraphically, and topographically that it forms most of the northwestern county line and the West Virginia–Virginia border.

Thus, Giles County, Va., lies at the western edge of the Valley and Ridge province (fig. 2). The Giles County locale, defined previously as the area within about 50 km of Pearisburg, lies mostly in the Valley and Ridge province but overlaps into the Appalachian Plateau province. Ground elevation ranges from about 0.6 km to about 1.2 km above sea level but is about 1 km in most places. Figure 2 shows locations of the five seismograph stations of the Virginia Polytechnic Institute seismic network that form a subnetwork in the Giles County locale.

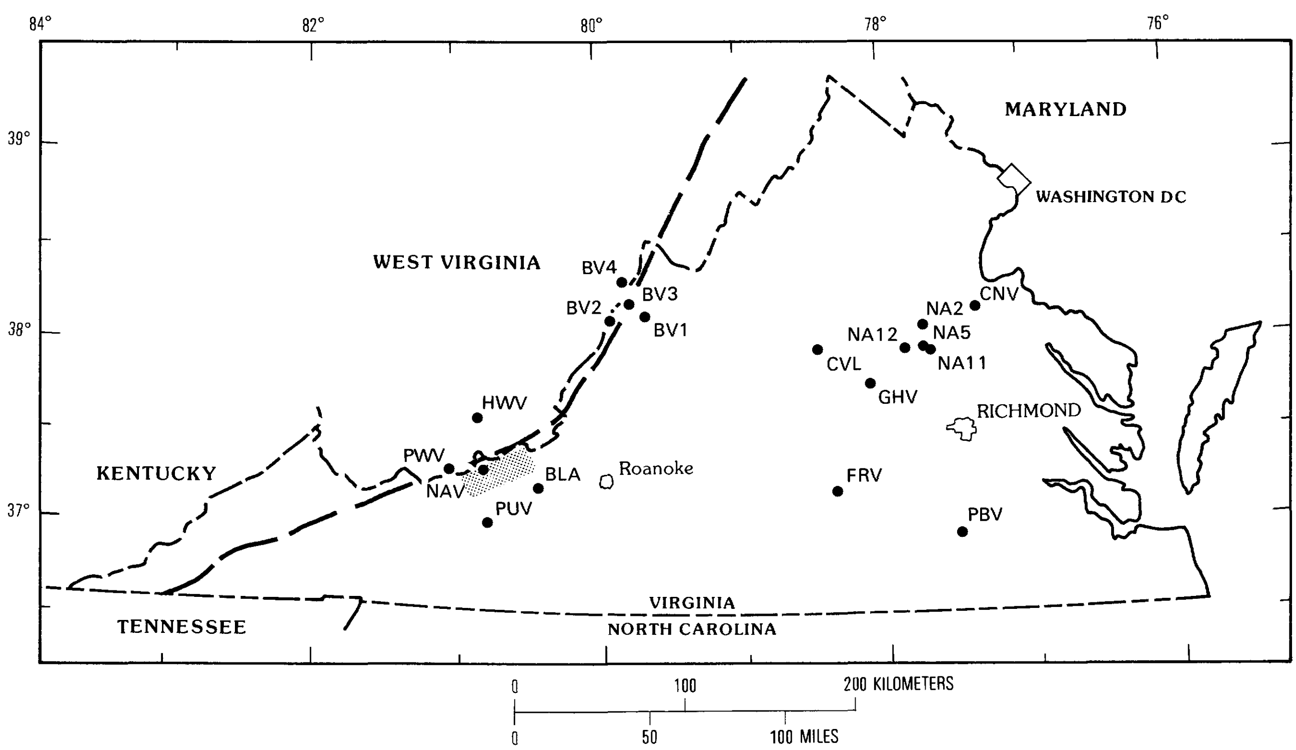


FIGURE 2.—Map showing the Virginia Polytechnic Institute Seismic Network. Modified from Bollinger (1981a) with permission. Locations of individual seismograph stations are shown by solid circles. Stations are identified by their three- or four-character formal names. Dashed line divides Appalachian Plateau (on northwest) and Valley and Ridge (on southeast) provinces (Rodgers, 1970, pl. 1A). Solid lines are State boundaries. Shaded area defines Giles County.

STRATIGRAPHY AND STRUCTURE

Most of the stratigraphic units and structures that are exposed in and near the Giles County locale are not pertinent to an interpretation of seismicity under that locale. The reason is that all pre-Mesozoic near-surface rocks from the eastern Appalachian Plateau province southeastward have been thrust to the northwest during Paleozoic deformation, and the earthquakes occur below the deepest known thrust faults. Evidence to support those statements will be given in following sections of this report, but their consequence is that whatever structures cause the earthquakes have nothing obvious to do with exposed structures or stratigraphic units.

Accordingly, descriptions of stratigraphy, structures, and tectonic evolution will be confined to general summaries and to details that will be used in subsequent discussions and arguments. Most structures and stratigraphic units of the Giles County locale and environs will not be mentioned because they have no bearing on our discussion. Also, most of the geological arguments of this report deal with stratigraphic units and structures that, in and near Giles County, remain buried under thrust sheets or have been removed by erosion. Our arguments will draw on evidence from areas where those units and structures are exposed or preserved. Some of those areas are distant from the Giles County locale; for example, up to several hundred kilometers away in western West Virginia. Thus, what may seem like a tendency to ignore the study area arises from the need to determine what is under Giles County, not what is exposed in it.

Stratigraphy and rock type have strongly influenced structural style in the parts of the Valley and Ridge and Appalachian Plateau provinces near Giles County. That influence is summarized in figure 3. Column *B* of figure 3 summarizes the rock types that dominate various parts of the stratigraphic column, as exposed throughout Virginia and West Virginia. The Precambrian basement complex is overlain by the Late Proterozoic to Lower Cambrian basal clastic rocks of the Appalachians. The lower Paleozoic rocks are mostly a thick sequence of carbonate rocks of Cambrian and Ordovician ages. Most of the rest of the preserved sequence is clastic, although limestones occur in the Silurian and Devonian part. Clastic wedges of Ordovician, Devonian, and late Paleozoic ages contain the erosional record of parts of the Taconic, Acadian, and Alleghany orogenic episodes. Mississippian and older rocks are mostly marine; younger rocks are mostly continental. The youngest preserved rocks are Permian, which overlie the Pennsylvanian coal measures of West Virginia and adjacent States.

Column *C* of figure 3 shows what might be termed a structural stratigraphy. The Paleozoic sequence has been thrust northwestward during Paleozoic deformational episodes. The distance of northwestward transport decreases to the northwest, because the thrust masses telescoped as they moved. Thrust faults occur preferentially in shaley, thin-bedded parts of the sequence. Paired arrows show the four places in the stratigraphic sequence in which large thrust faults most commonly formed. The deepest thrusts are found in the Lower Cambrian shales (locally named the Rome Formation). Rocks below the Rome Formation form structural basement, which is not known to have been thrust northwestward. Shallower rocks have ridden northwestward on one or more of the intervals indicated in figure 3, on smaller thrust faults elsewhere in the section, or on both. Which thrust faults were active at a particular time and place varies in complex fashions. Not all thrusts formed or moved at once, and facies changes cause the shaley sequences that localize the thrusts to rise and fall stratigraphically, and to thicken and thin both along and across strike. Such complexity does not alter the important fact that rocks above the Rome

A	B	C
PERMIAN PENNSYLVANIAN	Sandstones, shales, coals	
MISSISSIPPIAN	Sandstones, shales	
DEVONIAN	Shales	⇋
	Sandstones, shales, limestones	
SILURIAN	Shales, limestones	⇋
	Sandstones, shales, limestones	
ORDOVICIAN	Shales	⇋
	Limestones, dolomites	
CAMBRIAN	Shales *	⇋
PRECAMBRIAN	Sandstones, shales, conglomerates, metamorphic and igneous basement	Structural basement

FIGURE 3.—Sketch summarizing the stratigraphy of the central and southern Appalachians in and near Giles County, Va., and its relationship to the dominant structure there. *A*, The age range of units exposed in and near the Giles County locale. *B*, Dominant rock types in various portions of the stratigraphic column. Asterisk identifies the shales of the Lower Cambrian Rome Formation. *C*, Paired arrows identify the stratigraphic and lithologic positions that commonly contain the largest thrust faults.

Formation, including all exposed rocks older than Mesozoic in and near the Giles County locale, have been thrust northwestward.

INDEPENDENCE OF SEISMOGENIC AND EXPOSED STRUCTURE

Interpretations of data from public and proprietary wells and reflection seismic surveys show that the sedimentary rocks are underlain by metamorphic and igneous basement at depths of 10,000–19,000 ft (3–6 km) subsea (section *F-F'* of Cardwell and others, 1968; Shumaker, 1977, and in Negus-deWys, 1979, and Negus-deWys and Renton, 1979; Kulander and Dean, 1978b; Compudepth map of Seay, 1979; Perry and others, 1979; Kelly, 1978). Those depths correspond to 4–7 km below ground level. Accordingly, structures in rocks shallower than 4–7 km are unrelated to deeper structures (for example, see fig. 4 of Perry and others, 1979, and the more generalized section V3 of Roeder and others, 1978). In particular, we know of no reason to suspect any simple relationship between (1) outcropping faults or other obvious aspects of surface geology and (2) structures at the depths of the seismicity in the Giles County locale (at least 5 km, as documented later in table 4).

On a scale of hundreds of kilometers, most large exposed, shallow subsurface and deeper crustal structures in the Appalachians, adjacent craton, and Coastal Plain are roughly parallel to each other. They are roughly parallel because the Atlantic Ocean opened approximately where an older ocean had opened and then had closed to form the Appalachians (Wilson, 1966; p. 28–37 of this report). Parallelism on such a scale is too general to be of much aid in geological interpretation of seismicity within small areas like Giles County and, indeed, may hinder such interpretation because it limits our ability to distinguish structures by their azimuthal orientations. Thus, such parallelism of structures in eastern North America does not affect our conclusion that we expect no simple relationship between surface structures and seismicity in or near Giles County.

Two lines of evidence might appear to conflict with that conclusion, but on examination, do not. First, independence of structure above and below the thrust faults is best established farther north, in the central Appalachians (Gwinn, 1964; Rodgers, 1963). The possibility remains of subtle control of Paleozoic depositional systems by ancient topography created by movement on then-active faults in the underlying basement. Then, the thicknesses of Paleozoic sedimentary units would reflect that ancient topography (Cooper, 1961, p. 100–118, 1964; Thomas, 1982b). Such control is perhaps more likely in the southern than in the central Appalachians although there are clear examples in western West Virginia, near the cratonward edge of the Appalachians (Schaefer, 1979; Shumaker and others, 1979; Donaldson and Shumaker, 1981; Nuckols, 1981). However, Geiser (1977) pointed out that the same sedimentological patterns could be produced by thrust-related anticlines that were growing during deposition of the sediments in question. Thus, any such sedimentological patterns would not necessarily be evidence that basement faults are reflected in surface geology.

Second, J. W. Dewey and D. W. Gordon (written commun., 1980) calculated the location of the 1969 Elgood, West Virginia, earthquake ($m_b = 4.6$, event J of table 7). They obtained a depth of 2.5 km below ground level for the hypocenter. All other reliable hypocentral depths in the Giles County locale are deeper, within the basement. The top of basement near Elgood is at an approximate subsurface depth of 4–5 km, so the Elgood focal depth is apparently well within the sedimentary rocks. However, from J. W. Dewey's and D. W. Gordon's results (written commun., 1980), the vertical semiaxis of the 90-percent confidence ellipsoid about the hypocenter is estimated to be 6 km. Thus, the probability is 0.90 that the depth of the Elgood earthquake was about 8.5 km or less. Furthermore, near Elgood the deepest thrust faults are only from 3 km to about 3.5 km below the surface (fig. 4 of Perry and others, 1979; W. J. Perry, Jr., oral commun., 1980). As much as another 3 km of unthrust sedimentary rocks underlie those deepest thrust faults, separating the faults from the top of metamorphic and igneous basement (Perry and others, 1979). Those sedimentary rocks beneath the thrust faults are structurally part of the basement (fig. 3, column *C*). Thus, the depth calculated by Dewey and Gordon for the Elgood earthquake, taken with the 6-km uncertainty implied by the confidence ellipsoid, is not inconsistent with the earthquake having occurred either in the metamorphic and igneous basement, or in the unthrust sedimentary rocks below the deepest thrust faults. In addition, Herrmann (1979) calculated a depth of 5 km for the Elgood earthquake, using surface-wave data, and Carts (1981) calculated a well-constrained depth of 13.6 km using the computer program HYPO71 (Lee and Lahr, 1975). The associated vertical location error parameter, ERZ, was 1.4 km using a locale-specific velocity model. Thus, of the three depths calculated for the Elgood earthquake, two place it below the thrust structures and the third has too large an uncertainty to contradict such a depth. We, therefore, retain our conclusion that seismicity in the Giles County locale appears to bear no simple relationship to surface geology. Later, we shall consider faults or classes of faults that are known or inferred to exist in the basement and that lack obvious expression in the surface geology, and which thus could be responsible for the observed seismicity.

SEISMICITY OF THE GILES COUNTY, VIRGINIA, LOCALE

NETWORK MONITORING PROGRAM

The Giles County network is a five-station subnetwork of the Virginia Polytechnic Institute's Seismic Network. The subnetwork is capable of detecting and locating accurately microearthquakes in the nearby area (fig. 2, table 2). Its central station is at Narrows, Va. (station call letters NAV), about 6 km west of Pearisburg, Va., and is located in Mill Creek Valley on the east side of Sentinel Point promontory. (See frontispiece.) The subnetwork aperture (greatest distance between stations) is about 60 km. Monitoring was initiated at NAV in October 1977, and the network installation was completed by mid-April, 1978. All stations have short-period (1 Hz), vertical (SPZ) transducers; however, the Pulaski, Va., (PUV) station also has two short-period horizontals (oriented north-south and east-west, and has been operational since early in 1980). Signals from all five stations are telemetered to a central recording facility on the Virginia Polytechnic Institute

TABLE 2.—*Site, instrumentation, and operation information for the Giles County, Va., subnetwork of the Virginia Polytechnic Institute Seismic Network*

[SPZ, short-period vertical seismometer; T_o, free period of seismometer; T_g, free period of galvanometer. Timing System: Systron-Donner Time Code Generator 8120[5]. Direction of motion of records: up on record for up on ground. System response curves (see fig. 4). Two horizontal sensors added at PUV early in 1980. Magnifications listed are for the visual recorders]

Site information

Code	Station name	Latitude (north)	Longitude (west)	Elevation (meters)	Date opened	Foundation geologic age
NAV	Narrows, Va.-----	37.3157°	80.7935°	610	10/77	Ordovician clastics.
PUV	Pulaski, Va.-----	37.0235°	80.8158°	652	2/78	Devonian clastics.
HWV	Hinton, W. Va.---	37.5905°	80.8408°	521	4/78	Mississippian clastics.
PWV	Princeton, W. Va.	37.3348°	81.0488°	820	3/78	Do.
BLA	Blacksburg, Va.--	37.2114°	80.4211°	634	1962	Cambrian carbonates.

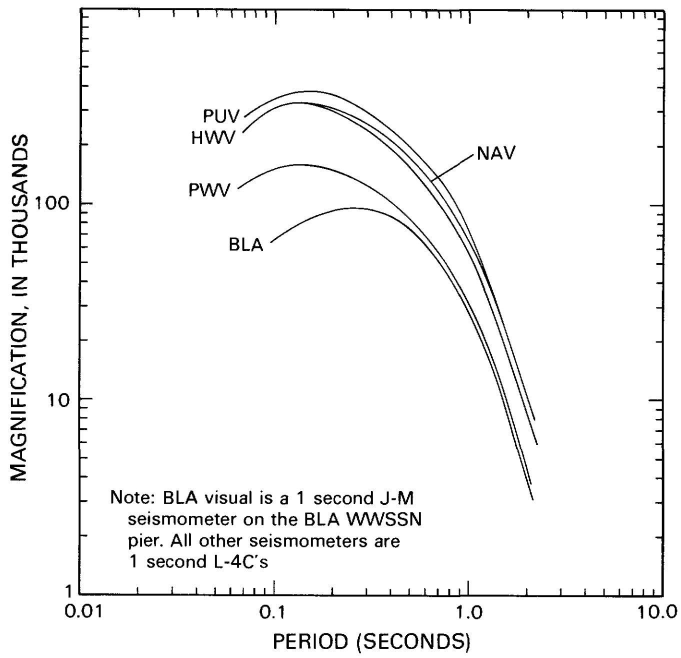
Instrumentation

Code	SPZ seis-mometer	T_o (sec-onds)	T_g[1] (sec-onds)	Type recording[2]	Magnifi-cation at T_o	Maximum magnification
NAV	(L4-C)[3]	1.0	0.1	V,F,T	65K	310K at 0.15 s
PUV	(L4-C)[3]	1.0	.1	F,T	75K	390K at .15 s
HWV	(L4-C)[3]	1.0	.1	F,T	53K	320K at .15 s
PWV	(L4-C)[3]	1.0	.1	F,T	32K	160K at .15 s
BLA	(J-M)[4]	1.0	.1	V,F,T	30K	97K at .30 s

Operation (To June 1, 1980)

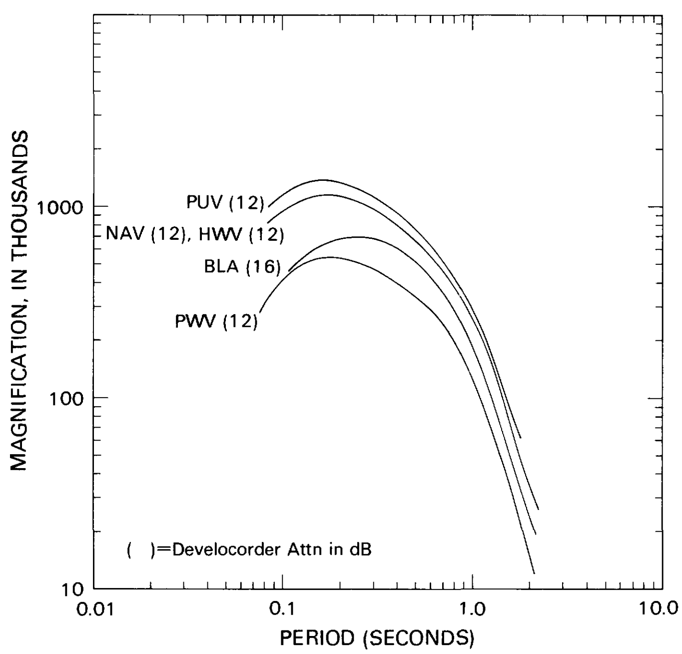
Code	Years of operation	Total down days	Percent down time
NAV	2.60	33	3.5
PUV	2.24	29	3.5
HWV	2.13	4	.5
PWV	2.21	27	3.3
BLA	2.60	1	.1

[1]High-cut filter setting.
[2]V, visual; F, 16-mm film; T, FM magnetic tape.
[3]Mark Products design[5].
[4]Johnson-Matheson design[5].
[5]Use of trade names in this report is for descriptive purposes only and does not constitute endorsement by the U.S. Geological Survey.

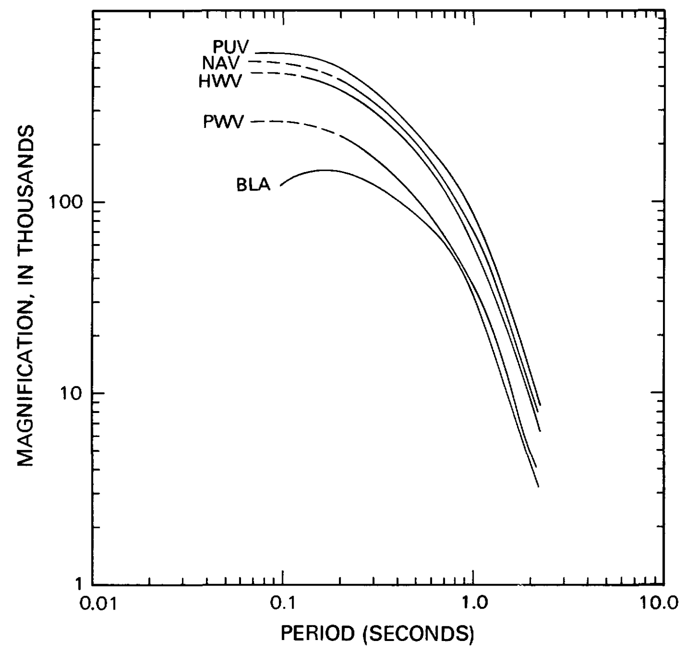


A. MAGNIFICATION CURVES – GILES COUNTY, VIRGINIA NETWORK. VISUAL RECORDER CALIBRATION – JANUARY 1979

Direction of motion (all stations): Up on record=Up on ground



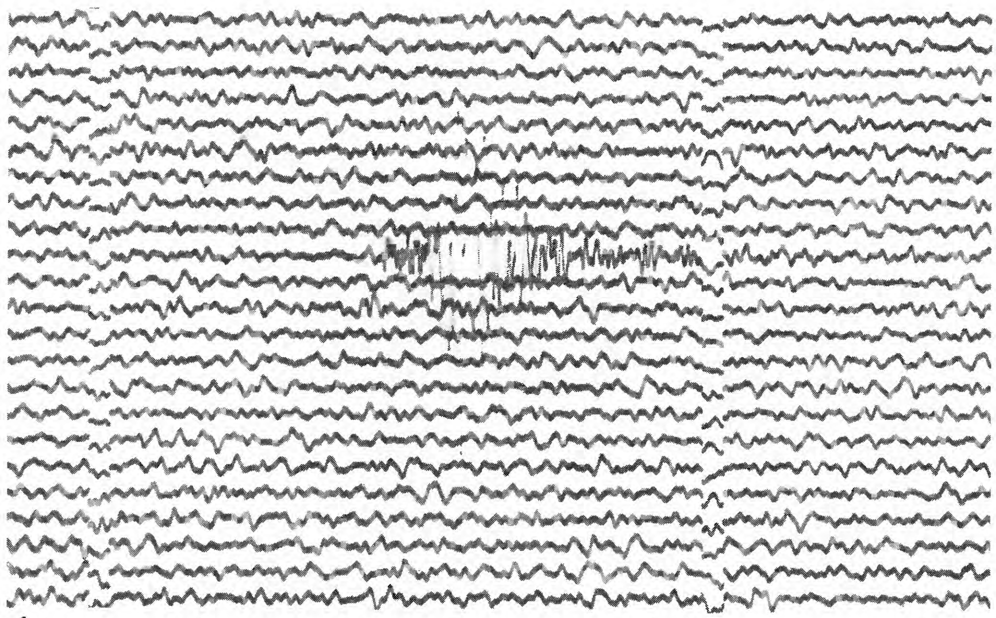
B. MAGNIFICATION CURVES – GILES COUNTY, VIRGINIA NETWORK. DEVELOCORDER CALIBRATION – JANUARY, 1979



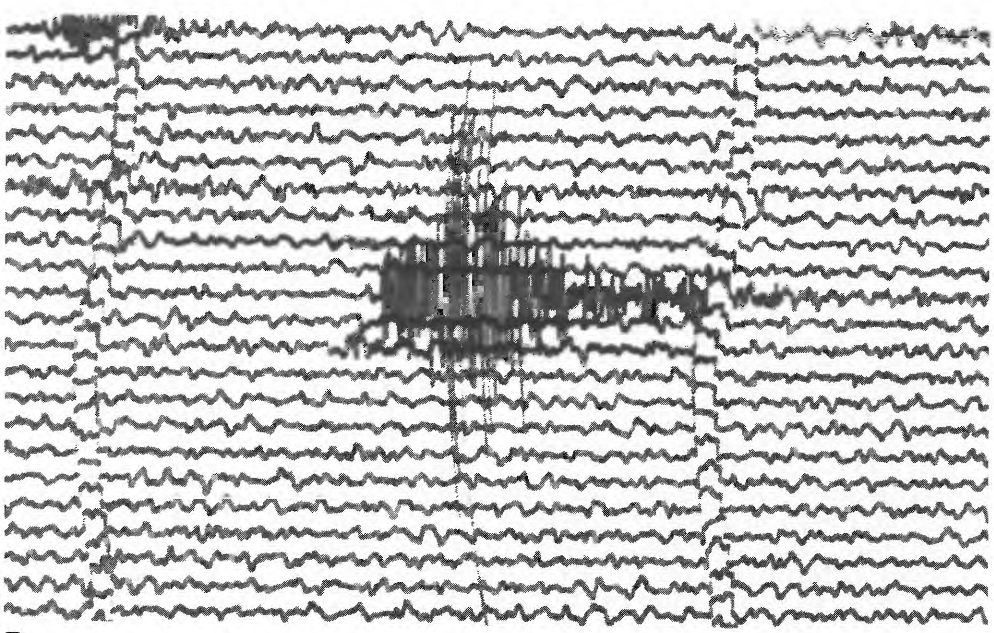
C. MAGNIFICATION CURVES – GILES COUNTY, VIRGINA NETWORK. ANALOG TAPE RECORDER CALIBRATION, JANUARY 1979

FIGURE 4.—Magnification curves for the seismographs comprising the Giles County subnetwork of the Virginia Polytechnic Institute Seismic Network. Seismograph stations located in Virginia (BLA-Blacksburg, PUV-Pulaski and NAV-Narrows) and in West Virginia (HWV-Hinton and PWV-Princeton). The magnification is the amount by which the seismograph magnifies the ground motion. Thus, a magnification of 100,000 (100K) means that a ground motion of 1 micron (0.001 mm) would appear (be magnified) on the seismogram as 10 cm (100 mm). The calibration of the seismographs necessary to determine these magnification curves was performed during January 1979. *A*, Visual (pen-and-ink) recorders; seismograph polarities set so that an upward motion of the seismogram trace corresponds to an upward motion of the ground beneath the seismometer. *B*, 16-mm-film recorder. Develocorder Attn in dB gives the attenuation switch setting on the 16-mm-film recorder. Magnifications shown are for the use of a film viewer with its own magnification of 20 times; *C*, Analog tape recorder-playback system. See table 2 for general network information.

campus where they are recorded on 16-mm film and on frequency-modulated (FM) analog magnetic tape. Signals from the central station (NAV) are also recorded on a visual recorder (pen-and-ink) and as an additional, low-gain (30-decibel (dB) reduction) channel on the tape recorder. Seismograph magnifications at the individual stations range from 30 to 300 K at 1 Hz, depending on station and recorder mode and are specified by individual magnification curves. (See fig. 4.) The frequency passband for all recording channels is set by filters at 1–10 Hz. Average microseismic levels, as measured on the 16-mm film records, range from 1 to 60 nanometers at 0.6–3.4 Hz (Sibol, 1980; Appendix A of this report). Figure 5 shows seismograms from a magnitude 1.6 event that occurred near Narrows, and which was recorded on both the BLA WWSSN SPZ (short-period vertical seismograph and the network BLA SPZ (1–10 Hz passband). The increased efficiency of microearthquake



A



B

FIGURE 5.—Sample seismograms for a microearthquake. Two different, short-period, vertical seismograms for the same microearthquake that occurred near Narrows, Va. (January 28, 1978; event no. 32, table 4; magnitude=1.6; minute marks every 60 mm on original seismograms. Scale here is 1 mm on the figure equals about 1.4 mm and 1.4 s on the original seismogram). Both seismometers located on the same pier at Blacksburg, Va. *A*, BLA WWSSN SPZ: magnification is 50K at 1 Hz and 4.5K at 10 Hz. *B*, BLA network SPZ visual: magnification is 28K at 1 Hz and 65K at 10 Hz. (See fig. 4*A*.) Note the increase in signal-to-noise ratio achieved by the increased magnification of the higher ground frequencies by the network station.

recording provided by the network passband is apparent in the figure. That increase is accomplished by emphasizing the higher (5–10 Hz) Earth frequencies.

The capability for detection and location of microearthquakes by the entire Virginia network, according to Tarr (1980), is illustrated in figures 6 and 7 (threshold

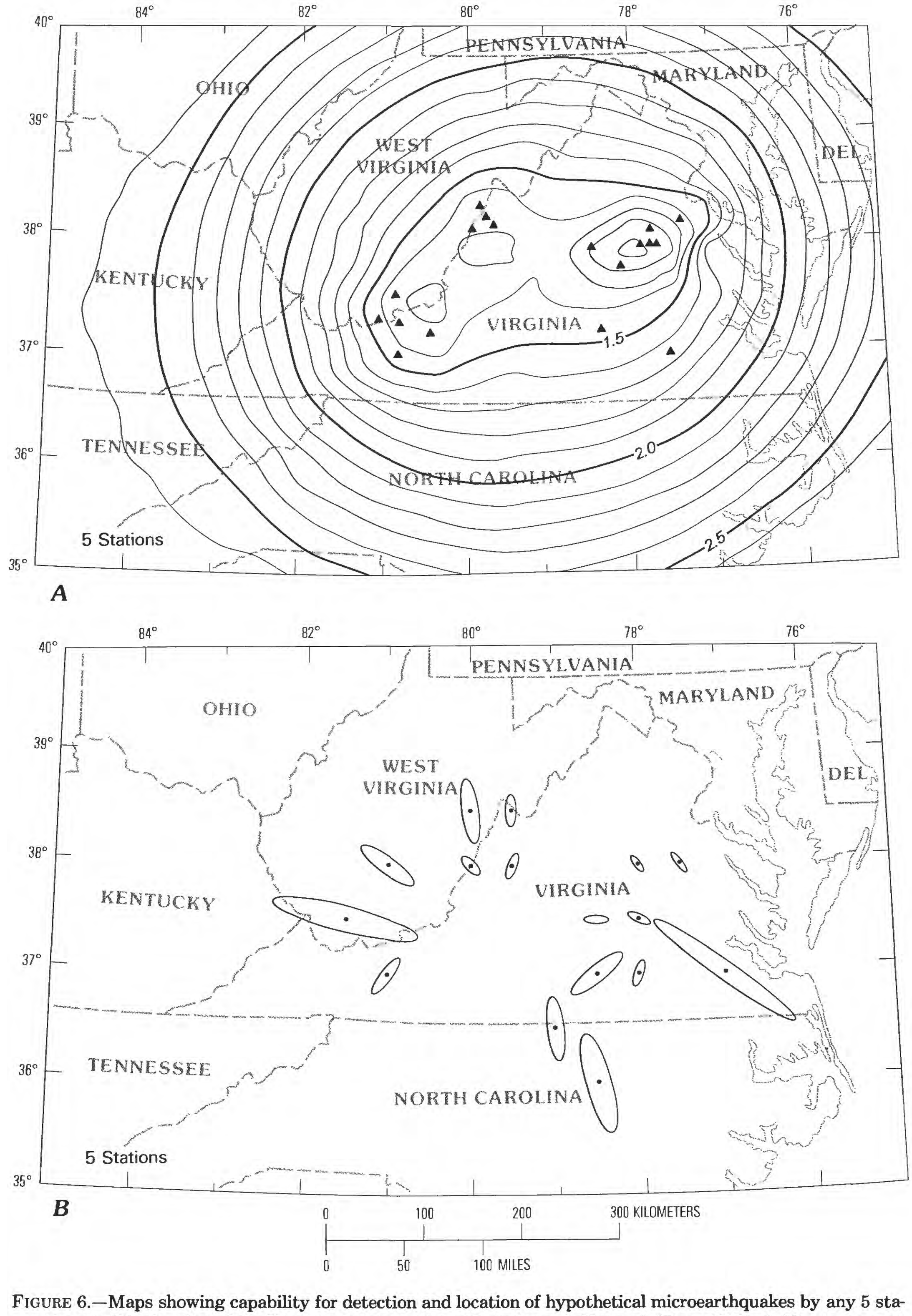


FIGURE 6.—Maps showing capability for detection and location of hypothetical microearthquakes by any 5 stations (solid triangles) of the Virginia Polytechnic Institute Seismic Network (Tarr, 1980). *A*, Ninety-percent probability threshold m_b magnitudes for detection by five or more stations. Contour interval 0.1 m_b unit. *B*, Ninety-percent confidence location ellipses, on a 0.5° latitude and longitude grid, for events detected by five or more stations. Ellipses are calculated for each 0.5° grid point but are not plotted if their semimajor axes are greater than 100 km or if their 95-percent confidence intervals on the focal depth are greater than 100 km. Interpolate only between adjacent grid points; do not extrapolate to grid points at which no ellipses are shown.

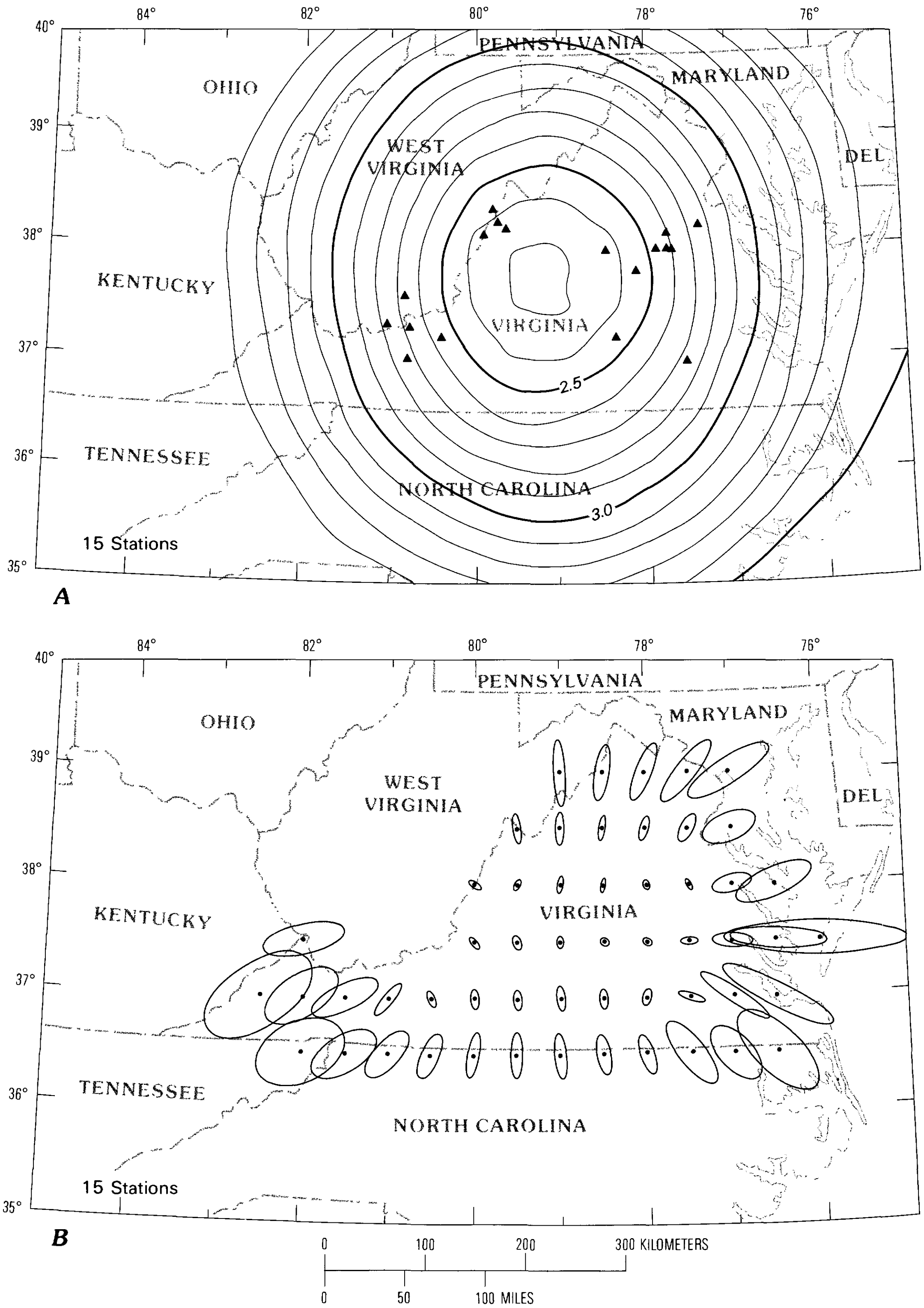


FIGURE 7.—Maps showing capability for detection and location of hypothetical microearthquakes by any 15 stations (solid triangles) of the Virginia Polytechnic Institute Seismic Network (Tarr, 1980). *A*, Ninety-percent probability threshold m_b magnitudes for detection by 15 or more stations. Contour interval 0.1 m_b unit. *B*, Ninety-percent confidence location ellipses, on a 0.5° latitude and longitude grid, for events detected by 15 or more stations. Ellipses are calculated for each 0.5° grid point but not plotted if their semimajor axes are greater than 100 km or if their 95-percent confidence intervals on the focal depth are greater than 100 km. Interpolate only between adjacent grid points; do not extrapolate to grid points at which no ellipses are shown.

body wave (m_b) magnitudes and 90-percent confidence ellipses for detection by 5 or more and 15 or more network stations). Note that inside the five-station Giles County network (see fig. 2), detection is complete down to a magnitude somewhat less than 1.5. Figure 8 shows the 90-percent confidence ellipses for $m_b=2$ and $m_b=3$ events detected by five or more network stations. For the Giles County locale, the location capability is seen to be quite good (small error ellipses).

Event size for locally recorded microearthquakes is determined by a duration magnitude relationship established for the Virginia Polytechnic Institute network by Viret (1980; See Appendix B, this report). For larger events, at distances greater than about 45 km, Nuttli's (1973) m_b formulas that use the short-period surface waves (*Lg* phase) are used.

A crustal velocity model for the Giles County network was determined by Moore (1979). He used conventional refraction techniques with local quarries and regional earthquakes serving as seismic sources. He also used a modification of the classical tripartite technique, perturbed to account for wave-front curvature of signals from regional quarry and mine blasts (Chapman, 1979), as an aid in determining the local velocity structure. Moore (1979) obtained two- and three-layer crustal models. A comparison of the error statistics estimated for the hypocentral locations derived from those as well as other available velocity models (Carts, 1980; Appendix C, this report) indicated that Moore's three-layer model, TPM2 (table 3), gave the smallest error estimates for hypocentral parameters. That velocity model has been used throughout this investigation.

ANALYSIS OF NETWORK EVENTS FROM JANUARY 1978 THROUGH DECEMBER 1980

Using the TPM2 velocity model, hypocenters were recalculated using HYPOELLIPSE (Lahr, 1980) for all the seismic events that had occurred since the beginning of network operation. Twelve of these events occurred in the area shown in figure 9. The reductions in the hypocentral errors, as compared to their pre-TPM2 values, were substantial, and 8 of the 12 epicenters (table 4, nos. 32, 33, 35, 37, 38, 46, 58, 63) coalesced to form a northeasterly trending alignment approximately 10 km in width and some 45 km in length (fig. 9). Four epicenters (table 4, nos. 34, 39, 40, 60) lie off the alignment and are interpreted not to have occurred in the Giles County seismic zone but to be part of the background seismicity of the surrounding region. The depth distribution of the 8 foci depicts a nearly vertical zone that extends from about 5 to 26 km in depth (table 4). These rough dimensions of 45 × 10 × 21 km (length × width × depth) define a seismic zone that is tabular (as opposed to planar or volumnar) in configuration.

TESTS OF THE SEISMIC ZONE

Because the seismic zone is defined by so few foci, evidence that is more objective than the simple visual impression of figures 9 and 13 is required. Appendix D contains discussions of statistical tests and other procedures that provide such evidence. The results of those tests and procedures allow us to conclude that the tabular zone is not random and that we have correctly estimated its orientation.

In addition to use of the hypocenters' statistical error measures to specify the geometry of the Giles County microearthquake zone, another form of testing can be done. By locating known quarry or construction blast sites from the arrival times of their *P* and *S* waves at the network stations and then comparing those calculated locations with the actual locations, the locational capability of the Giles County network can be demonstrated. Thus, the procedure is to pretend that the quarry explosion is an earthquake at an unknown location. Actually, only the epicenter (the horizontal coordinates of latitude and longitude) is tested in such a procedure, because the blasts are only at the surface and not at the deeper earthquake focal depths. But, if the hypocenters determined from the blast data indicate the correct very shallow focal depths, then this is evidence that the velocity model is suitable.

Such a test of the Giles County network and the velocity model (TPM2) was performed. Blasting for a highway bypass being constructed around Pearisburg, Va., was first monitored during December 1979 and then again during May 1980 as a confirming experiment. HYPOELLIPSE (Lahr, 1980) locations were calculated using only network *P* and *S* arrival times. Next, the actual blast locations were spotted on 7.5-minute topographic maps by the shooter. Figure 10 shows the blast locations (designated as A, B, C) and their HYPOELLIPSE locations. Tables 5 and 6 give the small epicenter errors (0.5, 0.9, 2.0 km), and they also show that, although there was lower accuracy in the focal-depth determinations, all determinations that were started well below the surface tended to become shallower than their starting trial focal depths. We interpret the results of these tests to indicate that our earthquake locational capability within the Giles County network is excellent. Blast C, which gave the largest error and the largest uncertainty, consisted of a significantly smaller explosive change than the other two blasts (A and B). The network signals of blast C as a result were not as clear (smaller signal to noise ratio), and thus its calculated location was

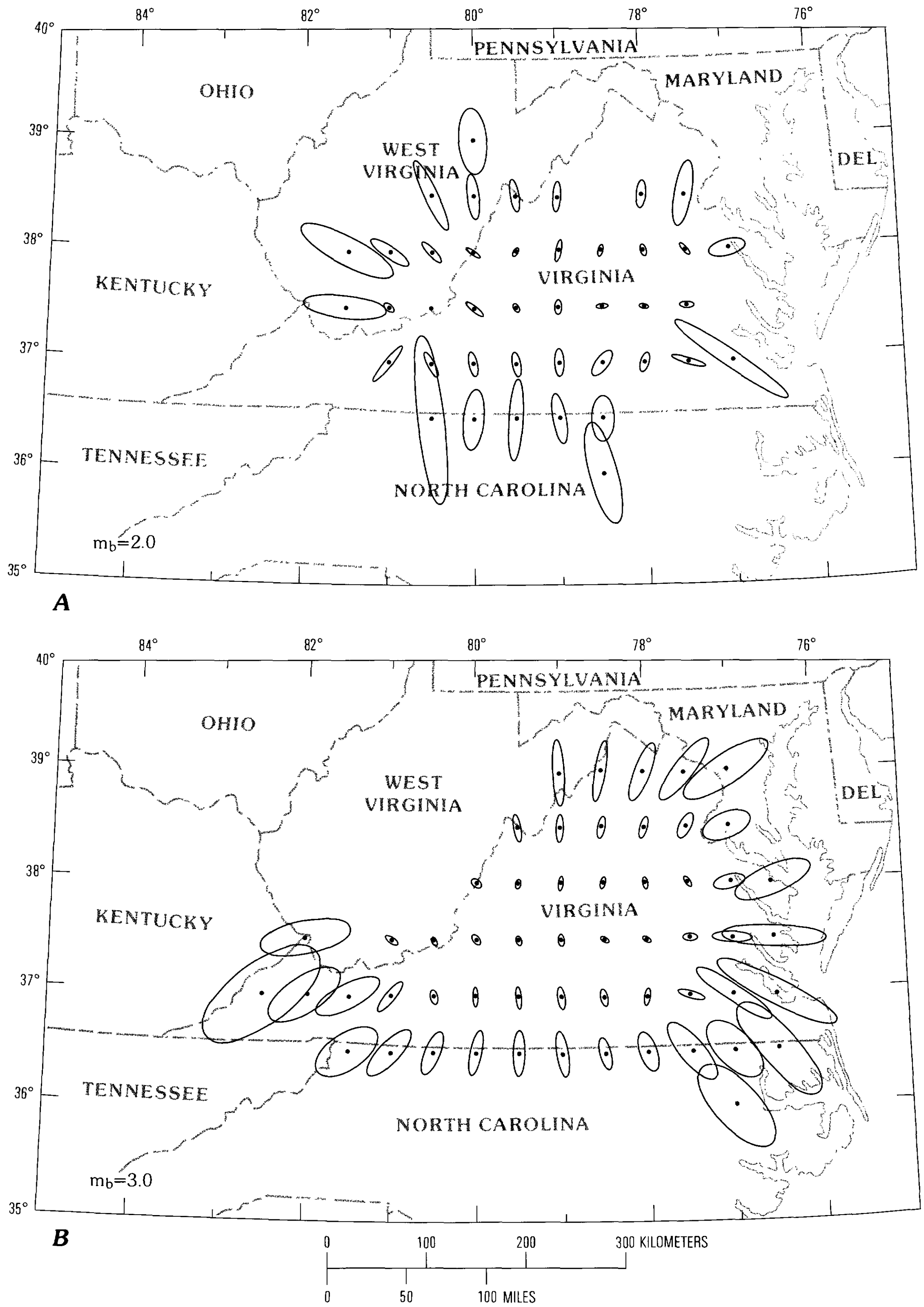


FIGURE 8.—Maps showing 90-percent confidence location ellipses. Ellipses are on a 0.5° latitude and longitude grid, for magnitudes m_b=2.0 (in *A*) and m_b=3.0 (in *B*) hypothetical events detected by five or more of the Virginia Polytechnic Institute Seismic Network stations (Tarr, 1980). Ellipses are calculated for each 0.5° grid point but are not plotted if their semimajor axes are greater than 100 km or if their 95-percent confidence intervals on focal depth are greater than 100 km. Interpolate only between adjacent grid points; do not extrapolate to grid points at which no ellipses are shown.

TABLE 3.—*Velocity model (TPM2) developed for the Giles County, Va., locale by Moore (1979)*

[km, kilometers; km/s, kilometers per second]

Depth (km)	P velocity (V_p, km/s)	S velocity (V_s, km/s)	V_p/V_s
0	5.63	3.44	1.64
5.7	6.05	3.52	1.72
14.7	6.53	3.84	1.70
50.7	8.18	4.79	1.71

expectably not as certain as those with the larger explosive charges.

An additional and important corroboration of the entire northeast-striking zone of microearthquakes was obtained from J. W. Dewey and D. W. Gordon (written commun., 1980). As part of their project to use Joint-Hypocenter-Determination (JHD) techniques (Dewey, 1971) to relocate historical Eastern United States earthquakes, they had relocated six events in the Giles County locale (table 7 and fig. 11). These were all events that were sensibly felt by people ($2.1 \leq M \leq 4.6$) and that occurred between 1959 and 1976, which is prior to the installation of the Giles County seismic network. Four of those six earthquakes relocated directly (within locational uncertainties) on the northeast-striking zone (figs. 11–15; note that the location of station NAV serves as a visual key from one figure to the next).

With the addition of the Dewey and Gordon (written commun., 1980) results, the definition of the Giles County seismogenic zone consists of 12 earthquakes that span four orders of seismic magnitude ($0 \leq M \leq 4$), span some 20 years of occurrence (1959–80), and have locations that were determined by two different research projects. Our judgment is that this constitutes a strong

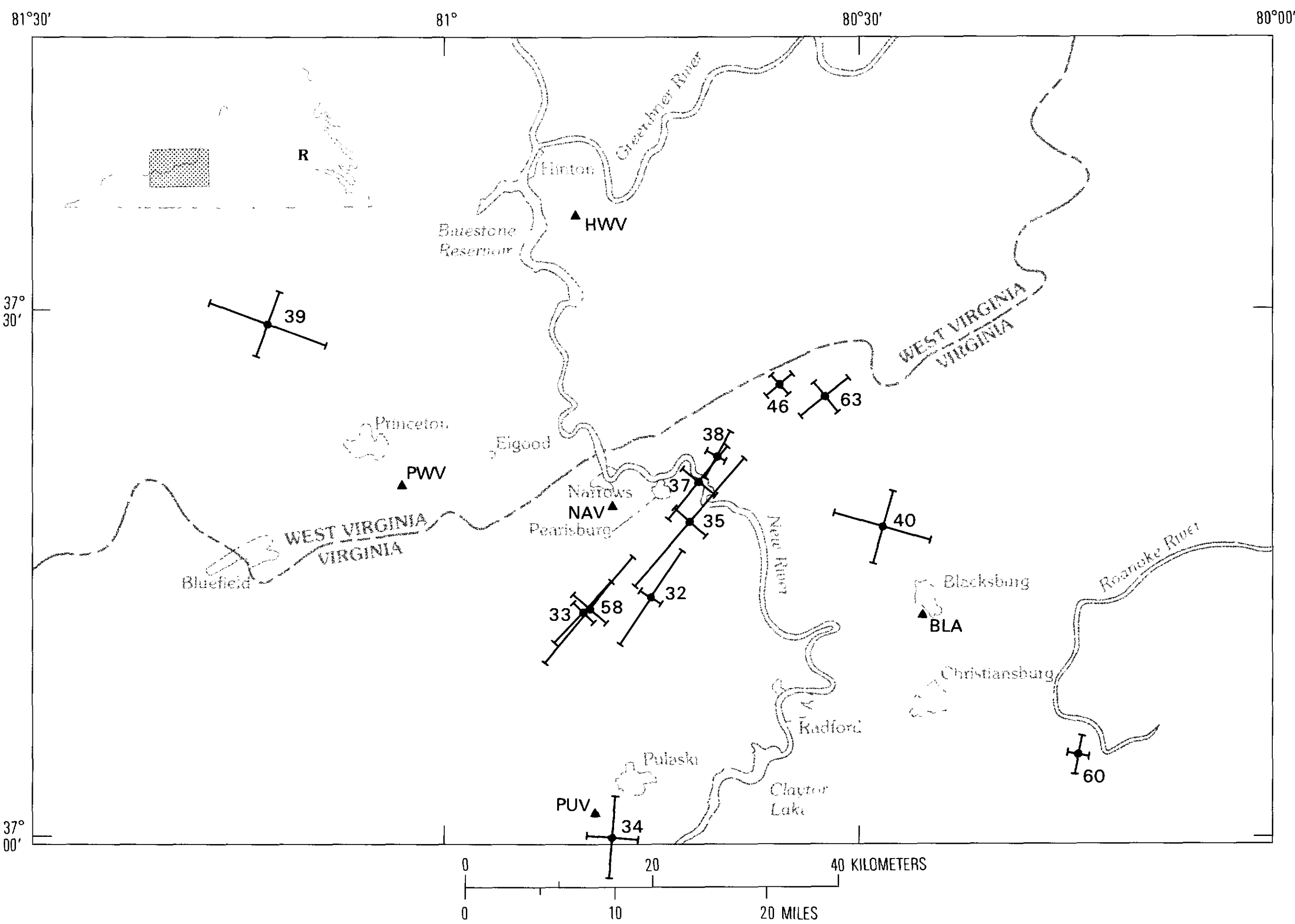


FIGURE 9.—Epicenter map for microearthquakes located with data from the Giles County, Va., subnetwork. Event identification numbers refer to the listing given in table 4. Sixty-eight-percent confidence-ellipsoid axes plotted at each epicenter (Lahr, 1980). Network seismic stations shown by solid triangles with three-letter codes. Inset map shows area of this figure and location for Richmond (R).

TABLE 4.—*Chronological listing of earthquakes that occurred subsequent to 1977 in the Giles County, Va., locale and were located using network data and the HYPOELLIPSE program*

[km, kilometers; s, seconds. Location within 50 km of Pearisburg, Va. HYPOELLIPSE from Lahr, 1979]

Event No.	Date			Origin time (UTC)			Latitude (north)	Longitude (west)	Depth (km)	Magnitude[1] (M_D)	RMS[2] (s)	Projection[3]					Quality[4]
												Horizontal				Vertical	
	Year	Month	Day	Hours	Minutes	Seconds						Length (km)	Trend (degrees)	Length (km)	Trend (degrees)	Length (km)	
32	1978	Jan.	28	23	13	23.4	37°13.68'	80°44.80'	4.5	1.6	0.10	1.3	-56	5.9	34	3.0	C
33	1978	May	10	4	19	9.6	37°12.80'	80°49.82'	26.2	.3	.09	1.5	-46	4.4	44	3.0	B
34	1978	May	25	8	30	25.1	37°00.01'	80°47.65'	12.1	1.5	.23	2.7	-86	4.3	4	3.8	B
35	1978	June	1	1	33	1.0	37°17.99'	80°41.98'	17.3	-0.2	.17	2.1	-49	8.8	41	9.1	C
37	1978	July	28	8	39	40.7	37°20.22'	80°41.41'	11.8	.6	.27	2.2	-51	4.9	39	8.1	C
38	1978	Aug.	30	2	19	38.2	37°21.71'	80°40.06'	8.4	.5	.09	1.0	-62	3.1	28	6.4	C
39	1978	Sep.	14	19	37	6.6	37°29.22'	81°12.80'	9.9	-0.4	.17	3.6	20	6.6	-70	17.4	D
40	1978	Oct.	14	1	50	51.0	37°17.68'	80°28.03'	20.1	.3	.06	3.8	16	5.3	-74	17.2	D
46	1980	Feb.	18	3	58	55.3	37°25.78'	80°35.54'	13.0	1.1	.25	1.2	-41	1.7	-131	3.6	B
58	1980	Oct.	9	1	47	1.1	37°13.01'	80°49.32'	23.5	-0.2	.25	2.3	-50	7.2	40	4.9	C
60	1980	Oct.	14	1	20	4.6	37°04.69'	80°13.82'	11.0	1.7	.35	1.1	-77	2.0	13	3.1	B
63	1980	Dec.	2	7	47	38.2	37°25.08'	80°32.25'	12.2	.4	.34	2.0	-39	3.2	-129	7.4	C

[1]Average network magnitude: M_D=-3.38+2.74 log (D) where D=average duration (seconds) at network stations from the onset of the P wave until return of vibrations to background microseismic level.

[2]Root-mean-square error of the travel-time residuals (observed seismic-wave traveltime minus calculated seismic-wave traveltime).

[3]Projection onto the Earth's surface (horizontal) and onto a vertical plane of the 68-percent confidence ellipsoid on the hypocentral coordinates. These projections are specified by giving the lengths and trends (plus-and-minus from north) of the semimajor and semiminor axes along with the length of the vertical semiaxis.

[4]Quality factor according to HYPOELLIPSE (Lahr, 1979): Lengths and azimuths of the axes of this ellipse are calculated as described in footnote 3. The greatest vertical deviation of the ellipsoid from the hypocenter is also calculated. Then a quality is calculated based on the largest of these three distances according to the following criteria:

Quality	Largest distance
A	Less than or equal to 2.5 km
B	Less than or equal to 5.0 km
C	Less than or equal to 10.0 km
D	Greater than or equal to 10.0 km

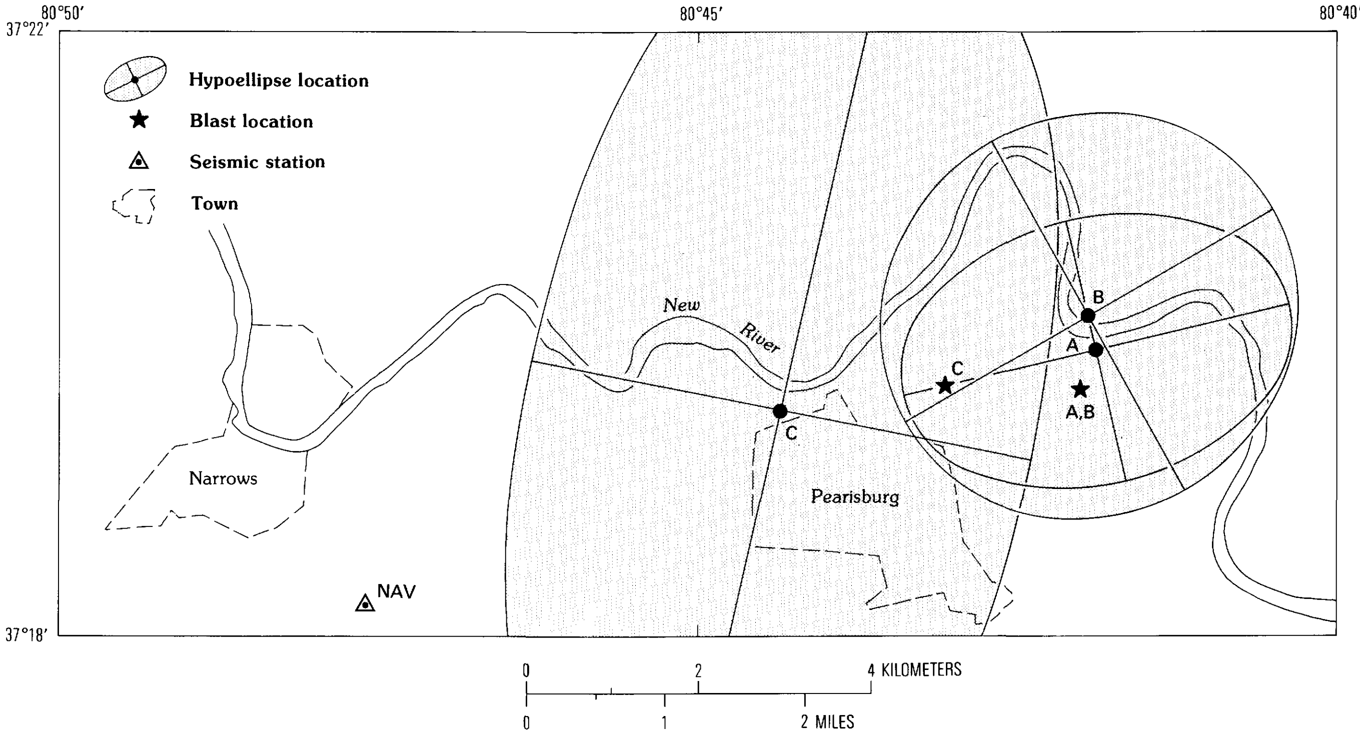


FIGURE 10.—Map showing a comparison of actual blast locations with those calculated using data from the Giles County, Va., subnetwork. Actual locations of blasts A, B, and C shown by stars. Computed locations shown by dots at the center of the error ellipses (projection onto a horizontal plane of the error ellipsoid). Location of Narrows, Va., seismic station (NAV) shown by open triangle with center-dot symbol.

TABLE 5.—*HYPOELLIPSE epicenter location errors for Giles County, Va., blasts*

[ERH, semimajor axis of the error ellipse that results from projection of the error ellipsoid onto a horizontal plane; km, kilometers]

Blast	Date of blast			Difference between actual and calculated epicenter (km)	ERH (km)
	Year	Month	Day		
A	1979	Dec.	3	0.5	2.2
B	1979	Dec.	6	.9	2.4
C	1980	May	20	2.0	5.7

TABLE 6.—*HYPOELLIPSE determination of focal depths for Giles County, Va., blasts*

[ERZ, greatest vertical deviation of the error ellipsoid from the hypocenter; km, kilometers]

Blast	Date of blast			Trial focal depth (km)	Solution focal depth (km)	ERZ (km)
	Year	Month	Day			
A	1979	Dec.	3	4.0	0.5	57.7
				[1]10.0	.2	99.0
B	1979	Dec.	6	.0	.0	99.0
				[1]5.0	2.5	16.7
C	1980	May	20	4.0	2.2	14.3

[1]Two trial focal depths were tested for blasts A and B.

case for the existence of the zone as we have described it even though the data base is not large.

FAULT AREA

Epicenter maps and vertical-section plots of foci are given by figures 11, 13, and 14. Figures 12 and 15 are illustrations designed to portray specific characteristics of the hypocenter data set in the horizontal (fig. 12) and vertical (fig. 15) planes. Figure 12 presents the epicenters, scaled according to magnitude, without any geography (except the location of the station NAV) or error ellipse axes so as to portray the seismicity in as direct a manner as possible. Figure 14 is a side view of the 12 earthquake hypocenters in vertical section, and figure 13 is an end view of those same 12 hypocenters, plus two off-zone hypocenters (nos. 34 and 40). In both figures two of the earthquakes (D and S) do not have vertical error semiaxes on the figure, because there were insufficient arrival-time data available to determine

TABLE 7.—*Chronological listing of earthquakes that occurred prior to 1978 in the Giles County, Va., locale and were relocated using joint hypocenter determination techniques*

[From J. W. Dewey and D. W. Gordon, written commun., 1980. Some of these data appear in table 1]

Event No.	Date			Locality	Origin time (UTC)			Latitude (north)	Longitude (west)	Depth (km)	Magnitude (m_{bLg})	Projection[1]	
	Year	Month	Day		Hours	Minutes	Seconds					Trend (degrees)	Semilengths (km)
D	1959	Apr.	23	Virginia-West Virginia border.	20	58	40.2	37°23.70'	80°40.92'	[2]5	3.8	98.1	12.9, 7.7
H	1968	Mar.	8	Narrows, Va.----	5	38	15.7	37°16.86'	80°46.44'	7.7	4.1	133.5	6.5, 6.1
J	1969	Nov.	20	Elgood, W. Va.--	1	00	9.3	37°26.94'	80°55.92'	2.5	4.6	132.7	6.2, 4.4
R	1974	May	30	Virginia-West Virginia border.	21	28	35.3	37°27.42'	80°32.40'	5.4	[3]3.7	122.7	8.6, 5.1
S	1975	Nov.	11	Giles-Bland Counties, Va., border	8	10	37.6	37°13.02'	80°53.52'	[2]1.0	3.2	144.8	11.6, 6.7
X	1976	July	3	Virginia-West Virginia border.	20	53	45.8	37°19.26'	81°07.62'	[2]1.0	[4]2.1	141.3	13.7, 6.5

[1]Projection onto the Earth's surface of the 90-percent-confidence ellipsoid on the hypocentral coordinates. This projection is specified by the trend of the semimajor axis and the lengths of the semimajor and semiminor axes, respectively.
[2]Focal depth fixed.
[3]Reagor and others (1980a) give a value of 3.6.
[4]Magnitude according to G. A. Bollinger (unpub. data, 1976).

adequately a focal depth even though the data were sufficient for calculation of an epicenter. In such cases, the depth is fixed at some arbitrary, but geologically reasonable depth, by the geophysicist performing the calculations. Figure 15 illustrates the range of fault areas allowed by the 10 hypocenters. That range, from 80 km² to 800 km², was determined by first projecting the hypocenters onto a vertical plane (*A-A'*; see figure 11) and then arbitrarily moving the hypocenters inside their error ellipses in the following manner:

Figure 15*A*—All hypocenters shifted toward the centroid of the hypocentral distribution. Note the superposition of groups of two and three hypocenters. A minimal area (80 km²) is defined by the shallowest eight hypocenters (shaded area). If the deepest two hypocenters are included (shaded plus hachured areas), then the area specified is 250 km².

Figure 15*B*—All hypocenters shifted away from the centroid of the hypocentral distribution, restricted to a minimum focal depth of 5 km, or both. A maximal area of some 800 km² (shaded area) is thereby defined.

Other ways of connecting the dots in figure 15 would produce slightly different inferred fault areas, but those areas could still vary by a factor of 10 times and yet be consistent with the locational accuracy of the hypocenters as specified by the error ellipsoids. Therefore, we do not have, at this time, an accurate estimate of the area of the Giles County, Va., seismogenic zone.

The definition of fault-plane area, 80–800 km², can be used to estimate the magnitude of an associated earthquake: M6–7. Thus, a variation of 10 times in the fault plane area implies a change of one in the magnitude of an associated earthquake (Wyss, 1979, 1980; Singh and others, 1980; Bonilla, 1980). However, the published plots of earthquake magnitude against the logarithm of fault-plane area contain approximately one unit of dispersion in each variable. We and Bollinger (1981a) have used the regression line of magnitude on log (fault-plane area) to estimate magnitude from area and so have not explicitly incorporated this variability. One could argue that that is wrong, because we are estimating the magnitude of the largest shock likely to occur on the seismic zone. However, we have already chosen an extreme value for the area that is consistent with locational uncertainties. The regression line gives the most probable magnitudes expectable from those extreme values of the area. We consider that if we added the uncertainty in the regression to the uncertainty of the area, the resulting magnitude range would be needlessly wide and conservative.

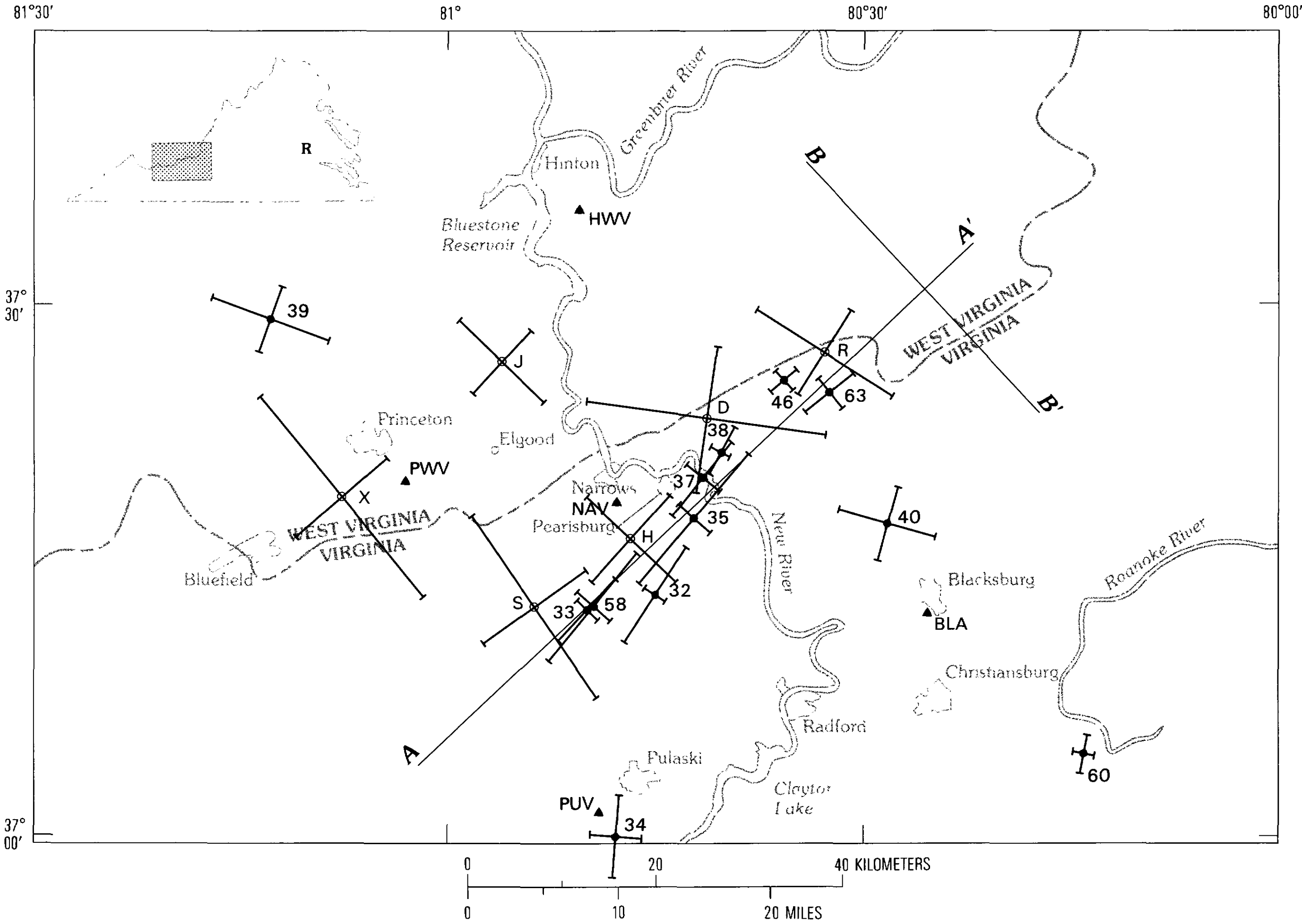


FIGURE 11.—Epicenter map for microearthquakes located by data from the Giles County, Va., subnetwork and for felt earthquakes relocated by J. W. Dewey and D. W. Gordon (written commun., 1980). The epicenters of felt earthquakes are designated by letters and listed in table 7. The epicenters of the microearthquakes are designated by numbers and listed in table 4. The six epicenters of felt earthquakes are shown by open circles and 90-percent confidence ellipsoidal axes, four of which are located on the northeast-striking zone. The locations of the vertical profiles are shown; the northeast-striking *A-A′* line is shown in figure 14, and the northwest-striking *B-B′* line is shown in figure 13. Inset map shows area of this figure and location for Richmond (R). Modified from Bollinger (1981a) with permission.

There are subjective aspects to the specification of seismogenic fault-plane area and estimation of the associated potential magnitude that bear further discussion.

1. Seismic rupture of the ground surface is unknown in or near Giles County. In such cases elsewhere that lack surface evidence, areas of fault planes are usually estimated from spatial distributions of aftershocks. We use here the spatial distribution of seismicity detected during an extended period of time because no aftershock sequences have been detected during our monitoring. However, the existence, orientation, and shape of the seismic zone as defined by the microseismicity are supported by the distribution of felt events. The zone has had nine felt events since 1959 (table 1). Four of the nine (events D, H, R, S in table 7) were relocated within the seismic zone by J. W. Dewey and D. W. Gordon (written commun., 1980; see also figs. 11–14) and events X and J were relocated outside the seismic zone.
2. The confidence ellipsoids used to estimate minimal and maximal fault-plane areas (fig. 15) are of two different types. Locations derived from the Giles County network were calculated using the HYPOELLIPSE program, which produces 68-percent confidence ellipsoids. These locations are shown as solid dots in figures 9–14. The relocations of Dewey and Gordon were calculated using the Joint-Hypocenter-Determination (JHD) techniques,

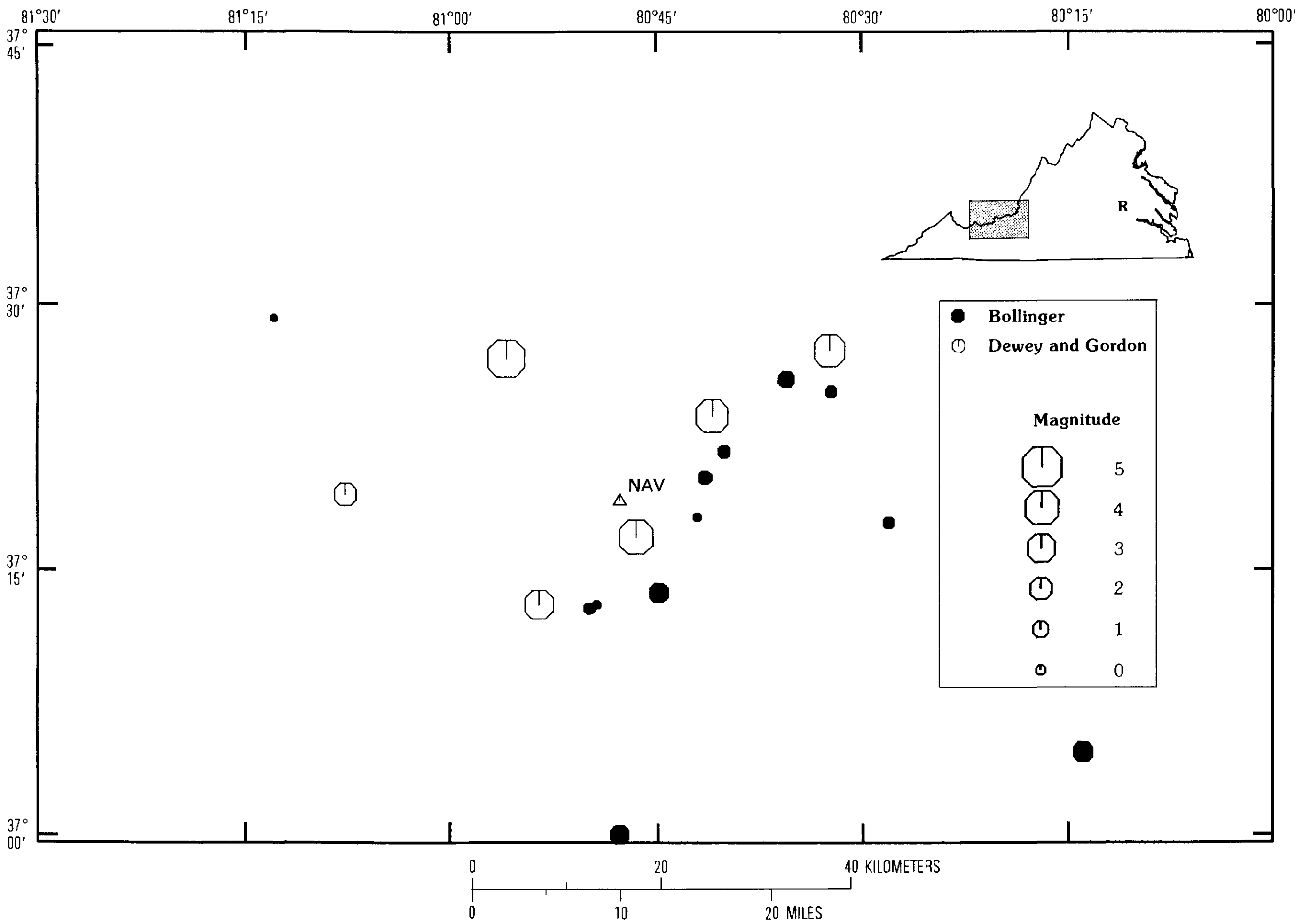


FIGURE 12.—The 18 epicenters of figure 11 scaled according to magnitude. The magnitudes are listed in tables 4 and 7. The epicenters are separated according to locational authority: Open symbols for epicenters according to J. W. Dewey and D. W. Gordon (written commun., 1980; see table 7); solid symbols for epicenters according to this study (see table 4). Inset map shows area of this figure. Modified from Bollinger (1981a) with permission.

which produces 90-percent confidence ellipsoids. These relocations are shown as open circles in figures 11–14. To combine the two properly, the eight 68-percent ellipsoids should be expanded, which would increase the estimated fault plane area, or the four 90-percent ellipsoids should be contracted, which would decrease the area. However, we consider that the resulting changes in the ellipsoid sizes, in the estimated areas, and in the resulting magnitudes would be negligible for our purposes. A recent study that applied the JHD technique to all 12 Giles County events in the seismic zone showed that the hypocenters relocated by the JHD technique have the same general location and trend as do those presented herein (Viret and others, 1981, 1986; Bollinger and others, 1982).

3. The confidence ellipsoids are three-dimensional shapes with various orientations in space. Figure 15 uses only the elliptical projections of the ellipsoids into horizontal and vertical planes. This distorts the estimates of fault plane area. A crude estimate of the amount of distortion may be obtained from a two-dimensional analogy that uses figure 15. The ellipses of figure 15 are drawn using vertical and horizontal semiaxes. Consider how the ellipses would be distorted if they were drawn using semiaxes obtained by projection of the ellipses of figure 15 into two other perpendicular lines lying in the plane of figure 15, say lines plunging 45° to the southwest and to the northeast. Such projected axes would allow fault planes with different orientations but whose areas would not be much different from those shown in figure 15.

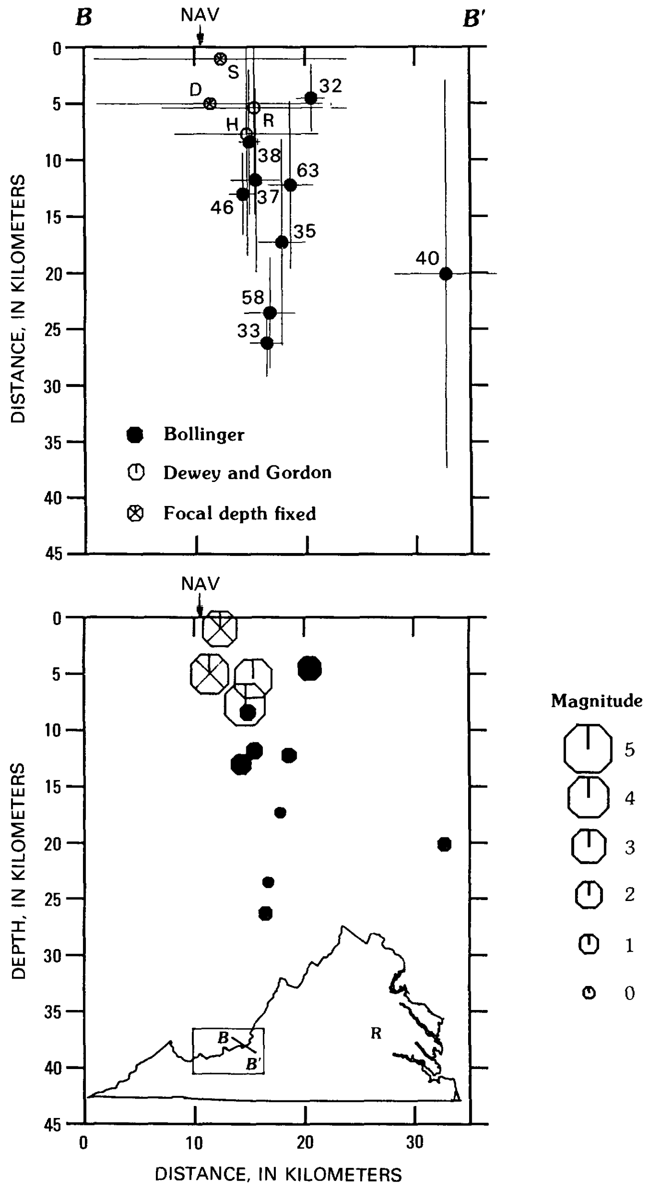


FIGURE 13.—Sections showing vertical distribution of the hypocenters projected perpendicularly into a northwest-striking plane *B-B′* (see fig. 11 for location of *B-B′*). Solid symbols, this report; open symbols, J. W. Dewey and D. W. Gordon (written commun., 1980) with focal-depth control; open symbols with X's, J. W. Dewey and D. W. Gordon (written commun., 1980) without focal-depth control (not enough arrival-time data), indicating depths shown were arbitrarily fixed during calculations. Top part of the figure shows error ellipse axes and event numbers and letters; in bottom part of the figure, hypocenter symbols are scaled according to magnitude (<1 to >4) of the individual earthquakes. Event numbers and letters refer to tables 4 and 7, respectively. Confidence ellipsoidal axes shown are at a 68-percent level for numbered events (from Giles County network) and a 90-percent level for lettered events (from J. W. Dewey and D. W. Gordon, written commun., 1980). Location of seismic station NAV shown by arrow on both parts of figure. The inset map shows the profile location and Richmond (R). Modified from Bollinger (1981a) with permission.

Thus, the effect of such a projection on the minimal and maximal fault plane areas would be negligible. Analogously, after consideration of the elliptical shapes as indicated by the semiaxes of figures 11 and 13–15 and after consideration of the ellipsoidal semiaxes of table 4, we conclude that this effect is also negligible for our purposes in three dimensions.

Recently, Bollinger (1981a) has presented an assessment of the potential hazard for use by public officials and emergency planners, but this assessment is not detailed enough to be of use in defining engineering specifications of structures. Such an assessment involves two major factors: (1) specification of a fault-plane area for use in estimating the potential earthquake magnitude; and (2) development of a hypothetical intensity map. The initial factor (1) has been discussed. The second factor (2), development of a hypothetical intensity map, attempts to utilize the geometric characteristics of local and regional isoseismal maps along with magnitude-intensity relationships and intensity-distance attenuation functions (Bollinger, 1981a). Application of these various characteristics, relationships, and functions to the Giles County data could, in principle, yield a range of possible results depending on initial assumptions and objectives. The specific results developed by Bollinger (1981a) for the study area were as follows: Potential earthquake size: M_S=7, I_o=IX (MMI); Hypothetical intensity map—all isoseismals elliptical in shape with principal zones of damage having areas of 785 km^2 (IX), 4,500 km^2 (VIII) and 31,700 km^2 (VII). The innermost isoseismals (VIII, IX) are postulated to have long dimensions that trend with the seismogenic zone (N. 44° E.), but the lower level isoseismals (VII and below) are to trend with the tectonic fabric of the surrounding portion of the Appalachians (N. 75° E.)(Bollinger, 1981a, p. 285).

Bollinger's (1981a) estimate of the size of the largest shock possible on the Giles County seismic zone is consistent with two suggestions of Nuttli (1981a, b). Nuttli compared Eastern and Western United States seismicity and suggested that (1) large Eastern shocks can arise from structures of only moderate size, and (2) most Eastern regions have probably not experienced their largest possible shock yet in historic times.

FOCAL MECHANISM STUDIES

A composite focal-mechanism solution (CFMS; fig. 16) was attempted for those 8 microearthquakes that have the most accurate locations and form the tightest spatial distribution in map view. According to our interpretation, they occurred on the same fault or fault

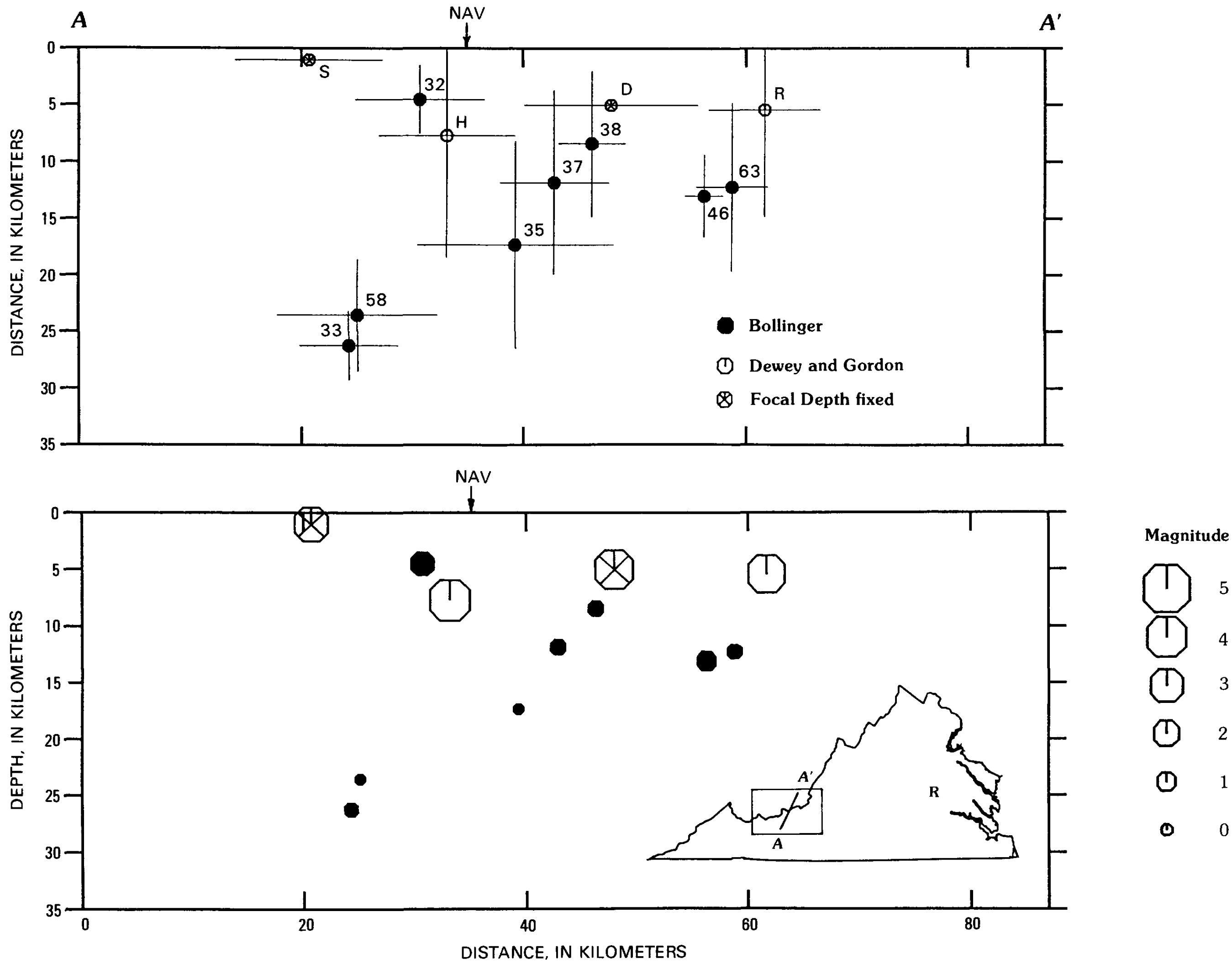


FIGURE 14.—Section showing vertical distribution of the hypocenters that define the seismic zone, along a northeast-striking plane *A*-*A*′ (see fig. 11 for location of *A*-*A*′). Solid symbols, this paper; open symbols, J. W. Dewey and D. W. Gordon (written commun., 1980) with focal-depth control; open symbols with X's, J. W. Dewey and D. W. Gordon (written commun., 1980) without focal-depth control; therefore, depths shown were arbitrarily fixed during calculations. Upper half of the figure shows error ellipse axes and event numbers and letters; in lower half of the figure, hypocenter symbols are scaled according to magnitude (<1 to >4) of the individual earthquakes. Event numbers and letters refer to tables 4 and 7, respectively. Confidence ellipsoidal axes shown are at a 68-percent level for numbered events (from Giles County network) and a 90-percent level for lettered events (from J. W. Dewey and D. W. Gordon, written commun., 1980). Inset figure shows the profile location and Richmond (R). Modified from Bollinger (1981a) with permission.

zone. Because of the small size (low energy level) of the individual shocks, only 14 *P*-wave polarities could be obtained (six impulsive, eight emergent; see table 8). A unique focal mechanism could not be obtained from the data set (we easily obtained three different solutions). Figure 16 gives a provisional composite focal-mechanism solution for these 8 well-located microearthquakes. To construct figure 16, we take the strike of that zone to be N. 44° E. and the dip to be 80° NW. The strike has been discussed previously, and the dip is our subjective visual fit to the foci in figure 13.

To glean as much as possible from the *P*-wave data set, the following procedures were employed to develop a focal-mechanism solution:

1. We used the microearthquake hypocenter distribution (figs. 9, 13) to define subjectively the preferred nodal plane as striking N. 44° E. and dipping 80° northwest, as follows. In Appendix D, we have noted that statistical analysis yielded a dip of 40° for the tabular seismic zone. However, we concluded there that the dip estimate was not clearly reliable. Further, if we had used a 40° dip for the

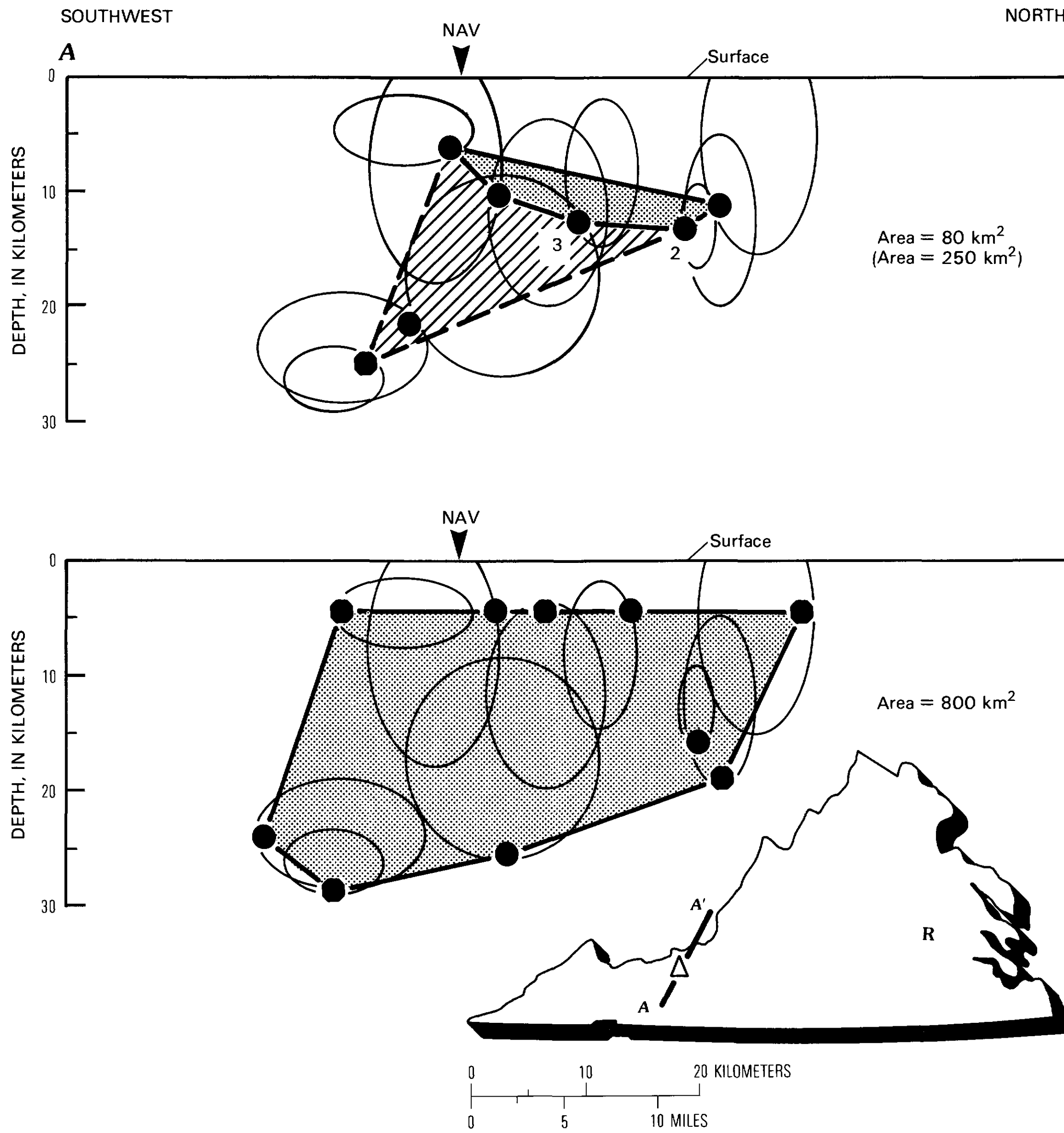


FIGURE 15.—Sections showing definition of possible fault-plane areas. Examples of hypocenter (solid dots; each with an associated error ellipse) distributions and interpretations of fault-plane areas that can be derived by first projecting the hypocenters onto a vertical plane (A–A'; see fig. 11) and then arbitrarily moving the hypocenters of figure 14 to various positions inside their error ellipses: Top—Minimal area (80 km^2, shaded region) and an intermediate size area (250 km^2, shaded plus hachured regions). Numerals indicate the number of hypocenters moved to the same point. Bottom—Maximal area (800 km^2, shaded region). Inset map shows locations of profile, the NAV station (open triangle) and Richmond (R). Note that events S and D have unknown focal depths (see figs. 13 and 14), so they were not used here; thus, only 10 points are plotted. Modified from Bollinger (1981a) with permission.

preferred nodal plane, the resulting CFMS would have predicted normal movement on that nodal plane. Such normal movement would be inconsistent with a maximum horizontal compressive stress trending east-northeast which is the orientation we shall infer in a later section on "State of Stress." Accordingly, we used a dip for the preferred nodal plane of 80° NW., estimated from a visual fit of a line to the foci of figure 13. We then found, by graphical means, an auxiliary plane that encompasses the compressional field (9 of 11 compressional P-wave polarity readings: table 8) observed in the northeast to south azimuths. Figure 16 shows that auxiliary plane to strike north-south

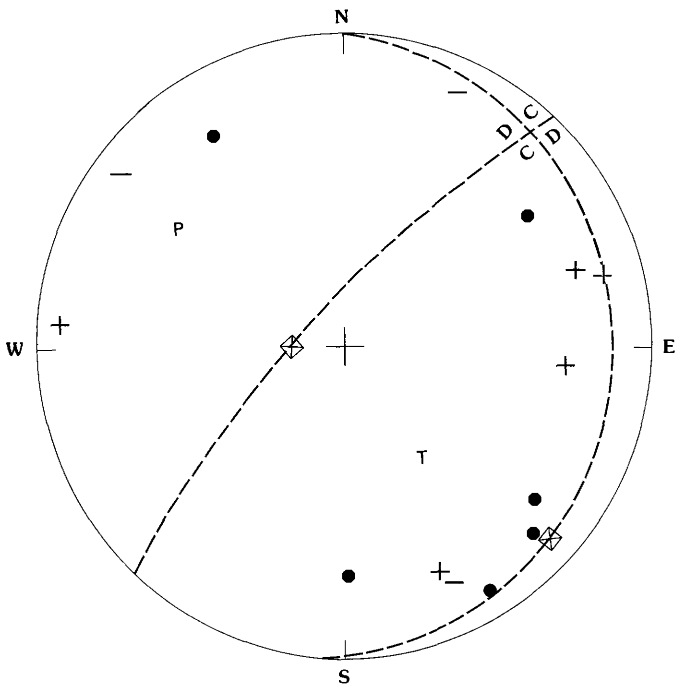


FIGURE 16.—Provisional composite focal-mechanism solution for events in the Giles County, Va., seismic zone. Lower-hemisphere equal-area projection. Symbols: Large plus, center of projection; solid circles, definite (impulsive) compressions; small plus signs, indefinite (emergent) compressions; minus signs; indefinite (emergent) dilatations; P and T, pressure and tension axes at the source, respectively; boxes with X's, nodal plane poles; dashed lines, nodal planes; C and D, quadrants about the source where the *P*-wave arrivals show compressional (away from source) and dilatational (toward source) first motions, respectively.

TABLE 8.—*P-wave polarity data for Giles County, Va., earthquakes*

[Event No., listing as shown in table 4; AZM, epicenter-to-station azimuth (in degrees); AIN, angle-of-incidence (measured from the downgoing vertical) at the focus (in degrees). P-wave polarity: C, compression; D, dilatation; e, emergent beginning of P wave; i, impulsive beginning of P wave. Station abbreviations are defined in table 2]

Event No.	Date			Station	AZM	AIN	P-wave polarity
	Year	Month	Day				
32	1978	Jan.	28	BLA	274	81	eC
				NAV	157	67	eC
33	1978	May	10	HWV	179	63	iC
37	1978	July	28	HWV	155	73	eD
38	1978	Aug.	30	BLA	307	80	eD
				HWV	149	81	iC
40	1978	Oct.	14	NAV	95	61	eC
				HWV	135	75	iC
46	1980	Feb.	18	BLA	328	69	iC
				NAV	55	62	iC
				HWV	129	68	iC
				PWV	75	76	eC
				PUV	24	79	eD
				CVL	72	68	eC

and to dip 14° to the east. The resulting CFMS (fig. 16) indicates right-reverse motion on the preferred nodal plane. The CFMS suggests that the reverse component is larger than the right-slip component of motion. However, we do not actually know which component is larger. That is because the relative magnitudes of the two components depend on the orientation of the auxiliary nodal plane. That orientation is uncertain, partly because the first motions are too few and too poorly distributed on the focal hemisphere to restrict the orientation of the auxiliary nodal plane (fig. 16), and partly because the hypocenters do not tightly constrain the dip of the preferred nodal plane (fig. 13; Appendix D).

2. Two of the four Dewey and Gordon (written commun., 1980) relocations in the seismic zone (table 7; fig. 11) have fixed focal depths (events D and S) and the other two (H, R) have rather large horizontal and vertical error estimates. Thus, it would be somewhat questionable to combine data from those shocks with data from the more precisely located microearthquakes. However, we note that the WWSSN Observatory BLA is always in the northwestern (dilatational) quadrant of the focal sphere (because we are considering a lower focal hemisphere; table 8). A check of the BLA seismograms for the D, H, R, and S events revealed only one clear reading (an impulsive dilatation from the event S) and one somewhat indefinite reading (a compression for event H). Thus, the single check we are able to make tends to agree with the CFMS, but not without some ambiguity. Resolution will come only with more data from larger earthquakes (M>3).
3. We can evaluate the solution with binomial tests. Details are in Appendix E. Statistical results discussed there provide objective support for our subjective opinion that the CFMS of figure 16 is valid, despite being based on a small number of first motions with several inconsistencies.
4. We compared the Giles County, Va., CFMS with other, nearby focal-mechanism solutions. With no previous focal-mechanism solutions for events in the seismic zone, a direct comparison of this type is not possible. There are, however, two nearby focal-mechanism solutions that will provide a measure of comparative value. Those solutions are for

the 1969 Elgood, W. Va., shock (event J in fig. 11 and table 7) and the 1973 Knoxville, Tenn., earthquake (epicenter: 35.8° N. 84.0° W.; origin time (UTC): 0748:41.2; m_b=4.6).

Herrmann (1979) used *P*-wave first motions and surface-wave amplitude and phase data from Love and Rayleigh waves to obtain a focal-mechanism solution for the 1969 Elgood, W. Va., shock. That solution showed predominately strike-slip motion. The nodal plane strikes were northeast and northwest, and the dips were near vertical. The northeast-striking plane (N. 33° E., 80° SE.) exhibited a left-lateral motion, with a small normal component. Thus, the strike and dip (but not the sense of movement) of one of the solution's nodal planes are similar to those for the Giles County, Va., zone. Note that the 1969 Elgood, W. Va., shock was not directly in the Giles County zone, but rather some 25 km to the northwest of that zone (fig. 11).

Focal mechanism solutions for the 1973 earthquake were obtained by Bollinger and others (1976) and by Herrmann (1979). The former investigators found a dip-slip mechanism, but could not, because of meager polarity data, differentiate between normal and reverse modes of faulting. That is, they obtained two equally likely solutions, one showing normal faulting (NE.- and NW.-striking nodal planes) and the other defining reverse faulting (both nodal planes had NW. strikes). The northeast-striking nodal plane (N. 49° E.; dip 70° SE.) has an orientation roughly similar to the strike and nearly vertical dip of the Giles County, Va., zone. Bollinger and his coauthors (1976) favored the reverse-faulting solution based on other data (trend of aftershock epicenters, the vertical distribution of the aftershock hypocenters, and the trend of regional in situ stress measurements). Interestingly, Herrmann obtained a predominately strike-slip mechanism for this shock (nodal planes with NNE. and WNW. strikes and steep dips). He rated the solution quality as "C" (average) and noted that, because of the skimpy data base, he had little faith in either his solution or that by Bollinger and others (1976). The 1973 Knoxville earthquake was located some 320 km along strike and to the southwest from Pearisburg, Va.

Thus, from other focal-mechanism studies we find some supporting evidence for seismically active, northeast-striking, steeply dipping seismic zones in the general area and in the same geologic province as the Giles County, Va., zone. The evidence favors right-reverse motion but is far too mixed and uncertain to be definitive at this stage.

TYPES OF FAULTS POTENTIALLY RESPONSIBLE FOR THE GILES COUNTY SEISMIC ZONE

INTRODUCTION

We shall now determine the type of fault that is most likely to be responsible for the seismicity of Giles County, including the earthquake of 1897, and estimate the portion of the East in which similar faults might be expected to occur. The discussion to follow is long and involved because pertinent data are sparse. However, the effort is worthwhile: We conclude that (1) the seismicity of Giles County probably occurs by compressional reactivation of a fault that formed when an ocean called Iapetus opened, in Late Proterozoic or early Paleozoic time, and (2) similar faults may occur under most of the western portion of the southern and central Appalachians and adjacent craton.

What is known of the geologic evolution of southeastern North America indicates that the crust beneath the region that includes Giles County has undergone four deformational episodes. Each episode is known or can reasonably be inferred to have produced faults that may have been reactivated under present-day stresses to produce the Giles County seismic zone. Each episode was caused by movements of the North American plate and other plates and is known or inferred to have produced faults with specific and predictable properties throughout all or portions of the region now occupied by the Appalachians and the Coastal Plain. Those fault properties can be compared to properties of the crust under Giles County and surrounding areas, to characteristics of the seismicity of Giles County, and to geological data. Of the four kinds of faults produced by the four deformational episodes, the one whose properties best match those of the Giles County locale is the one most likely to include the structure that produces the seismicity of Giles County.

The four deformational episodes occurred (1) about a billion years ago during the Grenville orogeny, (2) during crustal extension in the Late Proterozoic or early Paleozoic, as the Iapetus Ocean began to open, (3) during crustal loading later in the Paleozoic as Appalachian thrusting reached the Giles County locale, and (4) during renewed crustal extension in the early Mesozoic as the Atlantic Ocean began to open (Wheeler and Bollinger, 1980).

CRUSTAL PROPERTIES AND SEISMICITY

We have documented that the current seismicity in the Giles County locale is concentrated in a nearly vertical, tabular zone that strikes N. 44° E. and extends from 5 to 25 km in depth; we have argued that that zone is probably the source of the 1897 shock; and we have concluded that structures responsible for the seismic zone lie in the basement, beneath exposed and near-surface rocks, folds, and thrust faults.

The Giles County seismic zone involves the upper half of the continental crust beneath the thrust faults. The best estimate of local crustal thickness is 51 km, which is based on traveltime analyses of local and regional earthquakes and quarry blasts and on an unreversed refraction survey (Moore, 1979; model TPM2 of Appendix C, of this report). That thickness estimate is consistent with a previous one of 50–55 km derived from regional analysis of seismic traveltime terms (James and others, 1968). Sbar and Sykes (1977), Dewey and Gordon (1980), and Acharya (1980b; but see Stevens, 1981) suggested that small earthquakes occurring deeper than about 10 km indicated a potential for large earthquakes. That is consistent with the suggestion by Bollinger (1981a) that the Giles County seismic zone could generate a large shock. Also, the depth distribution of hypocenters of the seismic zone is consistent with a suggestion by Chen and Molnar (1981) that continental regions are characterized by an aseismic lower crust. The lower crust could be aseismic because it is too ductile to support high stresses (Meissner and Strehlau, 1982), perhaps because the grains of common minerals that support stress recrystallize in the lower crust (Toriumi, 1982).

Because of the size of the seismic zone, any structure or structures responsible for the zone must be of crustal scale. It seems reasonable to expect that any such large, nearly vertical, presumably planar structures are faults or fault zones that had their origins in processes operating on regional, continental, or plate scales. Only such processes could stress the entire upper crust and cause it to fail. Eventually, data with which to identify clearly such deep, seismogenic faults in the Giles County locale may result from interpretation of deep seismic-reflection lines, from detailed modeling and interpretation of new and existing gravity and aeromagnetic data, from new geologic mapping and analyses, or from analysis of future seismicity beneath the Giles County network. In the meantime, consideration of the geologic history of the Giles County locale and its surroundings can help to define the probable type, age, and motion of such seismogenic faults, as well as the geographic area within which there may occur analogous faults with similar potential for seismic hazard.

Here, we should note an assumption that underlies most of our geological interpretation of the seismological data. We assume that if the Giles County seismic zone does occur on a fault or fault zone, then that fault or fault zone is an old one that is being reactivated in the present stress field. It is not a fresh crustal break formed in unfractured rock in direct response to today's stress field. There are two reasons for making this assumption.

First, where continental basement is exposed, it is commonly cut by old faults and shear zones of various ages, sizes, orientations, and movement histories. For instance, Odom and Hatcher (1980) described examples from the Appalachians, and Isachsen and McKendree (1977) mapped similar features in the Adirondacks. Many geologists have long argued that, in intraplate regions, reactivation of older faults may be the rule and formation of new faults, the exception. Recently, Hamilton (1981) has summarized evidence that suggests that large Eastern earthquakes occur on reactivated rather than new faults.

Second, regardless of the stress state at the fault, a weak zone that is at or near the optimum failure orientation will yield before fresh rock will. The following sections demonstrate that ancient, crustal-scale faults probably formed in the region that is now occupied by Giles County, with the orientation and size that we observe for the seismic zone. Some of those ancient faults formed in Late Proterozoic or early Paleozoic time, as an ocean called Iapetus opened. Since then, no events are known to have affected Giles County that are likely to have significantly deformed, annealed, or otherwise strengthened most such faults. Thus, it is probable that some of the ancient faults are still weak and would be reactivated in preference to forming new faults.

GRENVILLE OROGENY

Roughly in middle Middle Proterozoic time (about a billion years ago) the Grenville orogeny occurred. The metamorphic and igneous basement rock under and near Giles County lies in that part of eastern North America that was deformed or recrystallized, or both, during the Grenville orogeny (Ammerman and Keller, 1979, p. 344; Bass, 1960; Bayley and Muhlberger, 1968; Black and Force, 1982; Lidiak and Zietz, 1976; Lidiak and others, 1981). Glover and others (1978) have identified rocks of Grenville age in the thrust sheets of the Piedmont province in central Virginia. Pertinent data are sparse, but any high-angle faults that formed during or before the local Grenville deformational or metamorphic peak(s) should have been sufficiently deformed or annealed, or both, by Grenville events that they no longer constitute important strength discontinuities. R. C. Shumaker (1982; written commun., oral communs., 1978–81) is analyzing published and unpublished structural, stratigraphic, geophysical, and oil- and gas-production data from central and southwestern West Virginia and eastern Kentucky. He has suggested that basement faults of Grenville age have been reactivated in that region throughout Paleozoic time. However, all the areas in which such reactivation is known to have occurred lie west of the New York–Alabama magnetic lineament (King and Zietz, 1978; Zietz and others, 1980), which crosses central West Virginia about 100 km northwest of Giles County (fig. 17). The magnetic lineament lies approximately along the ill-defined southeastern edge of a large Paleozoic graben called the Rome trough (this report, fig. 25; Harris, 1975, 1978; Shumaker, 1977). Reactivated basement faults in central and southwestern West Virginia and eastern Kentucky are probably parts of the Rome trough and probably are not analogs of the Giles County seismic zone.

Odom and Hatcher (1980) discussed the potential for reactivation of faults formed before, during, and after the occurrence of Paleozoic metamorphic peaks of the Appalachian orogenies. Those Paleozoic metamorphic peaks occurred tens of kilometers southeast of Giles County, and so did not affect the rocks under consideration here.

IAPETAN NORMAL FAULTS

Late Proterozoic or early Paleozoic normal faults could be the sources of Giles County seismicity. Such faults formed in North American cratonic crust as an ancient ocean opened, early in the development of a passive (Atlantic-type) continental margin. Features in the Bouguer gravity field over the Appalachians are used here to suggest the extent and limits of the area beneath which such faults may be expected to occur. We will argue that the southeastern limit of such faults is probably a large eastward rise in the Bouguer anomaly field that runs the length of the Appalachians. We will also suggest that the likelihood of encountering such faults decreases gradually to the northwest of the gravity rise, over a distance of several tens to several hundreds of kilometers.

IAPETUS OCEAN

The predecessor ocean of the Atlantic began opening in Late Proterozoic time, and closed progressively throughout the middle and late Paleozoic to produce the various Appalachian and Atlantic Caledonide orogens from Alabama, U.S.A., to Spitsbergen, Norway. The ocean was named the proto-Atlantic by Wilson (1966). However, the same term applies to the early stage of the Atlantic Ocean. Accordingly, Harland and Gayer (1972, p. 305) took the less confusing name Iapetus (from Greek mythology) for the northern part of the Paleozoic ocean, which separated the Eurasian and North American cratons. (See also reviews by McKerrow and Ziegler, 1972a; Cocks and others, 1980.) South of New England, the Paleozoic ocean opened and closed later than did Iapetus proper (Harland and Gayer, 1972), because of the involvement of a plate carrying the African and South American cratons rather than the Baltic craton (McKerrow and Ziegler, 1972b). The evolution of the southern Paleozoic ocean was further complicated by microplates caught between the converging cratons. Regardless, Williams (1978) and Williams and Max (1980) applied the name Iapetus to the area from Spitsbergen south to the southernmost Appalachians, and we follow that simplifying usage here.

GRAVITY MAPS AND THE IAPETAN CONTINENTAL EDGE

A steep gravity gradient runs the length of the Appalachians, with Bouguer gravity values rising eastward across the gradient as much as 80 mGal (Woollard and Joesting, 1964; Earth Physics Branch, 1974; Haworth and others, 1980). Figure 18 shows the part of the gradient in the region of interest here, near Giles County. The base of the gradient is shown by the −60 mGal contour in central Virginia, and the −80 mGal contour in western North Carolina. The top of the gradient is shown by the O mGal contour throughout the area of figure 18. The position of the gradient is clear from central Alabama to southern Vermont, but farther north the shape of the Bouguer field becomes more complex (Woollard, 1948; Griscom, 1963; Woollard and Joesting, 1964; Diment, 1968; Diment and others, 1972; Earth Physics Branch, 1974; Haworth, 1975; Haworth and others, 1980). Because of the complexity of the Bouguer field in New England and because New England lies beyond the geographic scope of this report, we shall restrict the following discussion to the central and southern Appalachians, and mostly to the area of figure 18. However, where pertinent, we shall cite papers and observations from elsewhere in the Appalachians.

R. W. Simpson, M. F. Kane, and coworkers have produced a set of gravity maps that show considerably more detail and complexity in the Bouguer field than is visible on most of the maps just cited (Simpson, Bothner, and Godson, 1981; Simpson and Godson, 1981; Simpson, Godson, and Bothner, 1981; Kane and Simpson, 1981; Kane and others, 1981; Kane, 1982). Their maps are derived from digitized Bouguer gravity values, contain computer corrections for terrain more than 0.895 km from the stations, are computer contoured and plotted in color, and show the Bouguer

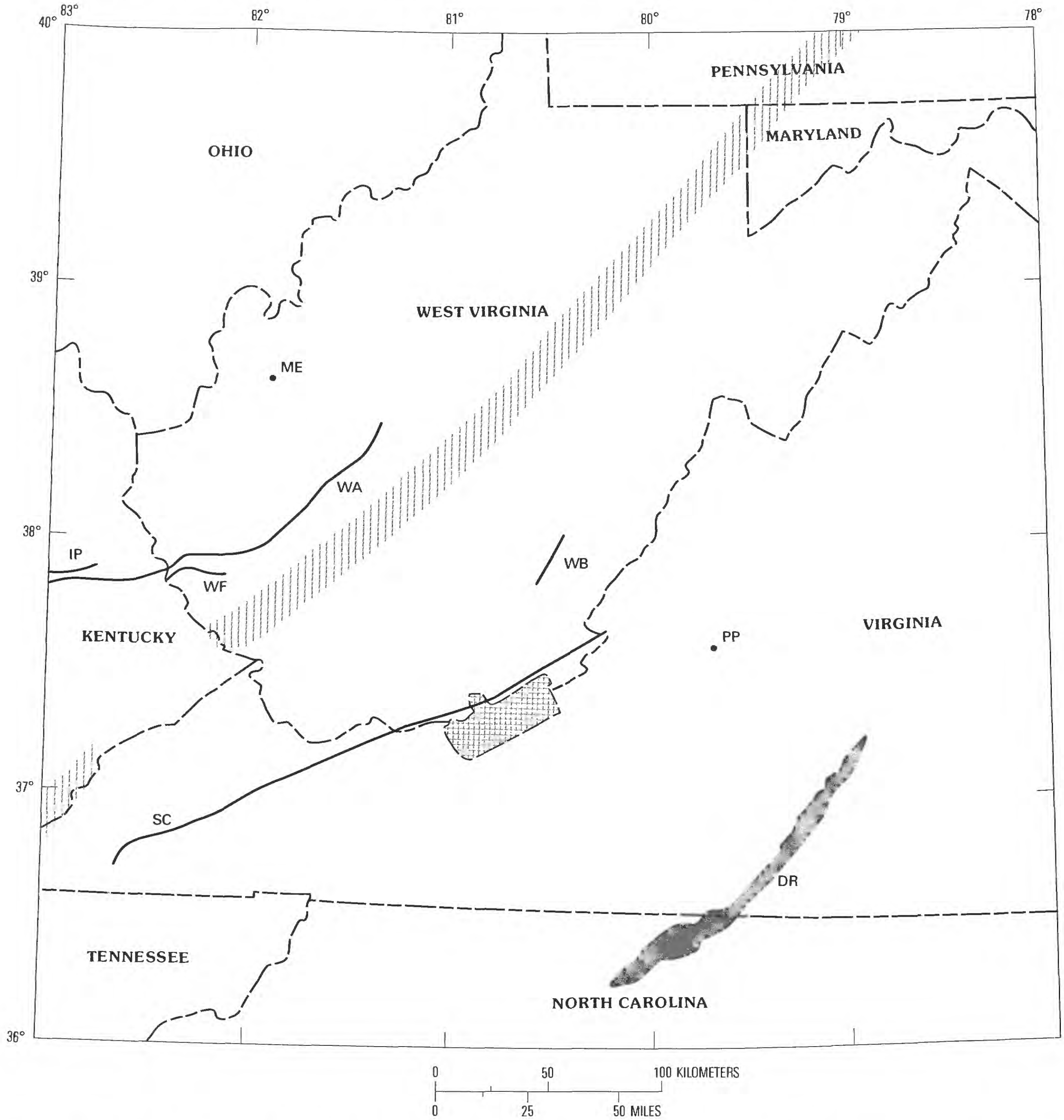
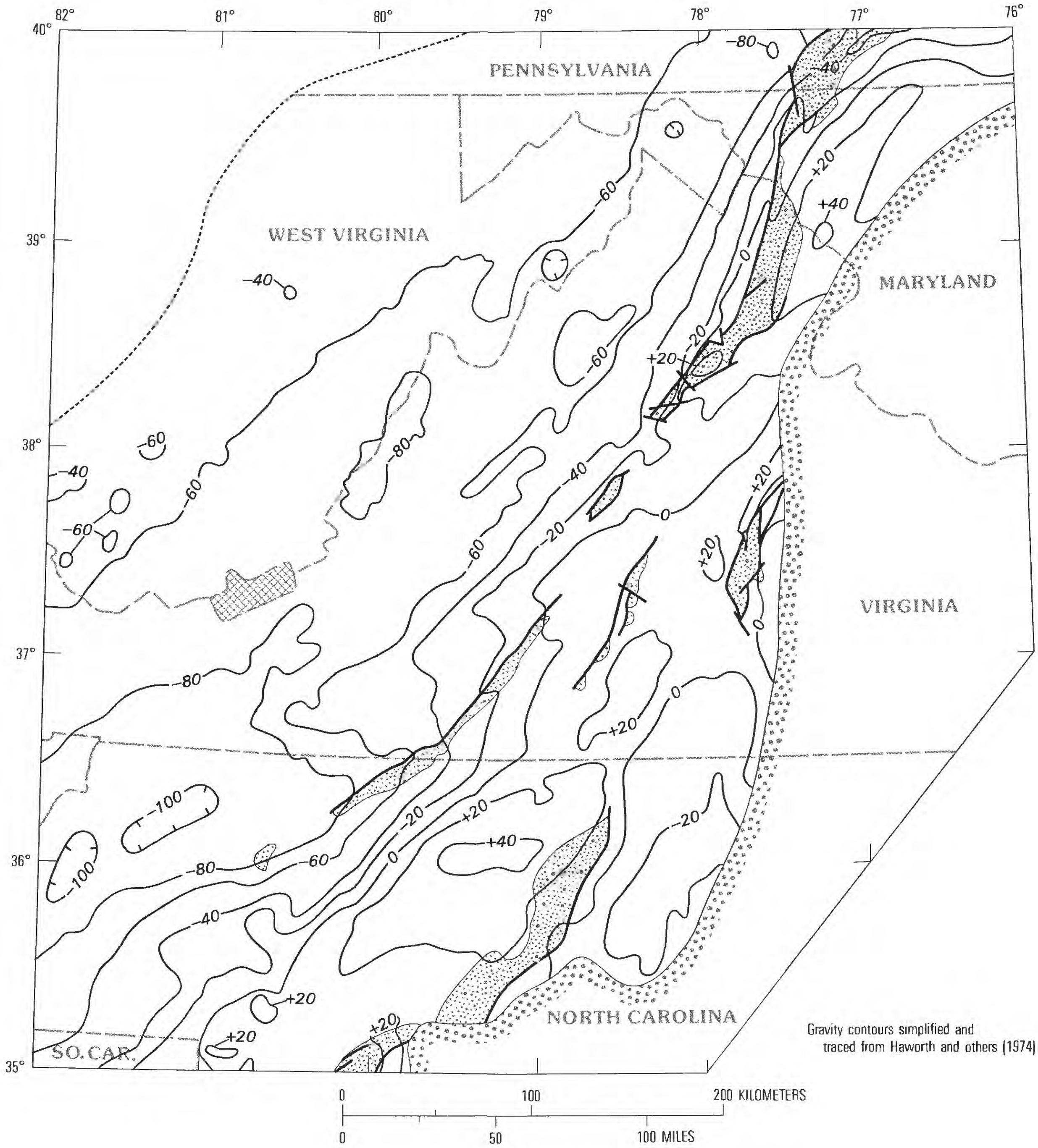


FIGURE 17.—Index map showing locations of some structures and other features named in the text. Hachured band approximates location of New York-Alabama magnetic lineament. Cross-hatched area is Giles County. DR, Dan River basin; IP, Irvine-Paint Creek fault; ME, Midway-Extra gas field; PP, interference structure formed by Pulaski thrust sheet and Purgatory Mountain anticline; SC, St. Clair fault; WA, Warfield anticline; WB, Williamsburg anticline; WF, Warfield fault. Modified from Rodgers (1970, pl. 1A), Cardwell (1976), Calver and others (1963), King and Zietz (1978).

anomaly field and several derivative fields calculated from the Bouguer values. The colored maps show that the gradient, part of which is shown in figure 18, is a geometrically complex eastward rise in Bouguer values. The rise and the portions of the Bouguer field on either side of the rise consist mostly of numerous irregularly



EXPLANATION

—-60— **Equal gravity contours**—Values are in milligals (1 gal = 1 cm/s^2). Contour interval is 20 mGal. Hachures point into closed lows.

Areas of Mesozoic sedimentary rocks—Fillings of extensional basins

Coastal plain onlap—Simplified edge of onlap from the Southeast of Mesozoic and Cenozoic rocks and sediments onto Paleozoic and older rocks of the Appalachians, and onto Mesozoic rocks filling extensional basins

Normal faults

Giles County

State boundaries

Edge of map—Northwest edge of gravity data mapped by Haworth and others (1974)

FIGURE 18.—Eastward gravity rise in exposed parts of central and southern Appalachians near Giles County locale. Geology from Williams (1978). Gravity from Haworth and others (1980). For more details of the geology and gravity field over larger areas, see maps of King and Beikman (1974), Woollard and Joesting (1964), Williams (1978), Haworth and others (1980), Simpson, Bothner, and Godson (1981), and Simpson and Godson (1981).

linear, anastamosing highs and lows. Anomalies of many widths are superimposed. The individual anomalies are separated across strike by second-order gradients of several milligals to several tens of milligals and replace each other along strike. Accordingly, rather than refer to a single gradient like that shown in figure 18, we shall refer to the eastward rise in the Bouguer anomaly field: The rise has as much internal structure as the parts of the Bouguer field that it separates, but the part of the gravitational field east of the rise generally has more positive Bouguer values than does the part west of the rise (R. W. Simpson, oral and written communs., 1981).

Simpson and coworkers used Fourier transform techniques to digitally separate the Bouguer anomaly field into a regional part, comprising all anomalies with wavelengths exceeding 100 km, and a residual part, made up of all anomalies with wavelengths less than 100 km. They performed a similar separation at a wavelength of 250 km (Simpson, Bothner, and Godson, 1981, their figs. 4, 5; Simpson and Godson, 1981, their figs. 4, 5). This process is called wavelength filtering. The two maps of residual (short-wavelength) fields, and especially the two maps of regional (long-wavelength) fields, all reflect the same presence and position of the rise as seen in the unfiltered field, from Vermont to Alabama. Figure 19 summarizes this spatial stability of the gravity rise for the region near Giles County.

A common interpretation of the prominent eastward rise in the Bouguer anomaly field is that it marks the southeastern edge of relatively intact North American continental crust. The edge is a relic of the early opening of the Iapetus Ocean (in addition to many of the papers already cited, see, for example, Fleming and Sumner, 1975; Rankin, 1975, p. 327–328; Long, 1979; Hatcher and Zietz, 1980; Price and Hatcher, 1980; Iverson, 1981; Kumarapeli and others, 1981; Cook and Oliver, 1981; Iverson and Smithson, 1982; Odom and Fullagar, 1982; Schwab, 1982; Thomas, 1982a). W. H. Diment (oral commun., 1981) noted that the rise could have different causes along different parts of its length. Interpretations of the rise by various authors previously cited include (1) eastward crustal thinning, caused by a change from continental crust to buried oceanic crust, or caused by an uplift of the upper mantle and lower crust on steep faults, and (2) eastward change to denser crust (oceanic, denser continental, or transitional) of the same or lesser thickness.

For example, several workers have computed geological models whose density distributions are consistent with the shape and amplitude of the rise. Diment (1968) suggested that the rise in Vermont could be caused by uplift east of the rise of dense lower crustal rocks along a steep fault. For northeast Georgia, Long (1979) suggested that the rise marks the west edge of a terrane of continental fragments separated from each other and the craton by remnants of a Paleozoic rift or rifts. For the same area, Cook and Oliver (1981) showed that a model based on density distributions typical of the modern Atlantic continental margin is consistent with the shape, position, and amplitude of the rise. Further, Kean and Long (1981) estimated from seismic-refraction arrival-time data that crustal thickness decreases about 13 km southeastward across the gravity rise in parts of Tennessee, the Carolinas, and Georgia. They showed a decrease of crustal thickness from a mean of 49 km northwest of the rise, to a mean of 36 km southeast of the rise, with a value of 33 km for the region immediately southeast of the rise. Their thickness estimate of 50 km at Blacksburg, Va., northwest of the rise, is in excellent agreement with the 51 km determined for the Giles County locale, about 25 km west of Blacksburg (Moore, 1979; see model TPM2 of Appendix C of this report). Similar eastward decreases in crustal thickness across the region of the rise were suggested by James and others (1968; decrease from 45 to 50 km to 35 to 40 km, as derived from seismic traveltime terms and corroborated by Chapman (1979)) and by Carts and Bollinger (1981; averaged thicknesses decrease from 40 to 33 km, as derived from an updated crustal velocity model based on recent earthquake arrival-time data).

Regardless of local causes of the eastward gravity rise, it is important for our purposes to note that we interpret the two maps of long-wavelength Bouguer gravity anomalies of Simpson, Bothner, and Godson (1981), and of Simpson and Godson (1981) to indicate that the North American craton extends at least as far east as the rise in the unfiltered Bouguer anomaly field, which is east of Giles County (fig. 19). This is presumed to be true for all crustal levels below the Appalachian thrust sheets, including those at the depths of Giles County seismicity (5–25 km). That interpretation is made because the process of wavelength filtering can be thought of in terms of the depths of the rock masses (sources) that produce gravity anomalies of various wavelengths (fig. 20). For sources of the same sizes and density contrasts, deeper sources produce broader (longer wavelength) anomalies. That correspondence between source depth and anomaly wavelength is not perfect because broad, shallow sources can also produce long-wavelength anomalies. However, in general, the process of wavelength filtering, which separates the total field into a short-wavelength (residual) part and a long-wavelength (regional) part, can be thought of as separating gravity anomalies that are caused by sources within different depth ranges. That is, the residual field from the 120-km wavelength filter is regarded as composed mostly of anomalies caused by sources at upper

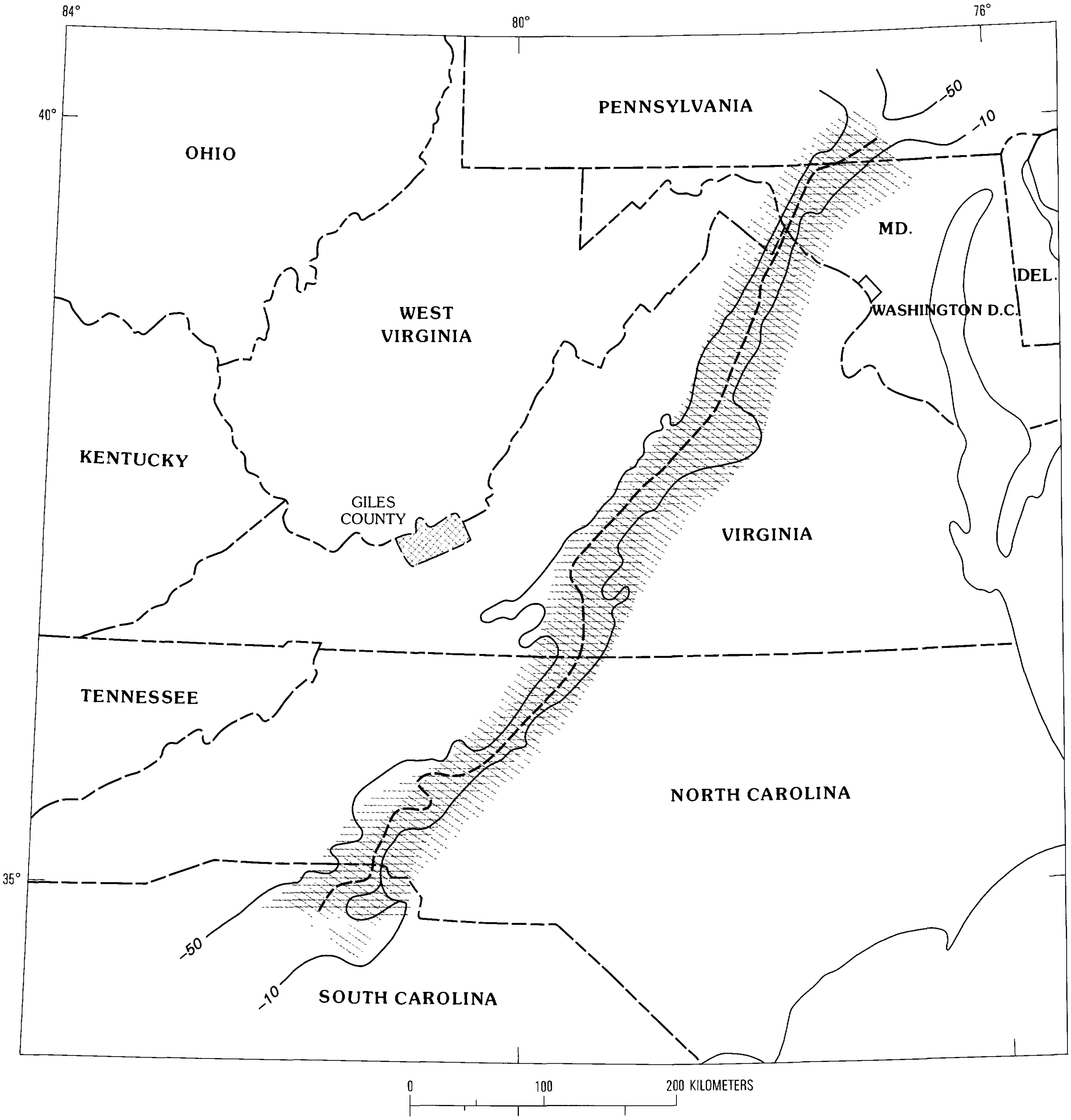


FIGURE 19.—Positions of eastward gravity rise in wavelength-filtered Bouguer anomaly fields. Isogal values of -50 mGal and -10 mGal define the bottom (northwest side) and top (southeast side) of the rise, respectively, in the unfiltered Bouguer field. Halfway in value, the -30 mGal isogal is shown by a heavy dashed line. Horizontal ruling shows position of gravity rise as it appears on map of anomalies with wavelengths longer than 125 km (ruling covers the area between -40 mGal and +10 mGal isogals). Diagonal ruling shows position of rise as it appears on map of anomalies with wavelengths longer than 250 km (ruling covers the area between -30 mGal and 0 mGal isogals).

Gravity data simplified and traced from unpublished maps supplied by R. W. Simpson (written commun., 1981), which combine the maps of Simpson, Bothner, and Godson (1981), and Simpson and Godson (1981), but use 125 km-wavelength instead of 100 km-wavelength.

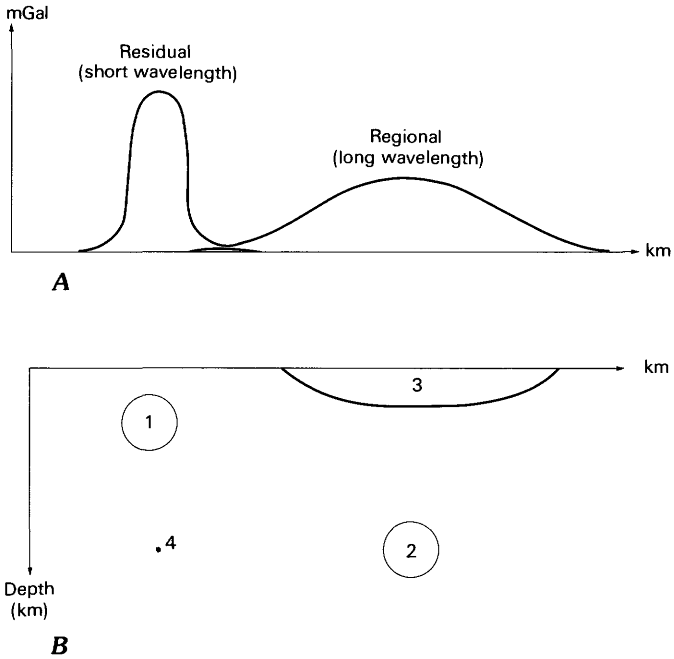


FIGURE 20.—Sketch illustrating interpretation of wavelengths of gravity anomalies in terms of the depths of the sources of the anomalies. *A*, Narrow (residual, or short wavelength) and wide (regional, or long wavelength) anomalies, respectively on the left and the right. *B*, Vertical cross section showing types of sources that might underlie and produce the anomalies of *A*. Sources 1–4 are denser than surrounding rock, and produce positive anomalies. Negative anomalies would be interpreted in the same way, but with sources that are less dense than surrounding rock. As the source of an anomaly deepens, the anomaly broadens. Source 2 is deeper than source 1 but otherwise identical, and produces a long-wavelength anomaly. However, such a long-wavelength anomaly can also be produced by a wide source like 3, at about the same depth as source 1. It is unlikely that a short-wavelength anomaly, such as that above source 1, could be produced by a deep source, such as 4. Such a deep source would have to be improbably small, improbably dense, or both. In practice, qualitative statements like the preceding ones are sharpened with numerical models calculated from specific values of densities, source depths and dimensions, and anomaly sizes and shapes.

crustal or shallower depths. The corresponding regional field contains anomalies from deeper sources, as well as those from shallow, wide sources, such as the sedimentary filling of the Appalachian Basin. Similarly, the residual field from the 250-km wavelength filter contains mostly anomalies arising from lower crustal and shallower sources; whereas, the corresponding regional field reflects deeper sources as well as shallow, wide sources (Simpson, Bothner and Godson, 1981; Simpson and Godson, 1981; Kane and others, 1981; R. W. Simpson, M. F. Kane, and W. H. Diment, oral and written communs., 1980, 1981).

Given that general association between anomaly wavelength and source depth, it is important for our purposes to note that the eastward rises in the unfiltered Bouguer field, in the regional field obtained from the 125-km filter, and in the regional field obtained from the 250-km filter, all coincide in map view in the region near Giles County (fig. 19). Locational mismatches between the eastward rises in the three fields have map dimensions usually less than half the map width of the rises themselves. We attribute such mismatches to the smoothing effects of the filtering process. We see no indication that the source of the rise migrates northwestward or southeastward with depth, although small amounts of such migration may be unresolvable at the scale of the maps we examined (1:5,000,000: R. W. Simpson, written commun., 1981). Thus, the source of the eastward gravity rise, which we and others infer to be the southeastern edge of relatively intact North American continental crust left from Iapetan opening, occurs at the same map position in both the upper and lower crust.

AREA OF EXPECTED OCCURRENCE OF IAPETAN NORMAL FAULTS

If a reactivated Iapetan normal fault or fault zone is responsible for the seismicity of the Giles County locale, where else in the Southeast might similar faults have formed? A foundation on which to build an answer is provided by the eastward gravity rise, and by its interpretation as the southeastern edge of the relatively intact continental crust that was left after Iapetan opening. The answer will consist of two estimated distances: how far to the southeast of the gravity rise, and how far to the northwest, Iapetan normal faults might have formed and been preserved until today.

SOUTHEASTWARD EXTENT OF IAPETAN NORMAL FAULTS

Recall that seismicity in and near Giles County occurs west of the gravity rise (fig. 19), beneath the sedimentary rocks that form the local thrust complex of the Valley and Ridge province (p. 5–8). That complex is the western tip of the much thicker thrust sheets of mostly metamorphic and igneous rocks that involve much of the upper crust of the southern Appalachians (Clark and others, 1978; Cook and others, 1979; Costain and Glover, 1980a; Cook and others, 1981; Cook and Oliver, 1981; Iverson and Smithson, 1982; Pratt and others, 1982) and perhaps farther north (Harris and Bayer, 1979, 1980, with discussion by Williams, 1980; Granger and others, 1980; Costain and Glover, 1980b;

but see also Ando and others, 1982; Taylor and Toksöz, 1982). If Giles County seismicity occurs on Iapetan normal faults, such faults will lie beneath the thrust complex and may be masked by thrusts on which upper crustal and shallower rocks have been transported to the west. Thus, Iapetan normal faults could exist at all crustal levels beneath the thrust sheets and at least as far east as the edge of relatively intact North American cratonic crust, which edge we consider to underlie the gravity rise.

We now consider whether an eastern boundary can be found for the area in which Iapetan normal faults may occur in cratonic crust. It is necessary for us to estimate separately regions of likely occurrence for Iapetan and Atlantic normal faults, even though both faulting episodes probably produced structures of comparable size, orientation, and style. The Appalachian thrustings and metamorphisms followed the Iapetan normal faulting but preceded the Atlantic normal faulting. Thus, the Iapetan and Atlantic normal faults could differ in properties, such as degree or type of annealing, which would affect their abilities to be reactivated in the present-day stress field. We will suggest that, in general, the eastward rise in the unfiltered Bouguer anomaly field is the eastern limit for Iapetan normal faults and that most such faults occur in the relatively intact North American crust west of the gravity rise. Local exceptions are possible, because the crust east of the rise is probably a heterogeneous mixture of many crustal pieces of many types. Some pieces may be parts of North American crust thinned or dissected by Iapetan normal faults. Most pieces may have been reworked by deformation and metamorphism during various Paleozoic subduction episodes. The reworking may have been so extensive that preexisting faults are no longer zones of crustal weakness.

Although the composition, thickness, and history of the crust east of the gravity rise are known only locally and approximately, it is now clear that much of that crust is not cratonic. For more than a decade (Brown, 1970), terranes of various sizes throughout the Appalachians have been shown or suggested to consist of Paleozoic island arcs, pieces of marginal or back-arc basins, or cratonic fragments with or without superimposed volcanic-plutonic edifices of Andean type. Examples of such terranes include Armorica (Van der Voo, 1979b, 1980a, 1982b) and various pieces of Avalon (for example, Simpson, Shride, and Bothner, 1980; Skehan and Murray, 1980; see review by Rast, 1980). Hatcher (1978), Long (1979), and Hatcher and Zietz (1980) inferred that various blocks of mafic, granitic, and mixed deep crust compose much of the southern Appalachians, including the region southeast of the gravity rise opposite Giles County. Osberg (1978) concluded that an island arc terrane comprises most of New England, and more terranes are being found or suggested at a quickening pace (Rowley, 1981; Spariosu and Kent, 1981; Williams and Hatcher, 1981; Zen and Palmer, 1981; Hatcher and Williams, 1982; Iverson and Smithson, 1982; Sinha and Zietz, 1982; Williams and Hatcher, 1982). Indeed, several workers (for example, Irving, 1979; Cook and others, 1981; Zen, 1981) considered a possible analog to be the mélange of over 50 distinct tectono-stratigraphic terranes that accreted onto western North America in Cenozoic, Mesozoic, and perhaps Paleozoic time after traveling unknown distances across the Pacific. (See reviews by Coney and others, 1980; and Ben-Avraham and others, 1981.) Some of these workers suggested, as a modern example, the complex of telescoping microplates and lithospheric shreds now caught between converging Australia and southeast Asia. (See maps by Hamilton, 1974a, b, c, 1978; Hayes, 1978.) Hatcher (1978) suggested, as another modern analog, the Pacific coast of Asia from Kamchatka to Japan and Korea, with its complex of peninsulas, island arcs, and marginal seas.

Further, the converging North American and Gondwanan continental margins of the Paleozoic were probably as irregular in map view as are present-day margins. If so, then geometric and geologic complexities like those inferred to be still developing in and around the Aegean Sea (Dewey and Şengör, 1979) may underlie one or more areas east of the gravity rise, opposite Giles County. Finally, the converging overall motion of the North American and Gondwanan plates may well have had irregular or strike-slip components. Such components would be most likely to occur as convergence ended and global plate motions began to reorganize to accommodate the loss of thousands of kilometers of subductive plate boundaries. If so, then much of the region east of the gravity rise may have evolved and accumulated throughout a history as complex as that suggested by Dewey and others (1973) for the Mediterranean region and the Alpine system.

Such known and suggested complexities in the tectonic evolution of the Appalachians can be used to postulate a resolution of an ostensible conflict. On the one hand, we hypothesize that the gravity rise marks the eastern edge of relatively intact North American continental crust, mostly formed in the Middle Proterozoic when the Grenville orogeny occurred. On the other hand, Glover and others (1978) identified exposed metamorphic rocks of Grenville age in central Virginia that lie east of the gravity rise. The rocks are involved in the thrust sheets of the Piedmont province, and so before thrusting they lay still farther east of the gravity rise. If these Grenville rocks were originally part of North America, they could have arrived east of the rise in various ways: the Grenville rocks could have remained attached to the relatively intact North American crust but linked to it by continental crust thinned by Iapetan normal faults; alternatively, the Grenville rocks could have been entirely separated from North America, by being rifted away (Hatcher, 1978; Hatcher and others, 1981; Glover and others, 1982), by strike-slip separation from a North American promontory, or by both, and later sutured back onto North America in their present relative position.

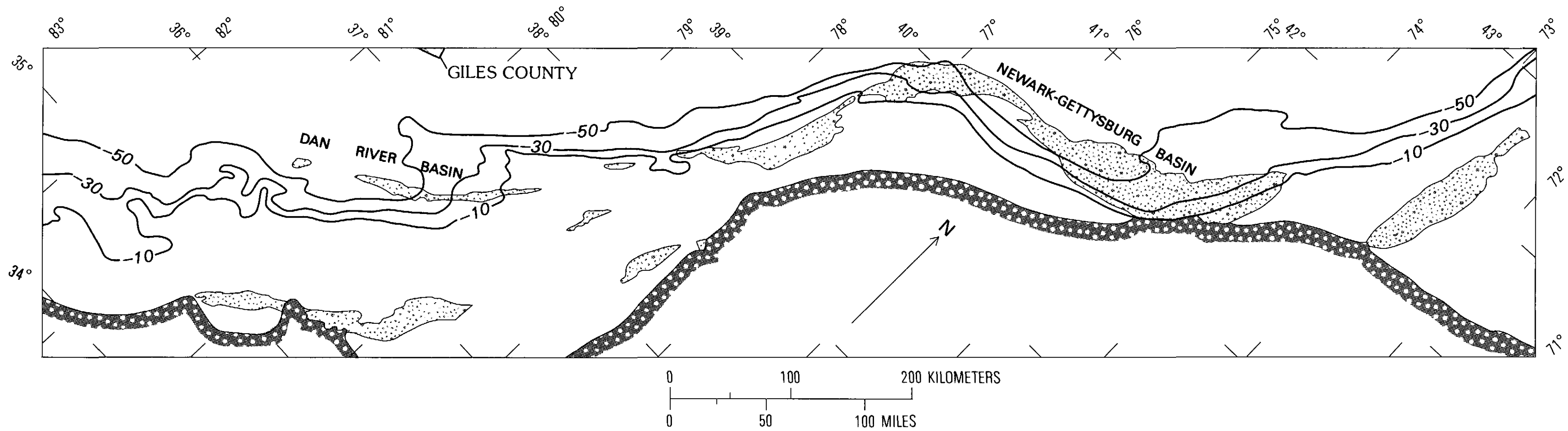
East of the gravity rise, such tectono-stratigraphic terranes could be of many sizes, shapes, and compositions. They are likely to be bounded and perhaps internally fragmented by plate-scale shear zones. Edges of pieces of continental crust could be further modified by Andean-type metamorphic and igneous activity. Further, any Iapetan normal faults that formed east of the gravity rise might no longer have an orientation suitable for reactivation in today's ambient stress field. That is, small plates could have been rotated when caught between larger plates carrying the North American and other cratons. Further rotation could have occurred during the many hundreds to several thousands of kilometers of left slip that is inferred to have occurred mostly or entirely in Carboniferous time (Kent and Opdyke, 1978, 1979; Van der Voo and others, 1979; Irving, 1979; Harland, 1980; Van der Voo, 1980a, b, 1981, 1982a, b; Kent, 1981; Van der Voo and Scotese, 1981; Williams and Hatcher, 1982). LeFort and Van der Voo (1981) suggested a model in which that left slip is consistent with the much smaller amount of coeval right slip inferred from the compilation of Bradley (1982). It seems unlikely that Iapetan normal faults would survive in such activity, at least not as weak zones of crustal size on which stress might be preferentially released by seismic slip.

If the crust east of the gravity rise is indeed an assemblage of heterogeneous terranes, it may be less cohesive or weaker than the cratonic crust west of the rise. Comparison of geologic and Bouguer gravity maps of the Eastern United States produces observations consistent with that suggestion (fig. 21). South from lat 43° N., Mesozoic extensional basins, most of them bounded by normal faults, lie on or east of the gravity rise, with two exceptions. The larger exception is the western part of the Newark-Gettysburg basin. However, there and elsewhere the western limit of the province of Mesozoic faults and associated basins follows faithfully abrupt bends in the gravity rise. The smaller exception is at about lat 37° N. where a sharp offset in the rise crosses the middle of the Dan River basin. Thus, the Mesozoic fragmentation of this part of the late Paleozoic supercontinent, Pangea, apparently followed and was restricted to the region suggested to be underlain by heterogeneous lithospheric fragments. It may be that those fragments are relatively weakly attached to each other and to the North American craton. Indeed, Grow and others (1982) independently suggested control of Mesozoic extensional faults by Paleozoic compressional structures.

NORTHWESTWARD EXTENT OF IAPETAN NORMAL FAULTS

It is reasonable to expect Iapetan normal faults to occur under and near the Giles County locale itself. The center of the rise in the unfiltered Bouguer anomaly field lies 50–100 km southeast of the locale (figs. 18, 19, 21). If the rise marks the eastern edge of relatively intact and unthinned North American cratonic crust, then analogies drawn from examination of present passive continental margins show that, in early Iapetan time, the locale was close enough to the lithospheric break that finally grew into the Iapetus Ocean that the Giles County locale could have experienced normal faulting. For example, on the edges of the Red Sea, Lowell and Genik (1972, their fig. 5) mapped normal faults that cut continental crust. On traverses across the Red Sea, as one approaches active and once-active spreading centers, such faults become abundant enough to have extended and thinned the crust. Lowell and Genik (1972) showed such faults occurring to about 100 km toward the craton from the seaward edge of relatively unthinned continental crust, and to some 270 km from the inferred boundary between new oceanic crust and old, fault-thinned continental crust.

Similarly, on and near the present-day United States continental margin off the central and southern Appalachians, the western edges of exposed, partly fault bordered Mesozoic basins show approximately how far into the pre-Atlantic continental crust of North America large normal faults formed when the Atlantic began to open. Continental crust is herein taken as extending east no farther than the western edge of the East Coast Magnetic High, which roughly follows the 2,000-m isobath between lat 31° and 40° N., 50–150 km offshore (Schouten and Klitgord, 1977; Grow and others, 1982). Over most of that area, continental crust is faulted but still relatively intact because it was apparently unthinned by Atlantic normal faults at least as far east as the overlap of the Coastal Plain onto Paleozoic rocks (boundary between Coastal Plain and Piedmont; James and others, 1968; Grow and others, 1982). However, offshore there are more normal faults (Sheridan, 1976; compilations by Wentworth and Mergner-Keefer, 1981a, b, c), and rift-stage crust becomes abundant at and east of the coast, or within 100–200 km of the East Coast Magnetic High (Klitgord and Behrendt, 1979). Thus, when the Atlantic opened, normal faults formed as



EXPLANATION

Areas of Mesozoic sedimentary rocks—Fillings of extensional basins

Southeast edge of exposed Paleozoic and Older rocks of Appalachians—South of about latitude 41°, the edge is at the onlap of Mesozoic and Cenozoic rocks and sediments of the Coastal Plain. North of that latitude, the edge is at the shoreline

—-10— **Selected contours of equal Bouguer gravity values**—Values are in milligals (1 gal = m/s^2). Contour interval is 20 mGal

Southeastern corner of Giles County

Figure 21.—Relationship of eastward rise in Bouguer gravity field to western edge of Mesozoic extension in central and southern Appalachians. Simplified and traced from Williams (1978) and Haworth and others (1980). Maps of Williams (1978) and King and Beikman (1974) show no Mesozoic extensional basins south or west of the area shown here. The three gravity contours that are shown are those that best depict the location and shape of the eastward gravity rise in the map area.

much as about 250 km to about 450 km inland from the present edge of the continental crust, and at least one-third to one-half of that distance represents normal faults in relatively intact continental crust. Similar values come from the margins of the Labrador Sea (Van der Linden, 1975), the Moroccan margin (Schlee, 1980), and several Australian, Red Sea, African, and Brazilian passive margins (Falvey, 1974; Talwani and others, 1979).

Thus, the area in which Iapetan normal faults are expected to exist and might experience seismic reactivation is bounded on the southeast by the eastward gravity rise, but has no sharp northwestern boundary. Southeast of the gravity rise, it is possible, but unlikely, that single Iapetan normal faults are preserved in a reactivatable state. To the northwest, Iapetan normal faults are expected to decrease in size, abundance, and slip gradually and irregularly northwestward into the North American craton over a distance of perhaps 100–200 km.

By analogy with other normal faults formed on passive continental margins, most Iapetan normal faults in eastern North America may be expected to strike northeast to north-northeast, particularly if they had not formed by reactivation of still older faults of diverse orientations. The Iapetan faults should dip steeply to either the northwest or the southeast. Where senses of net dip slip can be determined, most should still be normal. However, because the faults were properly oriented to have been reactivated in later compressional episodes of the Appalachian orogenies, some net dip slips could have been changed from normal to reverse if the original dip slips were small. Because today's greatest horizontal compressive stress trends northeasterly and not perpendicularly to the ancient Iapetan continental margin (Zoback and Zoback, 1980, 1981; this report, p. 51), seismic reactivation of such faults may (but need not) have a strike-slip component, probably right-slip. Such faults formed as the upper portions of fault systems that acted to extend the continental crust, and so should have dimensions comparable to the thickness of at least the brittle upper part of the crust.

SUMMARY

Of the three types of Paleozoic and Mesozoic basement faults that reasonably could have formed under the Giles County locale and that may be responsible for much of its present seismicity, we consider an Iapetan normal fault to be the most probable. Before considering the other two fault types, we summarize here reasons for favoring Iapetan normal faults. The Giles County locale is well within the region of North American continental crust expected to have undergone such faulting: west of but within 100–200 km of the Iapetan continental edge that is inferred to underlie the steep eastward rise in the unfiltered Bouguer anomaly field. The Giles County seismic zone has the proper orientation, shape, size, and depth range to be occurring on such a fault—reactivated in today's ambient stress field. Sparse direction-of-motion data on P waves are unable to give a composite focal mechanism by themselves but are consistent with right-reverse reactivation of such a fault (fig. 16). Finally and reassuringly for the evaluation of such subtly expressed and well-hidden structure, we know of no evidence that is inconsistent with the hypothesis that an Iapetan normal fault is responsible for the Giles County seismic zone.

ALLEGHANY THRUST-LOAD FAULTS

Late Paleozoic faults of a type here named "thrust-load" could be sources of Giles County seismicity. The likelihood that a thrust-load fault is responsible for the Giles County seismic zone will be evaluated by comparing relative ages of central and southern Appalachian thrusting in and near Giles County. The hypothesis of a thrust-load fault allows relative ages to be deduced from observed map relations, and that deduction can be tested against relative ages of thrusting inferred from stratigraphic and structural observations.

Thrust-load faults are hypothesized to form in front of or beneath recently emplaced thrust sheets, as the crust fractures under their weight in a brittle analog of the foredeeps known to form under and in front of thrust masses and continental ice sheets. Alternatively, thrust-load faulting may occur by reactivation of older basement faults that are suitably oriented, again under the load imposed by newly emplaced detached masses (W. G. Brown, oral communs., 1980, 1981; Berry and Trumbly, 1968; Buchanan and Johnson, 1968; Hopkins, 1968; Beiers, 1976; Bush and others, 1978; M. K.-Seguin, oral and written communs., 1981; Seguin, 1982).

A TESTABLE DEDUCTION FROM THE THRUST-LOAD HYPOTHESIS

The Giles County locale and its N. 44° E. striking seismic zone lie in the western part of the Valley and Ridge province (fig. 2) of the southern Appalachians. The locale is also near the juncture of the southern and central Appalachians. The southern Appalachians are characterized by east-northeast-trending thrust faults and related structures (fig. 22); in contrast, the central Appalachians are characterized by north-northeast

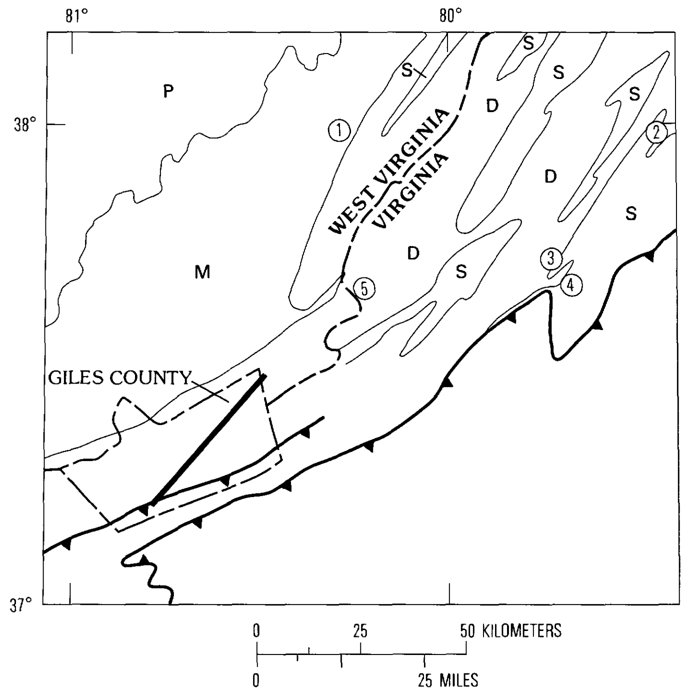


FIGURE 22.—Approximate orientations of Giles County seismic zone and of central and southern Appalachian thrust structures. Thick solid line trending about N. 44° E. shows approximate center of the seismic zone. Southern Appalachian structures trend east-northeast as indicated by traces of outcrops of main thrust faults (solid lines with sawteeth on upper plates). Central Appalachian structures trend north-northeast as indicated by main folds as outlined by traces of systemic boundaries (S, Silurian and older rocks; D, Devonian rocks; M, Mississippian rocks; P, Pennsylvanian rocks). Systemic boundaries southeast of thrust-fault traces are complex and not shown here. Geology and structure simplified from compilation of Willden and others (1968). Circled numbers show localities discussed in text.

trending thrust-related folds and other structures. Thrust faults and folds of both central and southern Appalachians involve Mississippian and older rocks, and to the northwest of Giles County, Pennsylvanian and Permian rocks. Many of the pre-Pennsylvanian rocks contain polymictic conglomerates that record the formation and erosion of substantial structural and topographic relief at several times and places. However, in and near the Giles County locale only the Pennsylvanian and Permian strata contain abundant, immature synorogenic clastic debris: molasse, derived from the southeast. Accordingly, the thrust masses presently exposed above the seismic zone are regarded as having been emplaced in the Giles County locale during the Alleghany orogeny of Pennsylvanian and Permian time. Older thrusts are known or possible, especially farther southeast, because thrusting developed successively northwestward (Perry, 1978). Any such older thrusts could have propagated northwestward, under Giles County, as blind thrusts (Boyer and Elliott, 1982, p. 1197). Thus, the thrusts now exposed in Giles County may have begun to form and move before Alleghany time. However, such earlier events would not affect the conclusion that the thrust-transported near-surface rocks and structures of the Giles County locale probably arrived above the seismic zone in Pennsylvanian or Permian time.

If the seismic zone occurs on a thrust-load fault, then that fault is probably either an Iapetan normal fault reactivated in Alleghany time, or a fresh crustal break of Alleghany age; recall that the introduction to this section suggests that high-angle faults older than Iapetan are unlikely to have survived under Giles County in any form able to be reactivated seismically. If thrust loading reactivated an Iapetan normal fault, the basement beneath the thrust masses would have to extend horizontally, in a direction at high angles to the strike of the reactivated normal fault. Fleitout and Froidevaux (1982, p. 43, figs. 5, 7) suggested a theoretical mechanism by which gravitational loading could produce a small amount of horizontal extension. Their suggestion is that emplacement of a thrust sheet would load and depress the lithosphere. Depression of the lower crust into the upper mantle would create a buoyant effect. However, if thrust loading and depression occurred fast enough that the depression occurred adiabatically, then depression of the upper mantle into the warmer, underlying mantle would create a sinking effect. If the sinking effect exceeded the buoyant effect, the crust could be drawn slightly in toward the center of the loaded area. The result could be horizontal extension under or near the loading thrust sheet. We do not argue that such horizontal extension under thrust loading actually occurred under Giles County, but we note that it is theoretically possible, and so must be considered here.

On the other hand, if the hypothesized thrust-load fault formed as a fresh Alleghany-age fracture, its strike should follow approximately the strike of the causative load gradient. In turn, the strike of the load gradient should follow the trends of thickness contours of the causative thrust complexes. Thickness contours follow trends of depth to basement, stratigraphic levels exposed today by erosion, and sedimentary facies, none of which change abruptly along strike in the region surrounding Giles County (Colton, 1970; King and Beikman, 1974). Consequently, the strike of any thrust-load fault that formed as a fresh Alleghany-age fracture should follow the general structural and stratigraphic trends of the exposed remnants of the causative thrust sheets.

The strike of the seismic zone is of an unambiguous central Appalachian orientation, rather than of a southern Appalachian one (fig. 22), so the hypothesized thrust-load fault would have been caused by emplacement of central Appalachian thrust sheets. However, the thrust and related structures now exposed in and near Giles County have southern Appalachian orientations (fig. 22). Those southern Appalachian structures are not known to be cut or otherwise affected by movement on the hypothesized thrust-load fault. Therefore, the thrust-load fault would have formed before arrival of the southern Appalachian thrust sheets that now overlie the seismic zone. This reasoning implies that central Appalachian thrust sheets arrived in or near Giles County first and were eroded before arrival of the southern Appalachian sheets, or were rotated by them or buried by them. Thus, the thrust-load hypothesis leads to the deduction that, in the Giles County locale, arrival of central Appalachian thrust sheets predates that of southern Appalachian thrust sheets.

This deduction can be tested. Relative ages of central and southern Appalachian thrusting are not clearly known, but several independent lines of structural and stratigraphic evidence are summarized in the following paragraphs. All favor southern Appalachian thrust sheets as having reached the vicinity of Giles County slightly before those of central Appalachian orientations. This conclusion contradicts the faulting sequence as deduced from the thrust-load hypothesis and, therefore, negates that hypothesis.

STRATIGRAPHIC TESTS

The straightforward stratigraphic approach of constraining the age of thrusting by determining the ages of youngest folded and oldest unfolded rocks cannot work here. Youngest folded rocks are Early Permian in the central Appalachians and are Middle or Late Pennsylvanian in the southern Appalachians (King and Beikman, 1974). However, Permian rocks are wholly eroded or were never deposited in the southern Appalachians, and Middle and Upper Pennsylvanian rocks are nearly as sparse there (Colton, 1970, p. 42; King and Beikman, 1974). On the other hand, more subtle stratigraphic arguments are fruitful. Arkle (1969, 1972, 1974) presented isopach and facies maps and current-direction data for units of middle Mississippian through latest Pennsylvanian or Early Permian ages. These maps and related information allow us to estimate relative ages of thrusting, by dating the main influxes of Pennsylvanian molasse. The following analysis and interpretation are consistent with those done independently by Donaldson and Shumaker (1981). Their analysis is more detailed and covers more of the Paleozoic and a larger region than does ours.

Pennsylvanian and Permian rocks of the central and adjacent southern Appalachians are "a series of shales and fine- to coarse-grained sandstones, locally conglomeratic, arranged in repetitious sequences with thinner coals, clays, lacustrine and marine limestones, chert and ironstone" (Arkle, 1974, p. 5). Pertinent stratigraphic names are summarized in figure 23. The sandstones are immature, and the various lithologies record terrestrial, fluvial, deltaic, and some shallow marine deposition (Meckel, 1970; Donaldson, 1974; Arkle, 1974; Horne and others, 1978). The sequence is synorogenic and records the topographic and erosional effects of emplacement of late Paleozoic Alleghany thrust sheets. At least the parts of those sheets that were close to areas of molasse deposition must have been the tops of the detached sedimentary fold-and-thrust complexes now exposed in the eastern Appalachian Plateau and the Valley and Ridge provinces. Farther southeast, the metamorphic and igneous rocks of the Appalachians were also being unroofed and dissected (Presley, 1981). Davis and Ehrlich (1974) inferred from petrography of metamorphic and igneous grains and rock fragments in the Pennsylvanian sandstones that, in Early Pennsylvanian time, sedimentary and volcanic debris accumulated from initial unroofing of the metamorphic and igneous rocks. Next, successively deeper erosion and the required kilometers of uplift shed debris first from low-grade metamorphic rocks, then from batholithic complexes, and finally by Late Pennsylvanian time, from the underlying migmatitic terrane.

The deposition of Pennsylvanian and Permian sediments occurred in two overlapping basins separated by a wide, diffuse hinge line (fig. 24; Arkle, 1969, 1972; Horne and others, 1978). Upper Mississippian (Englund and others, 1982) and Lower to Middle Pennsylvanian clastic debris flowed in from the southeast and accumulated mostly in a subsiding trough called the Pocahontas basin, southeast of the hinge line (fig. 24*A*, *B*). Lower Pennsylvanian rocks of the Pottsville Group (figs. 23, 24) north of the hinge line are much thinner than are correlative rocks south of the hinge line (fig. 24*B*, *C*). In Middle and Late Pennsylvanian time, the clastic sources lay to the east and northeast and much thinner units accumulated mostly on a stable platform called the Dunkard basin, northwest of the hinge line (fig. 24*D*). Williams and Bragonier (1974) documented a southeastern source for Early Pennsylvanian time in much of western Pennsylvania, but, even so, thicknesses were much less there than southeast of the hinge line in southeastern West Virginia.

The Pottsville Group and the overlying units of the northern coal field both have northeastern sediment sources, and isopach lines indicate southwestward

System	Series	Stratigraphic Units: Northern coal field	Stratigraphic Units: Southern coal field
Permian	Lower	Dunkard Group	
Permian or Pennsylvanian			
Pennsylvanian	Upper	Monongahela Formation 7	
		Conemaugh Formation 6	
	Middle		Charleston Sandstone 3
		Allegheny Formation 5	Kanawha Formation 2
	Lower	Pottsville Group 4	New River Formation, Pocahontas Formation 1
		NW SE	NW SE

FIGURE 23.—Permian and Pennsylvanian stratigraphy of West Virginia coal fields. Sources: Englund and others (1979) and Arkle (1974); also Berryhill and Swanson (1962) and Cardwell and others (1968), unless there is a conflict with the two newer sources. Nondeposition and erosion followed deposition of the Dunkard Group and Charleston Sandstone. In both coal fields, basal Pennsylvanian strata are conformable on Mississippian beds southeast of the hinge line of figure 24 and unconformable northwest of it. Circled numbers refer to isopachs of figure 24. Note that "Allegheny Formation" is spelled with an e; whereas, "Alleghany orogeny" takes an a (Rodgers, 1970, p. 30).

thinning for both sequences (fig. 24*C*, *D*). However, in western Pennsylvania the Pottsville Group also resembles the approximate correlative units in the southern coal field, because both received sediments from the southeast and both have isopachs that show northward thinning (fig. 24*B*, *C*). Thus, the Pottsville represents a transition between an older sediment source in the southern Appalachians, and a younger source in the central Appalachians.

Thus, the sedimentary record of Alleghany tectonism indicates an older, Early and Middle Pennsylvanian (Pocahontas, New River, and Kanawha Formations, and Charleston Sandstone) age in the southern Appalachians near Giles County, but a younger, Middle and Late Pennsylvanian (Alleghany, Conemaugh, and Monongahela Formations) age in the central Appalachians.

The greater age of the southern Appalachian deformation compared to that of the central Appalachians can be demonstrated by additional stratigraphic evidence. First, in eastern Pennsylvania, a central Appalachian synclinorium contains tightly folded coal measures of Middle and Late Pennsylvanian age—all deformed by the oldest of the Alleghany structures (Wood and Bergin, 1970). Thus, in that portion of the central Appalachians, Alleghany deformation occurred during or after Late Pennsylvanian time.

Second, Horne and others (1978) and Cavorac and others (1964) reported an abrupt southward thickening of lowest Pennsylvanian (lower Pottsville Group) strata across the east-striking Irvine–Paint Creek fault system

FIGURE 24 (facing page).—Distributional patterns of Pennsylvanian units in West Virginia and parts of adjacent States. See figure 23 for stratigraphy. Double broken line shows position of hinge line of Arkle (1969, 1972), separating northern and southern coal fields of central Appalachians and figure 23. Location of hinge line is approximate: Donaldson (1974) gives its width as 25–50 mi (50–80 km). In parts *B*, *C*, and *D* heavy lines show isopachs selected from the maps of Arkle (1974). Circled numerals at ends of isopachs match the isopachs with their units, numbered in figure 23 from oldest to youngest. Isopachs shown here were selected to summarize the approximate present shapes and thinning directions of the units as shown in the more detailed maps of Arkle (1974). Thickness values of the selected isopachs are shown next to them, and for Parts *B*, *C*, and *D* are variously one-third to three-fourths the largest values shown on Arkle's maps. Boxed numerals show approximate locations of maximum thicknesses of indicated units in the area shown here, as inferred from map patterns of facies distributions and from the isopach patterns shown and discussed by Arkle (1969, 1972, 1974). Arrows on isopachs indicate approximate directions of sediment flow and unit thinning.

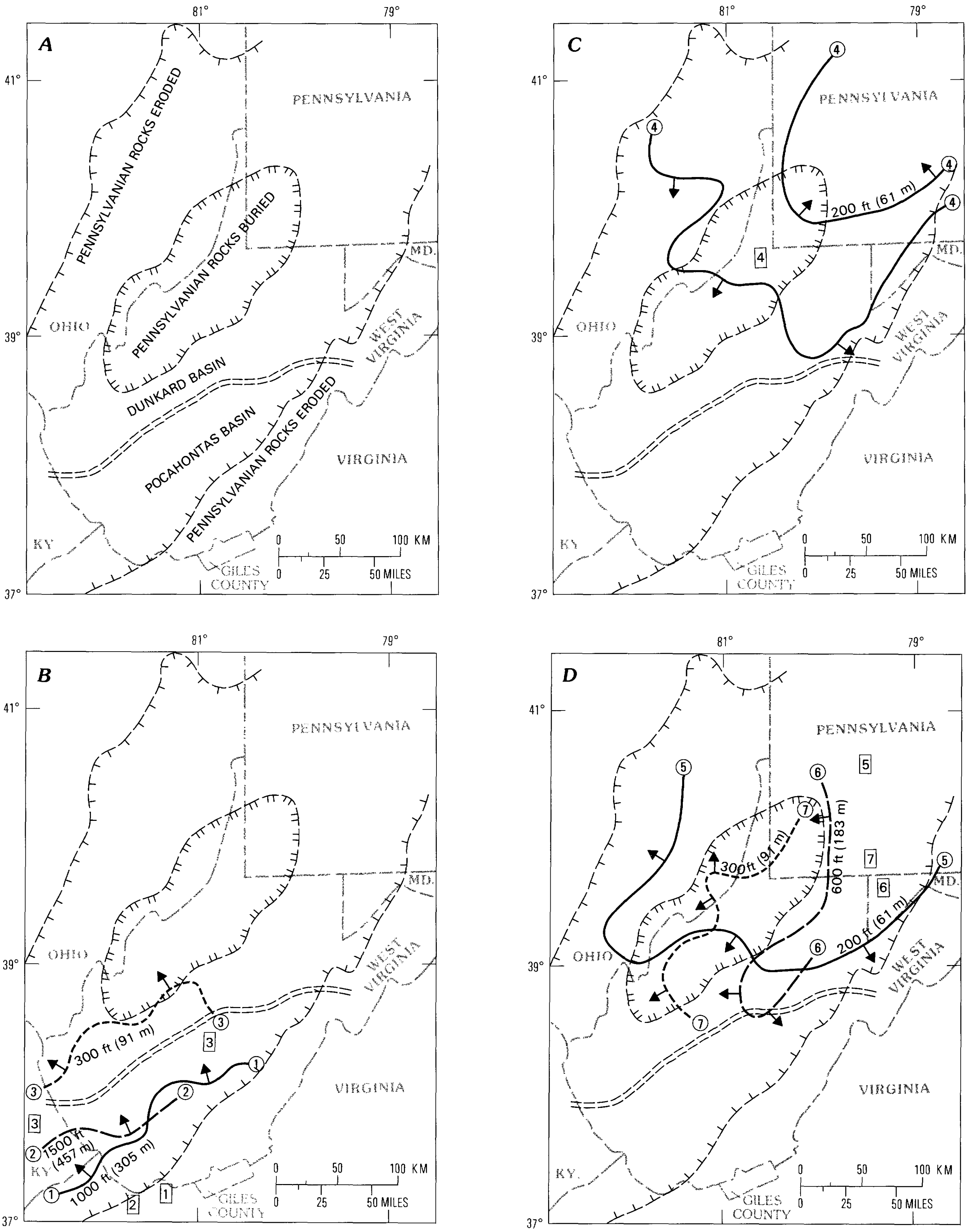
A. Locations of outcrops of base (single hachures) and top (double hachures) of Pennsylvanian System, greatly simplified. Hachures point inward toward center of late Paleozoic Dunkard basin.

B. Distributional patterns of Lower and Middle Pennsylvanian units, which entered the southern coal field of the Pocahontas basin from a southeastern source.

C. Distributional patterns of Lower and lower Middle Pennsylvanian Pottsville Group of the northern coal field, which is approximately correlative with most of the sequences represented in part *B*. Sediment entered the northern coal field of the Dunkard basin mostly from a northeastern source, but also with influx from a source in the southeast in western Pennsylvania (Williams and Bragonier, 1974).

D. Distributional patterns of Middle and Upper Pennsylvanian units, which entered the northern coal field of the Dunkard basin from eastern and northeastern sources. Arkle (1974) also shows isopach and facies maps of two Permian units. Their distributional patterns are consistent with those shown here but the Permian units are preserved over such small areas that they are not represented here.

of eastern Kentucky (fig. 17). Moreover, the southeastward tilting of depositional surfaces south of the hinge line occurred in Early Pennsylvanian time. The southeastward tilting possibly was a crustal response to loading by advancing southern Appalachian thrust sheets, and the activity on the Irvine–Paint Creek faults possibly was reactivation of older faults by thrust-loading. Later, a marine transgression, known to have occurred in Conemaugh time, could have been a response to formation of a foredeep by central Appalachian thrust-loading (Merrill, 1981). Such early southeastward tilting in front of advancing southern Appalachian thrust sheets, and later, northeast-trending depression in front of advancing central Appalachian thrust sheets, would be consistent with observations of Kulander and others (1980). Within the outcrop belt of Mississippian rocks that is shown in figure 22, Kulander and others mapped extension


A
81°
79°
41°
39°
37°
PENNSYLVANIA
PENNSYLVANIAN ROCKS ERODED
PENNSYLVANIAN ROCKS BURIED
OHIO
MD.
DUNKARD BASIN
POCAHONTAS BASIN
PENNSYLVANIAN ROCKS ERODED
WEST VIRGINIA
VIRGINIA
KY
GILES COUNTY
0 50 100 KM
0 25 50 MILES
C
PENNSYLVANIA
200 ft (61 m)
4
OHIO
MD.
WEST VIRGINIA
VIRGINIA
KY
GILES COUNTY
B
PENNSYLVANIA
OHIO
MD.
WEST VIRGINIA
300 ft (91 m)
3
1
2
1500 ft (457 m)
1000 ft (305 m)
VIRGINIA
KY
GILES COUNTY
D
PENNSYLVANIA
5
6
7
300ft (91 m)
600 ft (183 m)
200 ft (61 m)
OHIO
MD.
WEST VIRGINIA
VIRGINIA
KY.
GILES COUNTY

fractures that strike north-northeast and east-northeast. The fractures formed later than middle Mississippian time because they occur in limestone of that age. However, the extension fractures formed before Alleghany folding of the limestone because they are overprinted by stylolites that are a precursor of that folding. The extension fractures may record tilting and flexing of the crust in response to thrust-loading (S. L. Dean, oral commun., 1981). Thus, we have evidence which can be interpreted to indicate that the southern Appalachian thrust sheets were emplaced earlier in Pennsylvanian time than were the central Appalachian thrust sheets.

Finally, Babu and others (1973) compiled and mapped data on compositions of West Virginia coals. Coal rank and fixed-carbon content are properties whose values are affected mostly by postdepositional processes, such as burial, tectonism, and metamorphism. Accordingly, one would expect contours based on coal rank and fixed-carbon content to follow Appalachian structural trends—and they do (Babu and others, 1973, fig. 2). Values of both variables increase to the south-southeast in southeastern West Virginia, adjacent to the east-northeast-trending southern Appalachians, and values increase to the east-southeast in northern West Virginia, adjacent to and in the north-northeast-trending central Appalachians. However, ash and sulfur contents are known to reflect the local depositional environment and details of paleotopography in the coal swamp, and so ash and sulfur contents can record trends of thrust-related anticlines that were growing during deposition (Donaldson, 1974, p. 73). Figures 3 and 4 of Babu and others (1973) are too generalized to reflect locations and orientations of individual folds, but contours derived from ash and sulfur contents have crude southern Appalachian trends in the southern coal field, where most coals are of Early Pennsylvanian age, and the contours have crude central Appalachian trends in the northern coal field (see also Kent and Gomez, 1971), where most coals are of Middle and Late Pennsylvanian age. Thus, the combined data on coal composition and age support the evidence of a paleotopography that was created by growth of thrust-related anticlines formed in Early and Middle Pennsylvanian time in the southern Appalachians, and in Middle and Late Pennsylvanian time in the central Appalachians.

STRUCTURAL TESTS

Sparse structural information supports or is consistent with the relative ages inferred from interpretations of stratigraphic results. The clearest structural information is that from the Mississippian Greenbrier Limestone in both the central and southern Appalachians, along a strike belt about 120 mi (190 km) long roughly centered at the location indicated by the circled numeral 1 in figure 22 (Dean and Kulander, 1977, 1978; Dean and others, 1979; Skinner, 1979; S. L. Dean, oral commun., 1981). Along the strike belt, stylolites formed in the Greenbrier Limestone on preexisting systematic joints. Stylolites and joints were rotated by folds that are visible on maps at scales from 1:24,000 to 1:250,000. Stylolites form as the rock dissolves under compressive stress, and the teeth on the stylolite seam form with their long axes parallel to that compressive stress. The stylolites in the Greenbrier Limestone have slickenlines (grooves or striae on slickensided surfaces (Fleuty, 1975)) parallel to the teeth. Over a large area extending about 50 mi (80 km) into the central Appalachians, stylolite teeth trend north-northwest and so they parallel other southern Appalachian structures. In the central Appalachians, in the southwestern part of the Williamsburg anticline (locality 1, fig. 22; and fig. 17), the stylolites have a southern Appalachian orientation and are folded by the anticline. Further, central Appalachian stylolites and slickenlines overprint those of southern Appalachian orientation, and a central Appalachian axial cleavage can be traced southwest, and there cuts obliquely and with constant orientation across southern Appalachian folds (S. L. Dean, oral commun., 1981).

However, interpretations of larger interfering structures are still too few to be conclusive. Perry (1978, p. 525–526) reinterpreted map patterns published by Bick (1973) at scales of about 1:40,000 to about 1:100,000 (localities 2–4, fig. 22 of this report). For locality 2, Perry concluded that the map by Bick records a thrust fault of southern Appalachian orientation that has been folded by an anticline of central Appalachian orientation. Bick agrees with this conclusion (written communs., 1978, 1981). For localities 3 and 4, Perry interpreted Bick's maps to show other central and southern Appalachian structures interfering with each other; however, subsequent mapping indicates that the interpreted interference does not or may not occur at those two localities (K. F. Bick, written commun., 1981). More recently near locality 4 (fig. 22), Bick (1982, 1986) and Henika and others (1982) interpreted an interference structure involving the Purgatory Mountain anticline and the Pulaski thrust sheet (fig. 17). They deduced different relative ages using different data. In another study of the area at the junction of the southern and central Appalachians, Olson (1979) mapped, at a scale of 1:24,000, folds of central Appalachian orientation lying northwest of and trending into the southern Appalachian St. Clair fault (figs. 17, 22, locality 5). Olson (1979, p. 88) concluded that one such unnamed anticline-syncline pair is not truncated by the thrust, but his map (his plate 1A, northeast half) suggests to Wheeler that the thrust may cut the fold pair. At the northeast end of the thrust, a second fold pair mapped by Olson may be folded, though not necessarily cut, by the thrust. Reconnaissance mapping by McDowell tends to support the interpretation of Olson (R. C. McDowell, 1981; oral commun., 1981). Thus, of six places where map-scale structures of southern and central Appalachian orientations are known or suspected to interfere, one shows an older southern Appalachian structure and five are inconclusive at present.

Such inconclusive relative ages as determined from map-scale structures are discouraging but, on reflection, not surprising. The sequence of deformation in the sedimentary parts of the central Appalachians and adjacent parts of the southern Appalachians is known to have been long and complex on both map and outcrop scales (Geiser, 1977, 1981; Perry and deWitt, 1977, p. 39–40; Perry, 1978; Bartholomew, 1979; Nickelsen, 1963, 1979, 1980; Van der Voo, 1979a, b; Berger and others, 1979; Hatcher and Odom, 1980; Roeder and Boyer, 1981; Wright, 1981; Bartholomew and others, 1982; Bick, 1982; Gray, 1982; Henika and others, 1982; Webb, 1982; Wheeler, 1982, 1986). Dahlstrom (1970) documented type examples in the foothills of the Canadian Rockies where the typical sequence of relative ages of thrusts, becoming younger toward the craton, is locally reversed. Roeder and others (1978) and Witherspoon and Roeder (1981) interpreted a series of partly balanced cross sections through the southern Appalachians as recording complex polyphase thrusting with similar local reversals. Given such complex internal deformation of a thrust complex like the Valley and Ridge and eastern Appalachian Plateau provinces, it seems likely that many more local map relations will be needed before relative ages of structures become clear. Such structural complexity is compounded because the central and southern Appalachians overlapped in time for part of their growth. Overlap can be inferred with particular clarity from the paleogeographic maps of Donaldson and Shumaker (1981). Mapping underway by several workers, such as K. F. Bick, S. L. Dean, B. R. Kulander, and R. C. McDowell (oral and written communs., 1978–1982), will eventually produce the clearest, most detailed determinations of relative ages. However, now and for our purposes, the chronology of folded stylolites of Dean and Kulander (1977, 1978), and conclusions drawn from the stratigraphic data mapped by Arkle (1969, 1972, 1974) and compiled by Donaldson and Shumaker (1981), probably give more reliable relative ages because both approaches average the results over large areas.

SUMMARY

The thrust-load hypothesis leads to the deduction that central Appalachian thrust sheets entered or formed in Giles County before those of the southern Appalachians. However, that conclusion is contradicted by relative ages inferred from stratigraphic and outcrop-scale structural data. We conclude that the Giles County seismic zone did not form originally as a fresh Alleghany fracture or fracture zone under thrust loading. Recall that this line of reasoning assumes that such a fresh break would parallel the causative thrust front. Thrust loading by southern Appalachian thrust sheets could have reactivated an older basement fault. If such an older fault were weak enough, it could be reactivated even if it made an angle of several tens of degrees with the front of the loading thrust sheets. Such an older fault would probably have originated as an Iapetan normal fault, according to the arguments and conclusions previously stated in this report (p. 27–28).

ATLANTIC NORMAL FAULTS

Mesozoic normal faults could be sources of Giles County seismicity. Such faults formed to extend Pangean continental crust as the Atlantic Ocean began to open. The faults now bound Mesozoic grabens and half grabens that are exposed at least from Massachusetts to South Carolina (for instance see figure 18; King and Beikman, 1974). The normal faults are known or inferred to bound Mesozoic basins. The basins have been detected as far east as the edge of the continental shelf by geophysical methods and by drilling through younger sediments and sedimentary rocks. (See compilation of Wentworth and Mergner-Keefer, 1981a, b, c, and references cited there.) The faults can be high angle, with net normal slip. Most strikes trend from east-northeast to north-northeast, but a few short segments strike northwesterly and subdivide or terminate some of the basins (King and Beikman, 1974). Dips can be to either side of the strike. At least some faults formed by reactivation of older faults (Ratcliffe, 1971). Near New York City, Aggarwal and Sykes (1978) inferred that some such Mesozoic faults are seismically active, although Ratcliffe (1981a, b, c) questioned that inference.

Atlantic normal faults of Mesozoic age are unlikely to occur in or under the Giles County locale for four reasons: First, exposures of Mesozoic sedimentary rocks of the types that fill the Mesozoic grabens and half-grabens are unknown as far northwest as the Giles County locale (fig. 18; this has also been noted by Bollinger, 1981a). Second, large faults with normal slip that cut Alleghany thrust structures are unknown as far northwest as the Giles County locale; a Mesozoic normal fault of small enough extent to have escaped detection thus far would probably be too small to generate a damaging earthquake, even if erosion had removed all evidence of any Mesozoic sediment that might have accumulated on the downdropped block. Third, as noted previously under "Iapetan Normal Faults," most Mesozoic faults and basins in the Eastern United States lie on or east of the steep eastward rise in the Bouguer anomaly field. (For the region near the Giles County locale, figure 21 shows this relationship.) We have previously suggested that Mesozoic extensional faulting was restricted to weaker, more heterogeneous crust east of the gravity rise; therefore, the Mesozoic faults should not be expected in the relatively intact North American cratonic crust inferred to lie west of the rise. Giles County is about 50 mi (80 km) northwest of the gravity rise, and the nearest known Mesozoic basin is the fault-bounded Dan River basin, about 70 mi (110 km) southeast of Giles County (figs. 17, 21). Fourth, it is possible, although structurally and stratigraphically improbable, that beneath the Giles County locale the very last movements on Alleghany thrust faults occurred in earliest Mesozoic time and buried Permian or very early Mesozoic basins that have bounding faults. However, we know of no evidence for such Permian or Mesozoic basins under the thrust sheets—The sedimentary fillings of Mesozoic basins of the Atlantic seaboard of the Southeastern United States have compressional velocities from 4.4 km/s (or less) to 4.85 km/s, although higher velocities are possible with admixtures of basalt (Stewart and others, 1973; Daniels, 1974; Ackermann, 1977; Talwani, 1977; Behrendt and others, 1981); however, for the Giles County locale, Bollinger and others (1980) found no compressional velocities less than 5.33 km/s, and the velocity usually used for depths down to 5.7 km is 5.63 km/s (table 3, this report). For these four reasons, we reject the hypothesis that Atlantic normal faults could be responsible for Giles County seismicity.

OTHER FAULT TYPES

Other fault types that cannot be conclusively ruled out as candidates for Giles County seismicity include (1) those associated with formation of a back-arc basin in response to subduction connected with one of the Appalachian Paleozoic orogenies, and (2) a continental rift such as that associated with present seismicity in the head of the Mississippi embayment (Russ, 1981). However, we know of no data to suggest that either process operated near the Giles County locale. The nearest fault of either type is the graben known as the Rome trough of western Pennsylvania, western West Virginia, and eastern Kentucky (fig. 25; Harris, 1975, 1978; Shumaker, 1977; Kulander and Dean, 1978b; Ammerman and Keller, 1979). However, the southeastern border fault or faults of the Rome trough have a striking aeromagnetic signature that trends northeasterly and forms part of the New York–Alabama magnetic lineament (fig. 17) of King and Zietz (1978) and Zietz and others (1980). The magnetic lineament is about 60 mi (100 km) northwest of Giles County (U.S. Geological Survey, 1978).

It seems worth stating explicitly that we know of no reason to suspect that any seismicity in or near Giles County occurs on thrust faults such as those known or suggested to be sources of earthquakes elsewhere (Suppe, 1981; Seeber and Armbruster, 1979, 1981; Armbruster and Seeber, 1981; Behrendt and others, 1981). There are three arguments against seismic reactivation of thrust faults in the Giles County locale: First, all well-determined hypocentral depths in the locale lie in pre-Appalachian metamorphic and igneous basement, below the deepest known Appalachian thrusts. Second, a deeper Appalachian thrust, in basement, such as thrusts found farther south and southeast and in other mountain ranges, is unknown in or near Giles County. Third, the Giles County seismic zone itself dips too steeply over too great a depth range—and is too tabular—to be part of a thrust complex.

There is other seismicity in the Giles County locale that is not a part of the Giles County seismic zone itself. It is possible that some of that other seismicity could occur on a pre-Appalachian thrust fault within the basement below any Appalachian thrusts. Presumably, such a deeper detachment would have originated during the Grenville orogeny of Middle Proterozoic age. Although the existence of such a fault cannot be ruled out by present evidence, neither can we see any seismological, geological, or geophysical reason to postulate it.

SUMMARY

Of the four kinds of faults that might have been reactivated to produce the Giles County seismic zone, Iapetan normal faults best fit local geological, geophysical, and locational evidence. Faults formed during the Grenville orogeny or older faults are unlikely to have survived until now in a condition that would allow them to generate earthquakes. The stratigraphic and structural arguments against a seismogenic thrust-load fault are not conclusive. However, we are not aware of any place where thrust loading has been clearly shown to have formed fresh fractures in previously unfaulted crust. Rather, two other responses to thrust loading seem to occur instead or in addition: the formation of broad foredeeps by crustal downwarping and reactivation of older normal faults that date from the early opening of an ocean. For the Giles County locale, the Pocahontas basin may be such a foredeep and any reactivated faults would, therefore, be Iapetan normal faults. Finally, Mesozoic normal faults are still less likely candidates for the source of Giles County seismicity because they only occur far to the southeast of the Giles County locale.

STATE OF STRESS IN THE GILES COUNTY, VIRGINIA, LOCALE

INTRODUCTION

The orientation of the greatest horizontal compressive stress acting today on the Giles County seismic zone is estimated from selected stress measurements. We show the estimated orientation of greatest stress is consistent with the orientation and sense of seismic slip on the zone, as inferred from *P*-wave polarities and hypocentral distributions.

STRESS ORIENTATIONS

We are aware of three compilations of measurements of stress orientations for part or all of the region surrounding Giles County. Overbey (1976) compiled stress orientations measured at sites from southwestern New York State to eastern Kentucky, all in the Appalachian Plateau province and adjacent foreland. His review paper does not evaluate the different methods used, and his tabled and mapped orientations range over 61° of azimuth. Because the reliability of stress orientations and their applicability to seismogenic depths are more important for our purposes than the number of sites at which the orientation is measured, we did not investigate the original sources of Overbey. We relied instead on the two more recent compilations.

Zoback and Zoback (1980) reviewed and evaluated measurements and estimates of stress orientations for the conterminous United States. They compiled and annotated those that they considered reliable. Two of the orientations meet all the criteria that we imposed, which we will discuss later. The measurements at well OH-1 in southeastern Ohio, and at well 20402 in Lincoln County, W. Va., meet our criteria (fig. 25; table 9). Both measurements were obtained by hydraulic fracturing of wells at depths exceeding 800 m (about 2,600 ft).

The other six wells and well measurements shown in figure 25 and table 9 are from the analyses and compilation by Evans (1979). He examined oriented cores of parts or all of a sequence of gas-bearing Middle and Upper Devonian shales, taken in 13 gas wells drilled in Pennsylvania, Ohio, West Virginia, Kentucky, and Virginia. The cores are fractured, in many cases, intensively. They were collected to evaluate the fracture permeability of gas reservoirs in the shales, and it was crucial to determine which fractures are natural and which induced by the coring operation, including drilling. Evans examined the cores with techniques developed by ceramicists and later applied to rocks in work funded by the U.S. Department of Energy (Kulander, Dean, and Barton, 1977; Kulander, Barton, and Dean, 1979). Evans concluded that most unmineralized, unslickensided fractures in most cores are induced by coring. Of the three types of fractures induced by the coring operation, only that called petal-centerline is of interest here. Petal-centerline fractures are described, figured, and interpreted by Kulander, Dean, and Barton (1977), Kulander, Barton, and Dean (1979), Dean and Overbey (1980), and GangaRao and others (1979). Those authors concluded that the fractures form in advance of the downcutting drill bit, as extensional fractures in an orientation not distorted by the core or hole but determined by ambient stress at core depth. The result is one or several mostly vertical, planar fractures that parallel the axis of the core. The petal-centerline fractures form perpendicular to the least compressive stress, and so record the orientation of the least stress at all cored depths at which such fractures are observed. By inference, the vertical fractures also strike parallel to the trend of the greatest horizontal compressive stress.

GangaRao and others (1979, p. 686) reported that in some cores petal-centerline fractures dip steeply but not vertically. In such cases, the principal stresses would not have been vertical or horizontal. However, even in those cases, the orientation of the fractures would be interpreted as defining the orientation of the least compressive principal stress, and defining the orientation of a steeply dipping plane that contains the orientations of the greatest and intermediate compressive principal stresses.

We consider petal-centerline fractures to give accurate and precise estimates of the orientations of greatest and least horizontal compressive stresses for five reasons:

1. The oriented cores come from depths from 290 to 2,027 m (table 9), well below the near-surface zone of weathering and intensified jointing that may be responsible for the notorious complexity of individual stress determinations from shallow depths (for example, see Zoback and Zoback, 1980, p. 6128).
2. Cores can be hundreds of feet (meters) long, and so their fractures can average out variations in

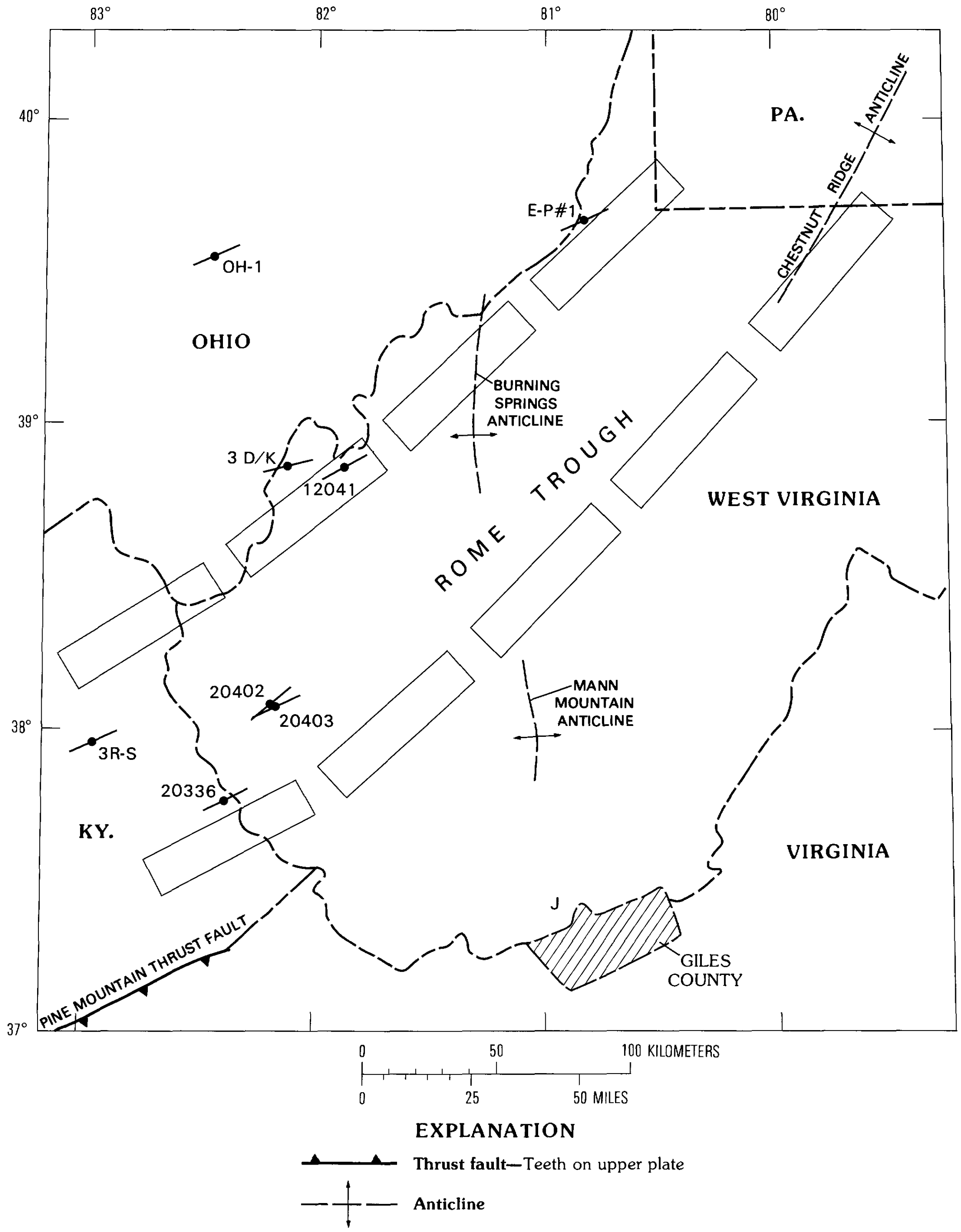


FIGURE 25.—Orientations of maximum horizontal compressive stress. Solid circles and lines through them show well locations and orientations of selected stress determinations from well cores (see text and table 9). Dashed lines show approximate locations of westernmost structures known to us to show significant thrust transport: Chestnut Ridge, Burning Springs, and Mann Mountain anticlines, and outcrop of Pine Mountain thrust fault. J shows approximate location of Elgood earthquake (table 7). Aligned open rectangles show approximate locations of southeast and northwest border faults of Rome trough: compiled from Ammerman and Keller (1979), Harris (1975, 1978), Kulander and Dean (1978b), and Shumaker (1977).

stress orientation between individual beds or groups of beds.

3. Such an oriented core can contain tens to many hundreds of individual petal-centerline fractures, many of which extend through several meters of core, so that orientations can be averaged over the entire cored interval.
4. Even cores with few petal-centerline fractures can

TABLE 9.—*Locations, sources, and values of selected stress orientations*

[Well code: usually permit number. Source, Zoback and Zoback (1980, Z in table) and Evans (1979, E in table). Code: parentheses after Z or E are well designators used by those authors. Method: HF, stress orientation obtained by hydrofracturing the well and then determining strike(s) of vertical cracks in hole wall. PCL, orientation determined by measuring strikes of vertical portions of petal-centerline fractures (defined in text) in an oriented core. n: Number of petal-centerline fractures used, leaders (----) indicate information not applicable or not given. Depth: Below ground level. References: Source of further information about the wells]

County and State	Well code	Source and code	Method	Stress orientation	Depth		n	References
					Meters	Feet		
Hocking, Ohio---	-----	Z(OH-1)	HF	N. 64° E.	808	2650	----	Haimson, 1974.
Martin, Ky.-----	20336	E(KY3)	PCL	N. 63° E.	758-1038	2486-3404	1573	Evans, 1979, p. 244-260; Wilson and others, 1980.
Johnson, Ky.----	3 R-S	E(KY4)	PCL	N. 65° E.	290- 457	950-1500	3	Evans, 1979, p. 261-276.
Wetzel, W. Va.--	E-P No. 1	E(WV7)	PCL	N. 67° E.	1859-2027	6100-6650	11	Evans, 1979, p. 161-176.
Mason, W. Va.---	3 D/K	E(WV5)	PCL	N. 75° E.	826-1042	2711-3420	1268	Evans, 1979, p. 125-146.
Jackson, W. Va.	12041	E(WV2)	PCL	N. 60° E.	981-1125	3220-3690	738	Evans, 1979, p. 67-84.
Lincoln, W. Va.	20402	Z(WV-1)	HF	N. 50° E.	835- 839	2738-2752	----	Evans, 1979, p. 106-124; Haimson, 1977; Abou-Sayed and others, 1978.
Do----------	20403	E(WV3)	PCL	N. 65° E.	829-1227	2720-4025	1215	Evans, 1979, p. 85-105.

give stress orientations that are consistent within the core and that match orientations from nearby wells (table 9). Indeed, in the cores described by Evans (1979), preferred orientations of petal-centerline fractures are exceptionally strong, with few fractures falling more than 10° away from the orientations listed in table 9.

5. If the core is preserved, it can be reexamined for stratigraphic and structural evidence with which to evaluate any changes along the core in the strike of the petal-centerline fractures. For example, we do not include stress orientations determined from petal-centerline fractures in a core from a well in southwestern Virginia (Wise County, well No. 20338). Evans (1979) and Wilson and others (1980) examined this core. They studied the vertical distributions and orientations of slickensides and slickenlines in the core, and structural, stratigraphic, and drilling information from the region surrounding the well. They concluded that the core bottomed in the sheared rock of the Pine Mountain thrust fault. The stress orientation of N. 57° E. from the core of well No. 20338 is consistent with other values obtained nearby (table 9; fig. 25), but to use such a result from rocks known to be thrust would violate one of the criteria of data selection that we shall discuss next.

CRITERIA USED TO SELECT DATA

Within about 300 mi (500 km) of Giles County, Zoback and Zoback (1980, plate 2; 1981, fig. 1) compiled 27 stress-orientation measurements made by various workers using various methods. Evans (1979) examined cores from 13 wells. These 40 measurements from Zoback and Zoback (1980, 1981) and Evans (1979) were reduced by us to the eight of table 9 by using the four following criteria:

1. We use only measurements likely to reflect stress orientations in the North American continental crust that is inferred to underlie Giles County at seismogenic depths, whether or not that crust has been reworked by Grenville metamorphism or fractured by Iapetan normal faults.
2. We use only stress orientations measured near Giles County, in the eastern parts of Tennessee, Ohio,

and Kentucky, in the western parts of Virginia and North Carolina, and anywhere in West Virginia.
3. We use only stress orientations likely to be representative of the stress field at seismogenic depths under Giles County. This means that measurements from rocks in thrust sheets are suspect because such rocks may be mechanically decoupled from underlying rocks.
4. We avoid stress orientations determined from measurements made at or within several tens of meters of the Earth's surface. Such measurements are notoriously variable and difficult to evaluate.

Criteria 1 and 3 require explanation. Criterion 1 restricts us to measurements made west of the steep rise in the unfiltered Bouguer anomaly field (fig. 18). East of that rise, we have suggested (p. 35) that the crust is an assemblage of pieces of various sizes, shapes, compositions, thicknesses, and origins. Thus, the stress field at seismogenic depths east of the rise may be more varied than, and differently oriented from, that under Giles County.

The stress data themselves contain support for this criterion. Zoback and Zoback (1980, 1981) divided the Eastern United States (apart from the Gulf Coast) into two stress provinces. In the region surrounding Giles County (fig. 19), the boundary between the two stress provinces coincides with the eastward rise in the Bouguer gravity field. To the east of the rise lies the Atlantic Coast stress province and to the west lies the Midcontinent stress province. The Atlantic Coast province is characterized by a variable but generally northwesterly trending greatest horizontal compressive stress, with the least compressive stress being vertical. From this, Zoback and Zoback inferred the existence of a coastal domain of reverse faulting involving compression at high angles to the Appalachians and continental margin. (Wentworth and Mergner-Keefer (1980, 1981a, b, c) arrived at the same conclusion.) Zoback and Zoback (1980, 1981) assigned the middle portion of the United States to the Midcontinent stress province, characterized by northeasterly trending greatest horizontal compressive stress. From this, they inferred the existence of a domain of reverse and strike-slip faulting. The eastward rise in the Bouguer anomaly field consistently separates data sites of the Midcontinent and Atlantic Coast stress provinces (also noted independently by Seeber and Armbruster, 1981). Thus, stress orientations measured east of the gravity rise are interpreted to come from a different stress province than that containing Giles County.

Criterion 3 restricts us to stress measurements made in rocks below the deepest thrust faults, or west of the westernmost rocks involved in thrusting, because thrust sheets might be mechanically decoupled from underlying rocks. Structural data consistent with such decoupling of Paleozoic stresses across thrust faults in the area shown in figure 25 are given by Wheeler (1980, p. 2173–2174), Werner (1980), and Wilson and others (1980). Seeber and Armbruster (1981) suggested similar decoupling for the modern stress field across a deeper thrust fault in Georgia. On the other hand, Zoback and Zoback (1980, p. 6136) pointed out that modern stress orientations in thrust rocks of the Appalachian Plateau province of western New York State and adjacent Pennsylvania are nearly parallel to those in nearby and underlying rocks not involved in thrust sheets. Thus, the question of decoupling remains open, and it seems possible that thrust rocks could be partly decoupled from underlying basement in some places and not in others.

In the region shown in figure 25, the deepest thrust faults cut upsection to the northwest. This is known from the dominant southeastward dips of outcropping thrust faults; from observing that more southeasterly thrusts expose older rocks that have been brought up from greater depths; and from abundant well and seismic-reflection data. For examples, see the cross sections of Rodgers (1963), Cardwell and others (1968), Gwinn (1970), Perry (1978), Roeder and others (1978), and Boyer and Elliott (1982, fig. 29). Because the deepest thrusts climb stratigraphically to the northwest, there is a western limit to rocks that have been involved in thrusting or anticlines produced by thrusting. Figure 25 locates westernmost structures known to us to have experienced important thrusting: Chestnut Ridge, Burning Springs, and Mann Mountain anticlines, and the outcrop of the Pine Mountain thrust fault. Small amounts of thrusting, or simple shear distributed over a stratigraphic interval without loss of cohesion, may occur west of those indicated structures (for example, see Shumaker, 1980). That situation may be most likely at shallow stratigraphic levels, such as within the Pennsylvanian and Permian rocks that are exposed around the wells shown in figure 25. Accordingly, we use only stress measurements made appreciably to the west of rocks likely to be thrust, or well below thrusted rocks, or both.

Other considerations apply to only one or a few cores, and cause us to discard data from those cores: Some cores examined by Evans (1979) contain no petal-centerline fractures, so stress orientation cannot be measured. Other cores have such fractures, but they exhibit no strong preferred orientation. Again, stress orientation cannot be measured on such cores. In examining one core, early workers did not distinguish natural from core-induced fractures. The core of well No. 20336 in Martin County, Ky., has 1573 petal-centerline fractures and preferred orientations of N. 33° E.,

N. 47° E., and N. 63° E. (Evans, 1979, p. 250). We use only the N. 63° E. orientation because that is most representative of fracture orientations in the bottom part of the core, below the effects of any near-surface thrusting in front of the outcrop of the Pine Mountain thrust fault.

Petal-centerline fractures in the core of well No. 20402 in Lincoln County, W. Va., present a problem. The 1,627 fractures give a strong preferred orientation of N. 33° E. (Evans, 1979, p. 114). That is anomalously more northerly than any of the other orientations shown in figure 25 or listed in table 9. However, a hydrofracturing experiment in the upper portion of the cored interval gives a stress orientation of N. 50° E. (Abou-Sayed and others, 1978), and the strike of petal-centerline fractures is N. 65° E. at well No. 20403, only 1 mi (1.6 km) away. Therefore, for well No. 20402 we use the hydrofracturing result (N. 50° E.) rather than that from the petal-centerline fractures (N. 33° E.), for the following reasons: First, the anomalously oriented fractures of well No. 20402 are not ubiquitous in the cored interval; the topmost 46 feet (14 m) of the 614 ft (187 m) cored contain petal-centerline fractures that strike N. 49° E. (Evans, 1979, p. 107, 116), in agreement with the hydrofracture result from that interval and with the orientation of petal-centerline fractures from the nearby well No. 20403 and the other wells of figure 25 and table 9. Second, both wells are within the Rome trough (fig. 25); as mentioned previously, some of the basement faults of that structurally complex graben were active intermittently throughout Paleozoic time. We speculate that the present-day stress field in the lower and larger part of the rock volume cored by well No. 20402 is distorted by past movement of some underlying fault associated with the Rome trough. We suggest that the top part of the cored volume of well No. 20402, and all the volume cored by well No. 20403, sample the regional and undistorted stress field.

This suggestion that stress fields have been distorted locally by past movement on basement faults, and have remained distorted, draws support from results of other workers. Such basement faults are known to be nearby and are inferred to have affected structures in the overlying sedimentary sequence. The Warfield fault and its associated Warfield anticline are about 15 mi (24 km) south of wells Nos. 20402 and 20403 (figs. 17, 25). At the Midway-Extra gas field, about 30 mi (48 km) northeast of wells Nos. 20402 and 20403 (figs. 17, 25), gas is produced from a fractured reservoir at the inflection line between an anticline and a syncline that are inferred to overlie another basement fault (Evans, 1979, p. 107; Schaefer, 1979; Cardwell, 1976). Further, distorted stress orientations in sedimentary rocks overlying basement faults of the Rome trough have been predicted from finite-element modelling, if the downdropped block is on the east side of the fault (Advani and others, 1977). Distortions are also inferred from anomalous orientations of cleat (planar fractures; systematic joints) in exposed coals, if the downdropped block is on the west side of the fault (Kulander, Dean, and Williams, 1977). Accordingly, we will use the hydrofracture result (N. 50° E.) instead of that from the petal-centerline fractures (N. 33° E.) for well No. 20402.

RESULTS

Orientations of greatest horizontal compressive stress that meet all the criteria just described are listed in table 9 and mapped in figure 25. They span about 170 mi (250 km) along the northeast trend of the Appalachians and about 110 mi (170 km) across the trend. They are in the western Appalachian Plateau province and adjacent foreland, from about 90 to 190 mi (150–300 km) west to north of Giles County.

Our selection criteria have produced a set of eight consistent estimates of stress orientations that cover a large area. The median trend is N. 64° E., with a range of 25° from N. 50° E. to N. 75° E. (fig. 26*A*). This median orientation agrees with the east-northeasterly orientations that Zoback and Zoback (1980) found for the eastern part of their Midcontinent stress province. Zoback and Zoback (1980, p. 6136) also suggested the existence of a transition zone about 200 km wide, comprising the eastern edge of the Midcontinent stress province and containing stress orientations that are roughly east-west, from Pennsylvania to Tennessee. We find no evidence of such a transition zone and suggest that the absence may be attributed to two factors. First, petal-centerline fractures can provide valid and accurate estimates of in situ stress orientation if measurements are made below the near-surface zone. However, some of the transition zone of Zoback and Zoback is based on measurements made at or near ground level. Second, perusal of the stress orientations in the hypothesized zone of transition indicates that most, but not all, are in thrust rocks. Thus, much of the transition zone may be caused by partial, local decoupling of rocks in thrust sheets from underlying rocks. The thrust rocks perhaps partly reflect stresses transmitted cratonward from the Atlantic Coast province. Both factors could produce a transition zone that reflects only near-surface stresses.

The result from the eight selected measurements can be improved slightly. We weight the measurements for geographic independence by averaging pairs of orientations determined in adjacent wells, which yield the results of figure 26*B*. For example, the two wells in Lincoln County, W. Va., are 1 mi (1.6 km) apart (Evans,

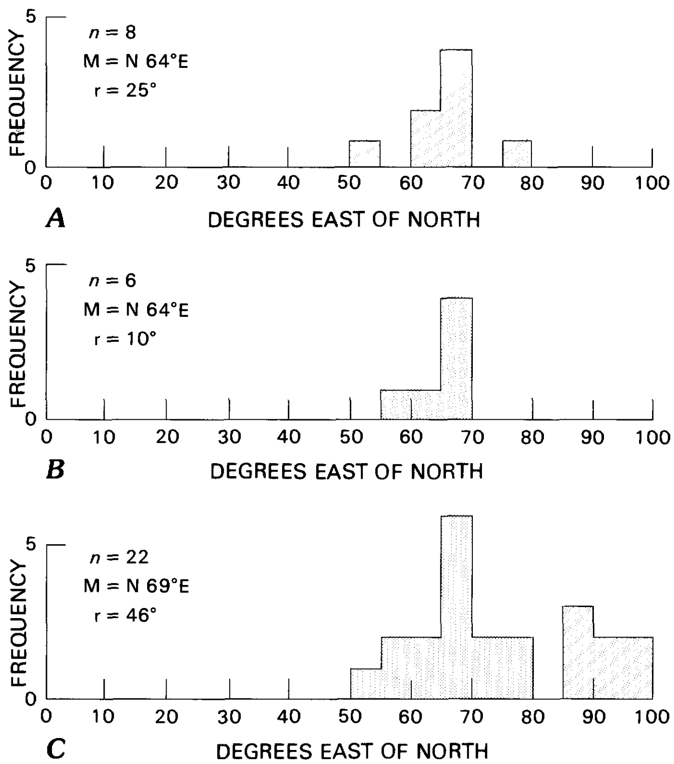


FIGURE 26.—Orientation distributions of measurements of greatest horizontal compressive stress. Class interval is 5 degrees, and n = number of measurements in a given set of orientations, M = median, r = range. *A*, The eight measurements that passed the selection criteria described in the text. *B*, The six measurements derived from those of *A* by averaging measurements from pairs of nearby wells. These six measurements are our preferred results. *C*, The 22 measurements obtained by adding to those of part *A*, 10 from Overbey (1976), 3 from Zoback and Zoback (1980), and 1 from Evans (1979).

1979, p. 92). The hydrofracturing determination from well No. 20402 averages with the determination from petal-centerline fractures in well No. 20403 to give a stress orientation of N. 58° E. The two wells in Mason and Jackson Counties, W. Va., are about 13 mi (21 km) apart (Evans, 1979, p. 72); their two determinations from petal-centerline fractures average to give N. 68° E. The resulting six estimates of the trend of greatest horizontal compressive stress still have a median of N. 64° E., but the range has decreased to 10°. The two extreme values are the two average orientations just described. This median and range are our preferred estimates for the stress orientation at and around the Giles County seismic zone (fig. 26*B*).

The selection criteria that we have used to produce table 9 and figure 25 have not changed the median stress orientation much, but the criteria have narrowed the range of individual site orientations considerably. Those results are shown by comparing medians and ranges between the three parts of figure 26. Figure 26*C* was obtained by including 14 other orientations that did not pass our selection criteria. In compiling figure 26*C*, we first include 10 of the orientations compiled by Overbey (1976). Those 10 orientations are mostly in the area shown in figure 25 and are apparently results of hydrofracturing at depth rather than near-surface strain-relief experiments. These 10 orientations are consistent with other nearby measurements and do not duplicate any individual results tabulated by Zoback and Zoback (1980) or Evans (1979). However, we have not determined whether any of those 10 measurements satisfy any of our criteria except the first, which is that they lie west of the gravity gradient.

The second group of orientations used in figure 26*C* are the three orientations compiled by Zoback and Zoback (1980, TN-3, OH-2, and WV-4) and one orientation by Evans (1979, VA-1 from well No. 20338 in Wise County, extreme southwestern Virginia). In compiling figure 26*A*, we originally deleted these four measurements because OH-2 is too far north, and the other three are known or suspected to have been measured in rocks that are shallow, thrusted, or both.

All 22 stress orientations together have a median of N. 69° E., only 5° more easterly than that of our preferred measurements (fig. 26*B*). However, the 22 measurements of figure 26*C* range over 46°, from N. 50° E. to N. 84° W. Thus, selection criteria that are designed with considerations of local and regional geology and structure, and which are combined with stress orientations measured by reliable methods can greatly improve the precision of estimates of stress over a large area. The accuracy of the estimate of stress orientation was also improved, but only slightly.

Another encouraging conclusion can be drawn from the consistency of stress orientations over the area shown in figure 25. This conclusion is that the Rome trough apparently has little effect on present-day stress orientations, at least over an area as large as that shown in figure 25. (Recall that we suggest a local effect for all but the top part of the interval cored by Lincoln County well No. 20402). This lack of effect on present-day stresses is noteworthy because some of the faults of the Rome trough were active from Cambrian through at least Pennsylvanian and perhaps Permian time, at least in eastern Kentucky (Cavorac and others, 1964; Black and Haney, 1975; Dever and others, 1977; Harris, 1978; Horne and others, 1978; Ammerman and Keller, 1979) and perhaps in central West Virginia (Kulander and Dean, 1978a; Kulander, Dean, and Williams, 1977, 1980). Of the eight well sites shown in figure 25, well OH-1 and probably wells E-P Nos. 1 and 3 D/K are northwest of the Rome trough. The other wells are within the limits of the trough. Apparently this major, long-lived crustal structure has not distorted the stress orientations of figure 25. The stress orientations determined by Evans (1979) from petal-centerline fractures indicate that the N. 64° E. orientation persists across the complex and long-active basement faults of the Rome trough.

The criteria that we used to select the eight measurements of figures 26*A* and 26*B* indicate that the N. 64° E. stress orientation represents the orientation of stresses at mid-crustal depths, under the thrust sheets of the Giles County locale. Such mid-crustal stresses would be those that cause the earthquakes of Giles County. The eight wells at which the stress measurements of figures 26*A* and 26*B* were made are separated from Giles County by some or all of the faults of the Rome trough, but we have just noted that the Rome trough does not appear to distort the regional stress field over the area of about 170 km by about 250 km that is covered by the wells plotted in figure 25. Thus, we shall extrapolate the N. 64° E. stress orientation 150 km to the southeast, across the Rome trough, to mid-crustal depths under Giles County. It now remains to test the N. 64° E. stress orientation for consistency with available seismological data from the Giles County seismic zone; we shall do that next.

CONSISTENCY WITH FOCAL MECHANISMS

Figure 27 illustrates an evaluation of the consistency of stress orientations deduced from the composite focal mechanism of figure 16, with those deduced from estimates of in situ stress that are shown in figure 25 and table 9. The in situ stress estimates are lines of zero plunge, parallel to strikes of vertical petal-centerline fractures. The vertical fractures formed by extension against the least compressive principal stress (S_3), which is also the least horizontal compressive principal stress (S_h) because the fractures are vertical. If we assume that the estimates can be extrapolated southeastward and downward to the Giles County seismic zone, the only constraint they provide is that S_h of figure 27 is parallel to S_3. The greatest and intermediate compressive principal stresses (S_1 and S_2, respectively) are constrained only to lie somewhere in the vertical plane perpendicular to S_h (represented by the dash-dot great circle of figure 27). Thus, S_1 and S_2 cannot be plotted directly in figure 27.

We concluded that the seismic zone probably occurs on an Iapetan normal fault that is being reactivated in today's stress field. For reactivation of an old fault that is weaker than surrounding rock, the angle between the reactivated fault and S_1 can depart from the ideal value of 30° that is typical of unfractured, homogeneous, brittle rock. Estimates of the size of the departure vary.

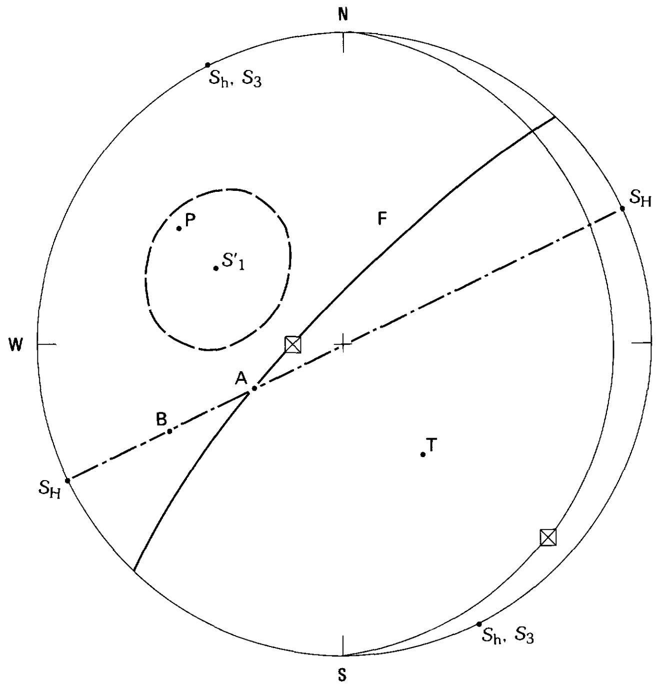


FIGURE 27.—Consistency of in situ stress orientation with orientation deduced from composite focal mechanism. Lower-hemisphere equal-area projection. Elements of focal mechanism of figure 16: solid curves show nodal planes, and boxed X's, their poles; F identifies preferred nodal plane, assumed to represent the orientation of the seismogenic fault or fault zone, which strikes N. 44° E., and dips 80° NW.; P and T locate compressional and tensional axes, respectively, at the seismic source. Elements of in situ stress field of figure 25 and table 9: S_H shows orientation of greatest horizontal compressive stress, a line that trends N. 64° E. and plunges 00° NE.; S_h shows orientation of least horizontal compressive stress, a line that trends N. 26° W. and plunges 00° NW.; dash-dot great circle shows plane perpendicular to S_h. Elements of greatest principal compressive stress, estimated from focal mechanism as recommended by Raleigh and others (1972): S_1' orients the stress; broken line is a small circle enclosing all orientations within 20° of S_1'; S_3 is least principal compressive stress. Points A and B are defined in text.

McKenzie (1969) noted that, in the most general case, S_1 determined from a focal mechanism is constrained only to lie within the compressional quadrant of the focal sphere. Thus, in principle, S_1 could lie as far as 90° from the P axis of figure 27. Raleigh and others (1972) suggested a procedure for estimating the probable orientation of S_1 from a focal mechanism. That estimate is S_1' of figure 27. It lies 15° from the P axis in the direction toward the preferred nodal plane. Raleigh and others suggested that in most cases the true orientation of S_1 should lie within 20° of S_1'. That range of orientations is enclosed by the broken line that defines the projection of a small circle in figure 27.

For strict consistency between the two orientations of S_1 deduced from the composite focal mechanism and from the in situ measurements, the dash-dot great circle of figure 27 should intersect the small circle around S_1'. Then, that intersection would define the possible range of orientations of S_1.

Thus, it appears that the stress orientations deduced from the in situ measurements are inconsistent with those deduced from the composite focal mechanism of figure 16. Further, a focal mechanism published by Herrmann (1979), for event J of our figure 11, appears to be inconsistent with both of our deduced stress orientations. Herrmann's solution will be discussed later. (Note: Event J occurred only about 25 km northwest of the Giles County seismic zone (fig. 11)). From these apparent inconsistencies, one could conclude that the stress state at seismic depths under and near Giles County changes markedly over horizontal distances as short as 25 km; such a conclusion is premature. In particular, the uncertainties in the sparse data allow the hypothesis that the in situ measurements, the first motions from the Giles County seismic zone (fig. 16), and at least some aspects of Herrmann's focal mechanism for event J might all reflect the same stress field, for the following reasons:

First, the 20° radius of the small circle is not an inflexible limit. Experimental data tabulated by Raleigh and others (1972) show considerable variation in the value of the angle between S_1 and the resulting reactivated fault. At least some of the variation can be attributed to differences in mean compressive stress, in the smoothness of the fault surface, and in lithology of the faulted rock. The dash-dot great circle of figure 27 lies only 13° from the small circle and consistency (point A is the point of closest approach).

Second, recall that the range of trends of S_H is 10°. Using the most easterly trending of the 6 orientations of figure 26*B*, point A would move another 2° closer to consistency.

Third, and most important, the discrepancy between the two estimates of the orientation of S_1 depends mostly on the orientation of the auxiliary nodal plane of figures 16 and 27. In particular, the orientation of the auxiliary nodal plane constrains the orientation of S_1'. That plane is constrained to include the pole of the fault nodal plane, but the dip of the auxiliary plane here is poorly constrained. For example, there are only six impulsive first motions from the seismic zone (fig. 16; table 8). If one uses only the five northeast- to southeast-plunging impulsive first motions, one could easily draw a steeper dipping auxiliary plane that would fit the five data about as well as does the shallowly dipping plane used here. If the auxiliary nodal plane were to steepen and strike more to the west in figure 27, S_1' and its enclosing small circle would be moved to the southwest to intersect the dash-dot great circle. If the auxiliary plane were to steepen from a dip of 14° to about 50°, the small circle would touch the great circle about at point B of figure 27, and the two stress estimates would be consistent within the limits suggested by Raleigh and others (1972). Then point B would represent the orientation of S_1.

Therefore, within the limits of our data, we can conclude that the in situ stress estimates are roughly consistent with the estimates deduced from the seismological data. S_1 probably plunges southwestward, toward points A and B of figure 27. Recall that the preferred nodal plane strikes northeast and dips steeply northwest (fig. 16). Motion on a fault with the orientation of the preferred nodal plane, and driven by compressive stress parallel to S_1, would be a combination of right slip and reverse slip. Because the orientation of the preferred nodal plane is determined by seismological data, Giles County seismicity is also produced by right-reverse slip. Such slip is consistent with the Midcontinent domain of reverse and strike-slip faulting suggested by Zoback and Zoback (1980). It is also consistent with most focal mechanisms compiled by them or given by Herrmann (1979) for unthrust rocks of the eastern craton of the United States. The relative importance of the reverse and right-slip components of motion on the Giles County seismic zone cannot be determined until the orientation of the auxiliary nodal plane is better constrained by more numerous impulsive *P*-wave first motions.

Herrmann (1979) used surface-wave data to derive a focal mechanism solution for the Elgood, W. Va., earthquake. That shock was located near but northwest of the Giles County seismic zone (fig. 25; Herrmann, 1979, event 11; our event J of fig. 11). His solution has a compression (P) axis trending northerly at low plunge and strike-slip motion on two steeply dipping nodal planes: left slip on a northeasterly striking plane and right slip on a northwesterly striking plane. Our selected stress orientations (fig. 25), extrapolated southeast to Elgood, are not consistent with the solution of Herrmann (1979, event 11). Our estimate of the orientation of greatest horizontal compressive stress (figs. 26, 27) falls near the dilatation (T) axis of Herrmann (1979). The most likely orientation of S_1 (fig. 27) also falls within the T field of Herrmann. However, the pattern of polarities of *P*-wave first motions in the northwest-to-northeast quadrant (dilatation) and the northeast-to-southeast quadrant (compression) are the same in both our and Herrmann's focal mechanism solutions. Applying the criteria of McKenzie (1969) and Raleigh and others (1972) to that similarity will allow a small area (about 1 percent of the focal hemisphere) wherein the S_1' about our P axis could include Herrmann's P axis. That is, Herrmann's P field includes roughly the northeastern quarter of the area in figure 27 that is enclosed by the small circle about S_1'. Thus, the markedly different focal mechanisms are not completely at odds with each other. However, resolution of the significant disparity that does exist lies beyond our scope here.

SUMMARY

The trend of greatest horizontal compressive stress has been measured at many places within several hundred kilometers north and west of the Giles County locale. The measured trends show much scatter, and were obtained in various geologic settings by methods of different reliability. We have formulated criteria based on considerations of local geology and structure, and on the reliability of the various methods that were used to obtain the stress measurements. Eight measurements satisfy these criteria. The eight measurements come from wells that were drilled 150–350 km north and west of Giles County, northwest of the thrust sheets that cover the Giles County locale and overlie its earthquakes. After averaging two pairs of duplicate measurements, the greatest horizontal compressive stress is estimated to trend N. 64° E., with little variability over the region covered by the wells. This orientation represents stress in the continental crust into which the wells were drilled and which extends southeast, under the thrust sheets of the Appalachians, to the location and depths of the seismicity beneath Giles County. The orientation of N. 64° E. is roughly consistent with stress orientations inferred from seismological data from the Giles County seismic zone, and it is partly consistent with stress orientations inferred from another earthquake that occurred near, but northwest of, the seismic zone. This consistency and the observed strike and dip of the seismic zone itself indicate that the seismicity of Giles County, including the damaging earthquake of 1897, probably occurs by right-reverse motion, in compression that trends about N. 64° E.

CONCLUSIONS

Our conclusions consist of specific statements, and of more general observations that should be borne in mind by users of the specific statements.

GENERAL OBSERVATIONS

This report presents the first direct instrumental evidence for a tabular seismic zone in Virginia, in the form of accurate and precise hypocenters that scatter but little about a plane. Additionally, by combining seismological and geological information, this report presents the first documentation of a seismically active basement fault or fault zone in the Southeast that does not parallel the trend of the known tectonic fabric of the host locale.

The excellent earthquake-location capability within the Giles County seismic network resulted in defining a seismic zone. The definition of the zone has two bases. First, the locale-specific velocity model (TPM2) has measured *P*- and *S*-wave velocities. Second, many of the microearthquakes are characterized by impulsive *P*- and *S*-wave phases (fig. 5), thereby allowing precise arrival-time determinations. Thus, accurate (*S–P*) time intervals from earthquakes within the network strongly constrain the hypocenter determinations in a manner that *P*-wave data alone cannot achieve. This situation is somewhat analogous to the independent determination of the origin-time procedure discussed by James and others (1969).

We have implicitly assumed throughout this report that the seismic zone we have defined is the same one that was the source of the 1897 shock. Clearly, the weight of evidence supports that assumption, but it cannot be proved. The intensity data are adequate to demonstrate that the greatest effects were, indeed, in Giles County (fig. 1; Bollinger and Hopper, 1971; Hopper and Bollinger, 1971; Law Engineering Testing Company, 1975). However, the presumption that Pearisburg was the probable epicentral locale is partly based on the fact that, as the county seat, it was the largest town in the county. Thus, the most numerous and detailed intensity reports came from there. Additionally, Campbell (1898), a U.S. Geological Survey geologist who visited the region in the early part of June, 1897, noted that: "The shock of May 31 was probably more severe in and about Pearisburg than any other point from which I have information."

There were two reasons for demonstrating the range of fault-plane areas that is allowed by the hypocenter data set to date (fig. 15). First, the demonstration conveys graphically the uncertainty of calculated fault-plane areas when a given level of statistical confidence is used as an error measure for the hypocentral locations. Second, in this case the demonstration shows that there can be a variation of a factor of ten in the implied fault plane area. Such uncertainty in the fault-plane area carries the potential for a change of one full unit in an associated earthquake's magnitude (Wyss, 1979; Singh and others, 1980). Realization of that potential would require that (1) the collection of individual hypocenters actually represents a single fault plane or zone, and (2) the entire plane or zone slips seismically all at once. However, we do know that in 1897 the locale experienced a shock roughly comparable in size to that associated with the minimum hypocentral area of 80 km^2 (m_b = 5.8, M_S = 6; Geller, 1976; O. W. Nuttli, written commun., 1980).

SPECIFIC STATEMENTS

The data presented and analyzed herein constitute a detailed instrumental and geological description of an individual seismic zone in the Southeastern United States. In our judgment, the evidence presented warrants the following conclusions:

1. A seismic zone has been defined in Giles County, Va., with the following seismological characteristics:
   - A. Strike—northeast; present data indicate N. 44° E.; Dip—near vertical; Depth range—from 5 to 25 km;
   - B. Horizontal length—40 km; centered at Pearisburg, Va.; Horizontal width—10 km.
2. The seismic zone also has the following geological characteristics:
   - A. Located within the basement and beneath the Appalachian thrusts;
   - B. Though the zone is in the southern Appalachians, it is subparallel in strike to the surface and near-surface structures of the central Appalachians to the north, and is at an angle of some 30° to the thrust-faulted tectonic fabric of the southern Appalachian host region.
3. Although conclusive evidence is lacking for the following aspects of the zone, we favor their likelihood:
   - A. The present-day motion on the inferred northeast-striking fault or fault zone is such that the southeast side is moving down relative to the northwest side;
   - B. High-angle reverse motion of the fault is more likely than is normal motion. At this time, it is impossible to determine which motion has occurred; nevertheless, high-angle reverse motion is the more likely because the seismic zone probably dips steeply northwest and because the region is probably under east-northeasterly trending compression at seismogenic depths;
   - C. Any strike-slip component of the motion is probably right-slip, though of unknown magnitude relative to the dip-slip component;
   - D. The zone defined in this report is the source of the 1897 shock. This implies an apparent resumption of strain energy release after a seismic quiescence of 4 to 5 decades;
   - E. The N. 44° E. seismic zone has probably resulted from compressional reactivation of a Late Proterozoic or early Paleozoic Iapetan normal fault or fault zone. Fault reactivation by late Paleozoic compression and Mesozoic extension is also possible.
4. Although flat or low-dip detachment faults have been found to or suggested to produce large earthquakes elsewhere, that is apparently not true for the Giles County seismic zone. Neither is it likely to be true for other earthquakes with well-determined depths in or near Giles County.

FUTURE WORK NEEDED FOR HAZARD ZONING

Our familiarity with the seismicity and geology of Giles County and the surrounding region enables us to offer suggestions that may be useful in guiding future workers who build on the conclusions of this report. The following suggestions in no way indicate that our findings here are preliminary, or that we expect future work to change these findings. It is unlikely that future seismicity in the Giles County locale will differ much from the seismicity analyzed in this report. The location of Giles County hypocenters below the Appalachian thrust sheets, and the lack of known surface ruptures associated with the 1897 shock, make it equally unlikely that important new geological or other geophysical data can be obtained quickly or at low cost. Therefore, we expect that future results will amplify and extend the conclusions of this report—but will not change them. Accordingly, the following suggestions are based partly on these conclusions, and partly on our more general experience with the geology, seismology, and hazard evaluation of the Southeast.

Our findings lead to three questions that must be answered to contribute to improvement of the existing hazard evaluations. At present, we know of no way to answer these questions quickly, but the following paragraphs suggest avenues of investigation that may eventually produce reliable answers:

1. Is a single fault or a single fault zone responsible for the Giles County seismic zone? Our only evidence for the existence of the seismic zone itself is the distribution of hypocenters shown in figures 11–14. From that distribution, one may infer the existence of a single fault or single fault zone. For example, figure 15 and the arguments based on it stem from such an inference.

Such an inference would be strengthened if the existence, orientation, and slip of one or more faults could be inferred from independent geophysical data, especially reflection seismic profiles. To test the existence of a fault or faults responsible for seismicity in Giles County, a reflection field experiment must be carefully designed to fit the reflector depths and geometries and fault offsets that are expected. Otherwise, equipment or processing may be selected that cannot resolve any fault offset that is present. For example, three-dimensional shooting geometries may be required to detect faults with very small offsets (J. K. Costain, oral communs., 1981).

Actual documentation of any outcropping fault or fault zone may be obtainable only through structural and other geologic mapping at scales more detailed than most hitherto done in the Giles County locale. Such mapping could seek and document small, systematic offsets of sharp contacts and structural elements or locate zones of unusually high intensity of joints and other fractures (Wheeler and Dixon, 1980).

Identification of a fault or fault zone responsible for Giles County seismicity is complicated by lack of any known rupture of the ground surface from the 1897 shock—or indeed from any accumulated motion on the seismic zone over time. However, we know of only one detailed search for such rupture and that is still in progress (McDowell, 1982, and his six preceding semiannual reports in the same series). Such a search is hindered by the comparatively moist climate, thick vegetation, and rapid erosion characteristic of the region and by the consequent sparseness of young and dateable geological materials that could record such rupture (Houser, 1981). Acharya (1980a, b) suggested that large earthquakes in eastern North America that do not rupture the ground surface must occur deeper than about 10 km. Such a depth would be consistent with instrumentally determined depths of microseismicity on the Giles County seismic zone (5–25 km) and would be consistent with the best estimate of the depth of the nearby Elgood earthquake of 1969 (table 1; Carts, 1981, depth 13.6 km). So, lack of known surface rupture from the 1897 shock is bothersome but not necessarily surprising.

A greater potential problem is the lack of any known surface offset that could be attributed to slip accumulated by the seismic zone by repeated activity over millions of years. However, the problem is still only a potential one because the problem is so poorly defined. For example, the seismic basement is overlain by several kilometers of complexly layered and faulted sedimentary rocks, and several thick shale sequences of largely unknown mechanical properties are contained within those rocks. It is not clear to us how fault slip would be transmitted through or dissipated within such a complex. Alternatively, the Giles County seismic zone might be only intermittently active. Such intermittent activity could arise from regular reactivation at long intervals. Such intermittent activity could also arise if the zone occurs on a fault that is but one element of a network of mechanically linked faults, which could relieve stresses imposed on the boundary of the network by concentrating them in turn at constantly changing points within the network. If the recoverable vertical strains within the crustal blocks of such a network were of about the same size as dip slips on the faults that bounded the blocks, and if the shifting stress concentrations within the network allowed such strains to alternately accumulate and relax, then the faults might experience alternating normal and reverse slip. Little or no net slip would accumulate, so little or no net slip would be visible at the surface. Therefore, the lack of known surface offset in Giles County is a complex enough problem that its further consideration lies beyond our scope here.

2. Are there other seismic zones structurally analogous to that in Giles County that lie along strike to the northeast or southwest? An eventual answer to this question will take one of two forms. One answer is that the Giles County seismic zone is unique in eastern North America. However, in addition to suggesting uniqueness, one should be able to explain it. For example, one might be able to show the presence of a northerly to westerly trending cross structure, under the thrust sheets, which might act to concentrate seismic release of stress on the Giles County seismic zone. Such cross structures could be of several kinds. For instance, the gradient in the unfiltered Bouguer field has a sharp S-shaped bend southeast of Giles County (figs. 18, 21). That bend may express the presence of an Iapetan transform fault. From analyses of gravity and aeromagnetic data, Phillips and Daniels (1982) suggested a marked change in the type of subthrust rock across that possible transform fault. Alternatively, Wheeler (1980) and Wheeler and others (1979) described a class of complex structures called cross-strike structural discontinuities (CSD's). Some CSD's apparently overlie basement faults of unknown or multiple ages and origins in structural settings similar to that of

Giles County. To our knowledge, CSD's have not yet been sought in Giles or most of its adjacent counties.

The alternative answer is that the Giles County seismic zone is not unique but is a presently active member of a class of similar zones that cover part of the terrane west of the gravity gradient. Before recommending this alternative answer, one should be able to suggest where other such zones might occur. This will be difficult. The Southeast is sparsely enough covered by seismograph networks that, over most of the region, such zones might not be recognizable. For example, of the 12 events that define the Giles County seismic zone, only the four that were relocated by J. W. Dewey and D. W. Gordon (written commun., 1980) exceed $M=2$ (fig. 12), and reliable location of smaller events is feasible only over a small area (fig. 6).

3. How far west or northwest of the rise in the unfiltered Bouguer gravity field may one expect to find Iapetan normal faults? An answer to this question is likely to be only approximate and might be expressed as the probability of finding such faults at specified distances from the rise. Such faults might not have a sharp cratonward limit but instead might decrease gradually in slip and in abundance away from the inferred Iapetan continental edge.

   Estimates of the spatial distribution to be expected of Iapetan normal faults may be obtained from modern Atlantic-type continental margins. A bound on such an estimate may be derived from the distribution of Mesozoic normal faults in eastern North America. That bound could be conservative from the viewpoint of hazard zoning by erring on the side of safety, overestimating the area underlain by Iapetan normal faults. If the crust east of the gravity rise is weaker than that to the west (as we have suggested), then normal faults might have formed farther inland from the Atlantic continental edge than they did from the Iapetan edge. Thus, an estimate derived from the Mesozoic faults might overestimate sizes, abundances, and area of occurrence of Iapetan faults. On the other hand, a nonconservative bound may be obtained from other modern margins, on which normal faults are commonly buried under younger sedimentary rocks and sediments. That estimate might be nonconservative because the more cratonward faults on such margins might be too small, too few, or both, to be resolved readily by standard geologic and geophysical techniques. Thus, both the numbers and cratonward extent of such faults could be underestimated. The two estimates might provide useful bounds for an estimate of the cratonward extent of Iapetan normal faults.

   A test of such estimates may be possible soon. Davies and others (1982) made numerous partly balanced cross sections across the Valley and Ridge and Appalachian Plateau provinces of the southern Appalachians. The sections were drawn to show numerous basement faults under the thrust sheets. By the arguments of this report, those basement faults are probably Iapetan normal faults.

Questions 2 and 3 posed in the preceding paragraphs deal with the uniqueness of Giles County seismicity in the Southeast. One hypothesis that bears on both questions is that of gravitationally induced stresses, which might reactivate Iapetan normal faults in the area that is outlined by the long Bouguer gravity low that flanks the steep eastward gravity rise on the northwest (Woollard and Joesting, 1964; Haworth and others, 1980; this report, fig. 18). The Giles County locale is in that long gravity low, and the rise passes 50–100 km southeast of the locale (fig. 18). Gibb and Thomas (1976) developed a composite model of crustal density distribution to fit Bouguer gravity profiles across four boundaries between Precambrian structural provinces in the Canadian Shield; Goodacre and Hasegawa (1980) used finite-element calculations based on that model to estimate shear stresses in the crust. Goodacre and Hasegawa applied their results to the Bouguer gravity rise where it passes through southeastern Quebec. They observed that seismicity in southeastern Quebec is concentrated in free-air gravity lows adjacent to free-air highs, where their calculations predicted that gravitationally induced shear stresses would be greatest. They hypothesized that the induced stresses have reactivated preexisting faults.

The hypothesis of Goodacre and Hasegawa (1980) is a possible explanation for the seismicity in the Giles County locale. Some of the largest, steepest parts of the Bouguer rise in the central and southern Appalachians are near Giles County. Also, the Giles County locale is in or near the long Bouguer gravity low, an area where the Goodacre-Hasegawa hypothesis predicts the greatest shear stresses at the depths of Giles County seismicity.

However, the hypothesis needs more detailed testing before being accepted for the Giles County locale for two reasons. First, there are several exceptionally steep parts of the rise and unusually strong positive and negative anomalies atop and at the bottom of the rise, between northern Virginia and northwestern South Carolina (Haworth and others, 1980). Indeed, Giles County itself is in a saddle between the two strongest negative anomalies, instead of within a negative anomaly as would be expected from a direct application of the results of Goodacre and Hasegawa (1980). Therefore, the hypothesis of gravitationally induced stresses requires further development to answer the question: Why is seismicity concentrated in and near Giles County, rather than at one of those other locales? Second, the models of Goodacre and Hasegawa and of Gibb and Thomas both attribute the induced stresses to lateral density contrasts that persist down to the base of the crust. From Pennsylvania southward, much of the size and steepness of the gravity rise is caused by the long gravity low adjacent to the rise on the northwest (Haworth and others, 1980; this report, fig. 18). That low lies about along the structural axis of the Appalachian basin, in which the sedimentary rocks are thickest, and the contours of the low approximately follow the mapped shape of the basin. How much of the rise is attributable to the sedimentary fill of the basin, rather than to density contrasts at the depths of Giles County seismicity? If the gravitational effect of the sedimentary rocks were removed by appropriate modeling, would finite-element calculations similar to those performed by Goodacre and Hasegawa predict large gravitationally induced shear stresses at the positions and depths of Giles County seismicity? Could induced stresses be further concentrated by cross structures similar to those mentioned in this section?

We suggest that the preceding questions and concepts be considered in designing future work on or near the Giles County seismic zone. The questions and their eventual answers will be important in drawing source zones for estimating seismic hazard, as well as in understanding the structural evolution of large parts of the North American continental crust. Currently, we have too few pertinent data to justify attempting to answer any of the questions. However, we know of no reason why carefully designed investigations should not eventually produce usably reliable answers to all of the questions we have posed.

ACKNOWLEDGMENTS

This research was sponsored in part by the National Science Foundation under Grants DES-75-1469-A1 and EAR-7923761 and in part by the Office of Nuclear Regulatory Research under Contract No. NRC-04-77-134, Mod. 5 and 6, all awarded to Bollinger.

The manuscript as a whole was improved significantly by suggestions of J. W. Dewey and C. M. Wentworth. Portions benefited from suggestions of S. R. Brockman, A. C. Donaldson, K. B. Krystinik, L. T. Long, W. D. Lowry, E. S. Robinson, R. C. Shumaker, Leonardo Seeber, R. W. Simpson, P. C. Thenhaus, and J. E. Tillman. We have been privileged to share the ideas, unpublished data, and preprints of many workers, including D. S. Albaugh, K. F. Bick, W. G. Brown, J. K. Costain, S. L. Dean, J. W. Dewey, W. H. Diment, A. C. Donaldson, D. W. Gordon, R. M. Hamilton, Y. W. Isachsen, M. F. Kane, B. R. Kulander, R. F. Lamb, P. A. Lentz, R. C. McDowell, O. W. Nuttli, W. J. Perry, Jr., H. W. Rauch, M. K.-Seguin, R. C. Shumaker, R. W. Simpson, A. C. Tarr, Rob Van der Voo, and C. M. Wentworth. Any remaining errors of science, fantasy, or propriety are ours alone.

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APPENDIXES

APPENDIX A

A NOTE ON MICROSEISMIC LEVELS FOR THE VIRGINIA POLYTECHNIC INSTITUTE SEISMIC NETWORK

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ABSTRACT

Six hundred amplitude and period measurements were made of the seismic waves comprising the short-period microseismic background at the World Wide Standard Seismograph Network (WWSSN) station in Blacksburg (BLA) and at eight Virginia network stations. The overall average amplitude level at BLA was 3 nm (daytime) and 5 nm (nighttime) at frequencies of 0.9-3.1 Hz. At the network sites, the average daytime amplitude level was 5 nm at 2.3 Hz; during the nighttime, it was 10 nm at 2.3 Hz.

INTRODUCTION

Noise surveys are usually employed to select sites for seismograph stations. However, follow-up measurements after a station or a network is installed and operational, are seldom made. There is normally little need for such measurements. However, if detection thresholds and network capability studies are to be made, knowledge of the ambient microseism levels is required. Additionally, specification of such levels can be useful for selection of additional sites in the region, and for engineering purposes related to radio telescopes, stable platforms, and other structures.

The Virginia Polytechnic Institute Seismic Network is perhaps representative of one class of network: short-period vertical transducers, with recording passband approximately 1–10 Hz. Stations are sited in four of the five major geologic provinces present in the Southeastern United States: Coastal Plain, Piedmont, Valley and Ridge, and Allegheny Plateau. Thus, noise measurements from the network could be used as approximations for expectable levels throughout the region.

PROCEDURE

A spectral analysis would be the optimum manner to specify microseismic levels. However, for many purposes, simple amplitude-period measurements are entirely adequate. Such a procedure was utilized for this study. A total of 600 such measurements were made from ten different station sites. Film seismograms, using a viewer (1 s of time=10 mm on the viewer screen) were employed for all measurements. These measurements were made according to the following scheme:

1. Choose the months of January, March, June, September, and December, 1979, as representative of seasonal variations.
2. For each month, select a "typical" day and for each day select typical 2-hour periods (for example, 07^h–09^h UTC, 2–4 a.m. EST; and 19^h–21^h UTC, 2–4 p.m. EST). Within those periods, select typical but arbitrary 2-minute periods.

For each 2-minute period, make measurements for the noisiest and quietest stations for the Giles County subnetwork and the central Virginia subnetwork. Also make measurements for WWSSN BLA. This procedure yielded 600 amplitude-period measurements at nine different stations. The average values at each of the stations are presented in table 10. Values missing in that table occur when a given station is neither noisiest nor quietest during a given month or during the day-to-night time frame.

ACKNOWLEDGMENTS

This study is based upon work sponsored by the National Science Foundation under Grant EAR-75-14691-A01 and by the Nuclear Regulatory Commission under Contract NRC-04-77-134. Thanks are due to Miss Christine Howell for assistance in the statistical analysis of the data for this report.

TABLE 10.—*Average microseismic amplitude levels and frequencies*

[Dash (-) indicates no record, see text explanation. Station abbreviations are HWV, Hinton, W. Va.; PWV, Princeton, W. Va.; NAV, Narrows, Va.; PUV, Pulaski, Va.; BLA, Blacksburg, Va.; CVL, Charlottesville, Va.; GHV, Goochland, Va.; FRV, Farmville, Va.; PBV, Petersburg, Va.]

Province	Station	January Day/Night	March Day/Night	June Day/Night	September Day/Night	December Day/Night
		Amplitude level (nanometers)				
Plateau---------	HWV	-/-	-/-	-/-	3/-	18/-
Do----------	PWV	11/15	3/3	2/-	3/3	4/5
Valley and Ridge	NAV	-/-	-/-	-/1	-/-	-/-
Do----------	PUV	23/60	2/3	2/1	-/9	-/32
Do----------	BLA	5/8	2/2	2/1	2/2	3/12
Piedmont--------	CVL	11/-	3/2	-/1	2/-	11/8
Do----------	GHV	-/12	-/-	-/-	-/-	-/-
Do----------	FRV	13/50	-/-	1/-	-/1	-/-
Coastal Plain---	PBV	-/-	2/2	2/2	1/1	3/4
Average-----------		13/29	2/2	2/1	2/3	8/12
		Microseismic frequency (Hertz)				
Plateau---------	HWV	-/-	-/-	-/-	2.3/ -	0.8/ -
Do----------	PWV	1.1/1.0	2.9/3.1	3.1/ -	3.0/2.9	3.1/1.7
Valley and Ridge	NAV	-/-	-/-	- /3.2	-/-	-/-
Do----------	PUV	0.8/0.6	2.8/1.4	2.5/3.4	- /0.9	- /0.7
Do----------	BLA	1.5/1.3	2.4/3.1	2.7/2.9	2.4/2.4	2.0/0.9
Piedmont--------	CVL	1.0/ -	2.7/3.4	- /3.0	2.7/ -	1.0/1.1
Do----------	GHV	- /1.0	-/-	-/-	-/-	-/-
Do----------	FRV	1.1/0.7	-/-	2.9/ -	- /3.1	-/-
Coastal Plain---	PBV	-/-	4.5/6.7	3.1/2.6	2.9/4.0	2.4/1.9
Average-----------		1.1/0.9	3.1/3.5	2.9/3.0	2.7/2.7	1.9/1.3

APPENDIX B

DETERMINATION OF A DURATION MAGNITUDE RELATIONSHIP FOR THE VIRGINIA POLYTECHNIC INSTITUTE SEISMIC NETWORK

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(Modified from Viret (1980))

INTRODUCTION

For the Virginia Polytechnic Institute Seismic Network, magnitudes of local and regional earthquakes are calculated using body-wave magnitude equations according to Nuttli (1973) and Bollinger (1979):

$$m_b(L_g) = 1.90 + 0.90 \log \Delta + \log (A/T) \quad 50 \text{ km} \leq \Delta \leq 400 \text{ km} \qquad (1a)$$

$$m_b(L_g) = -0.10 + 1.66 \log \Delta + \log (A/T) \quad 400 \text{ km} \leq \Delta \leq 2000 \text{ km}, \qquad (1b)$$

where Δ is the epicentral distance in kilometers, A is the sustained maximum ground motion, from center to peak, in microns, and T is the corresponding period in seconds.

Equation 1a does not apply for distances less than 50 km. In that distance range, Richter's local magnitude equation,

$$M_L = \log A - \log A_o + \log (G(WA)/G(Net)) \qquad (2)$$

is used (Richter, 1958). The log (G(WA)/G(Net)) is an adjustment term to allow for differences in magnification: G(WA) is the magnification of the Wood-Anderson seismograph (2800) and G(NET) is the Virginia Polytechnic Institute Network station magnification. A is the trace amplitude (half of the maximum peak to peak amplitude in mm), and A_o is Richter's standard earthquake amplitude (dependent on distance). The quantity ($-\log A_o$) is tabulated by Richter (1958, p. 342).

There are several sources of possible error in various schemes of magnitude determinations. That is, application of the available formulas in an uncritical manner can result in large errors. Possibly the most significant of the error sources are the following:

1. *Near distances.*—At epicentral distances less than 50 km, the use of M_L here includes no adjustment for differences in seismograph response between the mechanical-optical Wood-Anderson system (involved in the definition of M_L) and the electromagnetic seismographs used by the network. Additionally, there is no adjustment for the differences in seismic-wave attenuation between California and Virginia. However, with the small distances involved, the attenuation factor probably does not cause too large a disparity.
2. *Wave Frequency.*—The $m_b(L_g)$ formula is based on waves whose periods are within 0.2 s of 1 s. Any observed waves whose periods depart from that range carry the potential for large error. Also, the network seismograph's passband (fig. 4) has a much greater emphasis of the higher Earth frequencies than does the Wood-Anderson seismograph (Anderson and Wood, 1925). This emphasis could result in the use of a different seismic phase, or a different portion of the same phase, for magnitude determination than would have been considered had a Wood-Anderson seismograph been used.
3. *Different Interpreters.*—When the maximum vibrational amplitudes exceed the recording range of the instrument, none of the aforementioned magnitude equations can be used. It then becomes necessary to use a magnitude relationship based on the

duration of vibration. There is considerable subjectivity involved in the estimation of the duration of vibrations on a seismogram. Whitehurst (1977) estimated that a variation of only about 0.1 magnitude unit up or down is attributable to this factor. There were at least three different interpreters involved in the collection of the data set being considered here. In principle, then, we have the potential for 0.2 magnitude units variation from this source.

Several investigators have found empirically that a linear relationship between magnitude and the logarithm of duration of vibrations was adequate to specify earthquake size at near distances. As epicentral distances increase, however, a distance term must be added to this relationship. Because of the nature of seismic coda waves[1] as backscattering waves from numerous, randomly distributed heterogeneities in the Earth (Aki, 1969; Aki and Chouet, 1975), a theoretical basis for the empirical linear relationship can be described (for a review, see Whitehurst, 1977, p. 9–16). Thus, using equations 2, 1b, and 1a to calculate amplitude magnitudes for local and regional earthquakes, a relationship between the duration of vibrations and the magnitude of the causal earthquake can be established over a rather wide range of seismic energy release.

PROCEDURE

The magnitude-duration relationship is that of a straight line:

$$M_D = A + B \log (D) \qquad (3)$$

where M_D is the average network duration magnitude, D is the average duration of vibrations (usually in seconds) for the event, and A and B are constants to be determined. How duration is defined can affect the magnitude determined from the relationship of equation 3. Some authors define the duration as the time interval from the onset of the P-wave until the time when the earthquake vibrations return to the ambient microseismic noise level. That definition was used in this study. Another definition uses the same beginning but fixes the end of duration at the time the trace amplitude returns to a predetermined arbitrary peak-to-peak amplitude (Whitehurst, 1977).

[1]Seismic coda waves are the "tail" or final portions of a seismogram of a local earthquake; they are that part on a seismogram after the arrival of major wave types such as P, S, and surface waves (Aki and Chouet, 1975; Whitehurst, 1977).

For this study, there were three sources of data. One source was a data set compiled from durations and magnitudes (M_L or $m_b(L_g)$) measured on the Virginia Polytechnic Institute Seismic Network. The other two sources used durations and magnitudes (M_L or $m_b(L_g)$) measured on the Phase-I (P1) and Phase-II (P2) networks used for seismic monitoring at the North Anna site (Dames and Moore, 1976). The instruments used at that site were similar, and in some cases identical, to those now in use at Virginia Polytechnic Institute. Thus, given the same host region and the same general class of instrumentation, the data sets should be from the same general population. In all cases, the durations and magnitudes for a single event are averaged at all network stations to produce a network average.

The Virginia Polytechnic Institute Network magnitude data were combined with the North Anna magnitudes (VPI+P2+P1) to produce the input data set. A least-squares best-fit line was first determined for all the data points and then every point more than one standard deviation from that line was arbitrarily deleted to reduce excessive scatter. Finally, a new line was fit to the remaining points. The result of the above procedure is the equation:

$$M_D = (-3.38 \pm 0.09) + (2.74 \pm 0.06) \log (D). \qquad (4)$$
$$n = 102$$
$$(SD) = 0.25$$

where n is the final number of points used to calculate the equation of the line and (SD) is the standard deviation of the points about that line. The plus-minus values refer to the standard deviations of the estimates of the slope and the intercept. See table 11 for a listing of these 102 input data pairs.

SUMMARY

We chose as a provisional duration magnitude relation the following equation derived from 102 data points:

$$M_D = -3.38 + 2.74 \log (D) \qquad (5)$$

Figure 28 shows a plot of this curve. It is interesting to note that equation 5 gives values similar to those derived from the WWSSN station BLA's equation: $M_D = -2.87 + 2.44 \log (D)$ as determined by Whitehurst (1977). Table 12 presents a list of the recalculated magnitudes for the Virginia microearthquakes located to date.

TABLE 11.—*Data set used in the determination of average network duration magnitude* (M_D) *when* M_D = *-3.38+2.74 log* D *and* n=*102*

[Duration D is in seconds. Leaders (----) indicate no data available]

D	M_L	m_b	D	M_L	m_b	D	M_L	m_b
15	0.3	---	16	-.2	---	188	3.3	---
16	.5	---	16	-.3	---	209	3.2	---
21	.9	---	16	.1	---	232	3.5	---
30	.8	---	17	-.2	---	361	4.0	---
30	.9	---	18	.3	---	361	3.7	---
30	1.2	---	19	.0	---	8	-.4	---
40	1.1	---	19	.2	---	8	-.5	---
44	---	1.2	20	.5	---	10	-.6	---
120	---	2.5	21	.2	---	10	-.6	---
120	---	2.6	24	.4	---	10	-.3	---
15	.1	---	24	.1	---	11	-.3	---
16	.0	---	25	.3	---	11	-.7	---
16	.1	---	26	.4	---	11	-.6	---
20	.3	---	26	.5	---	30	.3	---
20	.4	---	26	.1	---	31	.5	---
21	.2	---	26	.2	---	32	.4	---
22	.1	---	26	.2	---	35	.6	---
23	.2	---	26	.3	---	36	.9	---
23	.5	---	27	.4	---	37	.8	---
23	.6	---	29	.5	---	38	.8	---
23	.6	---	34	.7	---	42	.8	---
25	.4	---	36	.7	---	44	.9	---
26	.7	---	37	.7	---	51	.9	---
28	.5	---	39	.8	---	86	1.6	---
29	.6	---	39	.7	---	117	2.7	---
29	.5	---	42	.8	---	137	2.4	---
30	.7	---	51	.9	---	160	2.2	---
12	-.4	---	52	1.3	---	179	3.0	---
12	-.7	---	53	1.4	---	180	2.9	---
13	-.5	---	62	1.7	---	232	2.8	---
13	-.3	---	64	1.2	---	287	3.4	---
13	-.3	---	100	2.3	---	294	3.1	---
14	-.4	---	159	3.1	---	392	3.7	---
15	.2	---	174	2.8	---	545	4.3	---

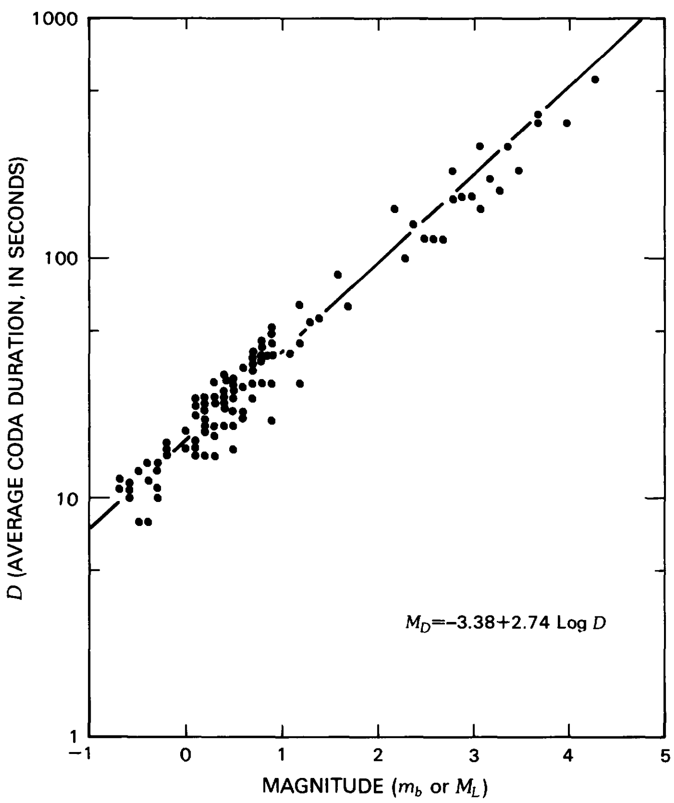

FIGURE 28.—Plot of average coda duration versus magnitude for earthquakes recorded by the Virginia Polytechnic Institute Seismic Network.

TABLE 12.—*Recalculated average network duration magnitudes using* M_D $=-3.38 + 2.74 \log D$

[Duration D is in seconds; leaders (---) indicate no data available]

Map No.[1]	Date Year	Month	Day	Time (UTC)	Average duration (D)	M_D	$m_b(L_g)$[2]	M_L[2]
21	1976	Sept.	13	18:54	193	2.9	3.1	---
32	1978	Jan.	28	23:13	66	1.6	(2.9)	(2.4)
32A	1978	Mar.	17	18:26	152	2.6	2.8	---
33	1978	May	10	4:19	22	.3	---	.6
34	1978	May	25	8:30	59	1.5	---	1.0
35	1978	June	1	1:33	14	-0.2	---	(.5)
36	1978	June	22	6:42	43	1.1	(2.27)	1.5
37	1978	July	28	8:39	28	.6	---	.6
38	1978	Aug.	30	2:19	25	.5	---	.7
39	1978	Sept.	14	19:37	12	-0.4	---	(.7)
40	1978	Oct.	14	1:50	22	.3	---	(1.0)
41	1978	Oct.	29	12:22	44	1.1	---	1.6
42	1978	Nov.	15	8:33	72	1.7	2.1	2.1
43	1979	Nov.	6	3:04	70	1.7	1.3	---
44	1979	Nov.	12	7:21	44	1.1	1.2	---
45	1980	Jan.	6	13:50	66	1.6	(1.0)	---
46	1980	Feb.	18	3:58	42	1.1	---	.9
47	1980	Apr.	10	22:33	30	.7	---	.9
48	1980	Apr.	22	3:14	110	2.2	(2.8)	---
49	1980	Apr.	26	3:59	58	1.4	(3.0)	1.7
50	1980	May	18	3:31	36	.9	---	1.2
51	1980	May	18	22:33	13	-0.3	---	(1.2)
52	1980	July	7	17:02	20	.2	---	(1.4)
53	1980	Aug.	4	10:13	30	.7	---	1.0
54	1980	Sept.	18	1:28	22	.3	---	.7
55A	1980	Sept.	21	10:02	56	1.4	(2.6)	1.5
56	1980	Sept.	24	6:41	42	1.1	(2.0)	1.4
57	1980	Sept.	26	1:31	89	2.0	(3.5)	2.2
57A	1980	Sept.	26	5:04	19	.1	---	(1.0)
58	1980	Oct.	9	1:47	14	-0.2	---	.4
59	1980	Oct.	11	22:40	31	.7	---	1.1
60	1980	Oct.	14	1:20	71	1.7	---	1.9
61	1980	Oct.	16	3:48	44	1.1	---	1.5

[1]Refers to table 4 and G. A. Bollinger (1980, unpub. data).
[2]Parentheses identify unacceptable values, probably caused by use of an inappropriate wave period or phase. Do not use these values.

APPENDIX C

VELOCITY MODEL TEST FOR GILES COUNTY LOCALE

By D. A. CARTS *and* G. A. BOLLINGER

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(Modified from Carts, 1980)

A microearthquake was detected at 04^h UTC, February 18, 1980, by seismograph stations of the Giles County, Bath County, and central Virginia subnetworks. A preliminary location placed the epicenter near the northeast edge of the Giles County seismic subnetwork.

The arrival times for this event were used to test the several velocity models that are available for the Giles County locale. Specifically, there are now five velocity models in use, three of which are "regional" (Southeastern United States) and two of which are "local" (Giles County). One of the regional models (MCC) and the two local models (TPMl, TPM2) were recently published (Bollinger and others, 1980). Each model has 2-4 horizontal layers of uniform velocities and thicknesses. Model names have no significance beyond identifying the models; most are initials of the authors cited in the table below. All the models are as follows:

[km/s, kilometers per second; km, kilometers]

Model	V_P (km/s)	Depth (km)	V_P/V_S	Remarks
GAB	6.24	0.0	1.70	Regional. Bollinger, 1970.
	8.22	45.0		
VPI	5.7	.0	1.70	Regional, hybrid. Unpublished.
	6.24	10.0	1.70	
	8.22	45.0	1.70	
MCC	6.34	.0	1.67	Regional. Chapman, 1979.
	8.18	45.0	1.73	
TPM1	5.63	.0	1.64	Local. Moore, 1979.
	6.53	10.0	1.70	
	8.18	49.0	1.71	
TPM2	5.63	.0	1.64	Local. Moore, 1979.
	6.05	5.7	1.72	
	6.53	14.7	1.70	
	8.18	50.7	1.71	

COMPARISON OF LOCATION CAPABILITY OF THE VELOCITY MODELS

Arrival time data were read from the seismograms and were used as input to HYPO71 (Lee and Lahr, 1975). Initial runs were made to eliminate arrival times with large traveltime residuals. Next, each model was tried with one or more different ratios of compressional velocity to shear velocity (V_P/V_S). All runs had a trial focal depth (TFD) set equal to zero. The results of the eight runs are tabulated:

[Error measures are calculated by HYP071 (Lee and Lahr, 1975): RMS, root-mean-square error of the traveltime residuals in seconds (s); ERH, standard error of the epicenter and ERZ, standard error of the focal depth in kilometers (km)]

Model	V_P/V_S	RMS (s)	ERH (km)	ERZ (km)	Quality[1]
GAB	1.70	0.33	1	386	C
VPI	1.70	.26	1	2	C
MCC	1.67	.46	1	542	C
MCC	1.73	.30	1	357	C
TPM1	1.64	.52	2	5	D
TPM1	1.70	.48	2	4	C
TPM2	1.64	.52	2	5	C
TPM2	1.70	.21	1	1	B

[1]The 68-percent confidence ellipsoid calculated by the HYPOELLIPSE program (Lahr, 1979) is projected onto horizontal and vertical planes, to give lengths and orientations of semimajor, semiminor, and vertical semiaxes. The largest of these three distances determines quality. Quality is A if the largest distance does not exceed 2.5 km, B if the largest distance does not exceed 5.0 km, C if the largest distance does not exceed 10.0 km, and D otherwise.

On the basis of the lowest RMS, ERH, and ERZ values and highest hypocenter quality, model TPM2 with V_P/V_S=1.70 appears to be the best model. Also, only the TPM2 and VPI models calculated a focal depth different from zero trial depth.

STABILITY OF FOCAL DEPTH ESTIMATED WITH CHANGES IN TRIAL FOCAL DEPTHS

The TPM2 velocity model with V_P/V_S=1.70 was used with several trial focal depths (TFD's). We observed the stability of the estimated focal depth as it was calculated with HYPO71 (Lee and Lahr, 1975). TFD's were chosen to be in each layer and near some layer boundaries. Results are as follows, in the form of calculated values of origin time, hypocenter coordinates, and error measures for each TFD:

[km, kilometers; s, seconds; min, minutes]

Trial focal depth (km)	Origin time (0358 + x) (s)	Latitude (N.) (37° + x) (min)	Longitude (W.) (80° + x) (min)	Focal depth (km)	RMS (s)	ERH (km)	ERZ (km)
00	55.28	25.68	35.37	13.1	0.21	0.8	1.5
3	55.34	25.72	35.36	12.1	.20	.8	1.5
4	55.27	25.70	35.37	13.2	.21	.8	1.5
5	55.35	25.63	35.34	12.1	.20	.8	1.5
6	55.34	25.57	35.42	12.6	.20	.8	1.5
10	55.24	25.62	35.44	13.9	.20	.7	.8
12	55.21	25.60	35.44	14.4	.19	.6	.8
13	55.21	25.56	35.45	14.6	.20	.6	.6
14	55.18	25.73	35.41	14.7	.19	.6	1.9
15	55.17	25.74	35.36	15.0	.19	.6	.8
18	55.17	25.66	35.31	14.9	.19	.6	1.8
20	55.17	25.71	35.21	14.5	.19	.6	.6
25	55.18	25.78	35.42	14.7	.19	.6	1.9
30	55.16	25.75	35.20	14.4	.19	.6	.6

It is seen that regardless of the initial depth estimate, a final focal depth near 14-km depth is obtained. Origin time and hypocenter coordinates vary little. All runs had B-quality solutions, RMS values of 0.20±0.01 and ERH values of 0.7±0.1. Runs with TFD near a layer boundary of the model TPM2 either tended to give ERZ values that were relatively large or did not change the focal depth from the TFD. Expectably, deeper focal depths are related to earlier origin times, but the latitude and longitude values were virtually independent of the focal depth.

PRELIMINARY CONCLUSIONS

The preliminary indications based on this one test are as follows: (1) The TPM2 model with V_P/V_S=1.70 is the best model for the Giles County area, (2) epicenter estimation is relatively stable with changing TFD, (3) shallower and deeper TFD's tend to produce slightly shallower and deeper focal depths, respectively, (4) TFD's near layer boundaries should be avoided, and (5) a TFD should be tried from each layer to ascertain stability of focal depth.

APPENDIX D

STATISTICAL TESTS OF THE GILES COUNTY SEISMIC ZONE

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INTRODUCTION

The nearly vertical, tabular seismic zone beneath Giles County, Va., is defined by only eight microearthquakes out of the 12 that have been located in and near Giles County during 1978-80. It is possible that the zone is only an artifact of the small sample size and would disappear with more data. The 12 microearthquakes recorded during 1978-80 are a sample of a population that might contain hundreds or thousands of earthquakes. The population consists of all the earthquakes that have occurred in or near Giles County during the past decades, centuries, and millenia, and all the earthquakes that will occur for a similar time into the future. The next few decades to millenia is the span of time about which evaluators of seismic hazard must be concerned. Accordingly, this appendix must evaluate the likelihood that a 3-year sample of 12 microearthquakes is an adequate basis for characterizing the population.

Properly chosen statistical methods that are applied to a representative sample can evaluate this likelihood, because such methods incorporate the effect of sample size. In particular, the methods lose power as sample size decreases. This power loss means that for small samples the methods used below will produce valid but conservative results. The methods may fail to detect associations that are only weakly significant, but the methods will not overemphasize the significance of marginal associations.

In Appendix D of Bollinger and Wheeler (1982), Wheeler attempted a straightforward statistical analysis of microearthquake locations. That analysis fell into a pitfall that might be called "the one-sample problem". Proper statistical procedure is to collect a sample of data and examine the structure of the sample. Results of the examination aid in the formulation of an hypothesis to be tested. Then the examined sample is set aside and the statistical test is performed on a second, unexamined sample of data. If examination and testing are both performed on the same sample, then the formulation of the hypothesis is likely to reflect vagaries in the sample more strongly than it reflects the structure of the population. The results of the statistical test are likely to be distorted toward statistical significance. In general, the amount of such distortion cannot be determined.

Only one sample of microearthquake locations exists for the Giles County locale. This sample consists of the 12 microearthquakes that were recorded during 1978-80. The sample is too small to divide into two parts. In Appendix D of Bollinger and Wheeler (1982), Wheeler used this single sample of 12 microearthquake locations for both examination and testing, and concluded that eight of these locations define a statistically significant spatial pattern. This conclusion might be correct, but the methods that Wheeler used in Bollinger and Wheeler (1982) cannot determine that because results of the analysis have been distorted by prior inspection of the tested sample.

Methods are needed that do not require inspection of data before performing a statistical test. This appendix applies such methods, in the third of the following four steps: (1) Arguments are presented that the sample is representative of the population from which it was drawn. (2) The epicentral alignment to be examined is shown to be the most appropriate alignment among the several that might be perceived among the 12 microearthquake locations. (3) A procedure that does not depend on prior inspection of the data allows evaluation of the suggested concentration of earthquakes in the Giles County seismic zone and allows evaluation of the suggested northeast alignment of locations of these earthquakes. (4) The strike and dip of the tabular seismic zone are calculated.

The 12 microearthquakes to be considered are represented in figures 9, 11, and 13. The microearthquakes are characterized by the data of table 4. Pertinent parts of these figures and data are included in the figures of this appendix.

REPRESENTATIVENESS

The 3-year sample of microearthquakes cannot be used to characterize the longer term seismicity of the Giles County seismic zone unless the sample is representative of that longer term seismicity. The sample must be representative in both space and time and must also be what is called complete. Representativeness in space means that earthquake detection and location have not been biased against earthquakes outside the Giles County seismic zone. Representativeness in time means that the sampled time interval, 1978–80, has not been anomalous in its seismicity. Completeness means that no earthquakes that occurred in the area of interest during 1978–80 have been missed, except perhaps earthquakes smaller than those in the sample. Because there is only one sample, and because it is small, quantitative demonstrations of representativeness and of completeness cannot be made. However, the following arguments suffice for our purposes.

The goals of this report do not require that the 3-year sample of microearthquakes perfectly represent the longer term seismicity of the Giles County locale. These goals only require that the sample be representative enough in time that the map pattern of the epicenters is defined clearly enough to justify geological interpretation. The geological and seismological arguments of the next two paragraphs support the conclusion that the sample achieves this level of representativeness in time.

The seismicity of Giles County occurs when existing faults are reactivated in an existing stress field (Hamilton, 1981). Probably important new faults are not formed, so the geologic structures that are responsible for the seismicity do not change over decades to millenia. The present stress field arises from motions and mechanical properties of the North American plate and from the interactions of the plate with adjacent and underlying material; these factors can change over millions of years but are unlikely to change appreciably over decades to millenia. Thus, it is geologically reasonable to regard the earthquakes sampled during 1978–80 as a representative sample of several decades to millenia of seismicity.

Seismicity of the Giles County locale is likely to change with time in two ways. First, earthquakes that are large enough for their recurrence intervals to exceed 3 years will occur in some 3-year samples but not in others. As an extreme example, only samples that include the year 1897 would sample an earthquake of intensity VIII. However, earthquakes that are large enough to have remained unsampled during 1978–80 will change the observed map pattern of epicenters only if they occur on faults that have not generated detectable microearthquakes during 1978–80. The geological arguments of the preceding paragraph make such variable activity unlikely. Second, the abundance of small earthquakes and microearthquakes can rise and fall with time, although locations of these earthquakes tend to remain more or less the same. The historical record of seismicity indicates that the number of earthquakes observed in and near Giles County does fluctuate over several decades (Bollinger, 1975). However, such fluctuations can be regarded as variations in the accumulation and seismic release of stress about a long-term average. Such fluctuations do not appear to indicate changes in the long-term level of seismicity or changes in which fault or group of faults are reactivated over decades to millenia. For these geological and seismological reasons, it is reasonable and safe to assume that the sampled microearthquakes are representative in time to a degree that justifies interpretation of their locations.

Whether the sampled microearthquakes are representative in space depends on whether the ability to detect and locate small earthquakes is the same throughout the area in and around Giles County. Figure 29 shows locations of the 12 microearthquakes that constitute the sample. The Giles County seismic zone is defined by the map locations of 8 of the 12 microearthquakes: nos. 32, 33, 35, 37, 38, 46, 58, and 63. The other four microearthquakes occurred outside the zone: nos. 34, 39, 40, and 60. Is a small earthquake as likely to be detected and located if it occurs outside the zone as if it occurs in the zone? An answer is obtained by comparing magnitudes and estimates of the quality of hypocentral locations for two groups of microearthquakes, the eight that lie in the seismic zone and the four that lie outside of it. Figure 30 makes this comparison graphically.

Figure 30 shows several relationships. As expected, there is a general improvement in the quality of hypocentral locations with increasing magnitude. This improvement occurs because larger earthquakes are more clearly recorded at more stations. However, there is no clear difference in locational quality between the microearthquakes inside the seismic zone and those outside the zone. Both groups of earthquakes have median quality of C. Also, there is no indication that earthquakes recorded inside the zone are preferentially smaller than those recorded outside the zone, as would be expected if detection ability and locational ability decrease away from the zone. The median magnitude of the earthquakes outside the zone is half a magnitude unit larger than is the median magnitude of earthquakes inside the zone, but for such small samples this difference is not important. In fact, the smallest earthquake of the 12 is no. 39, which occurred farthest from the seismic zone (figs. 29, 30). Therefore, it seems

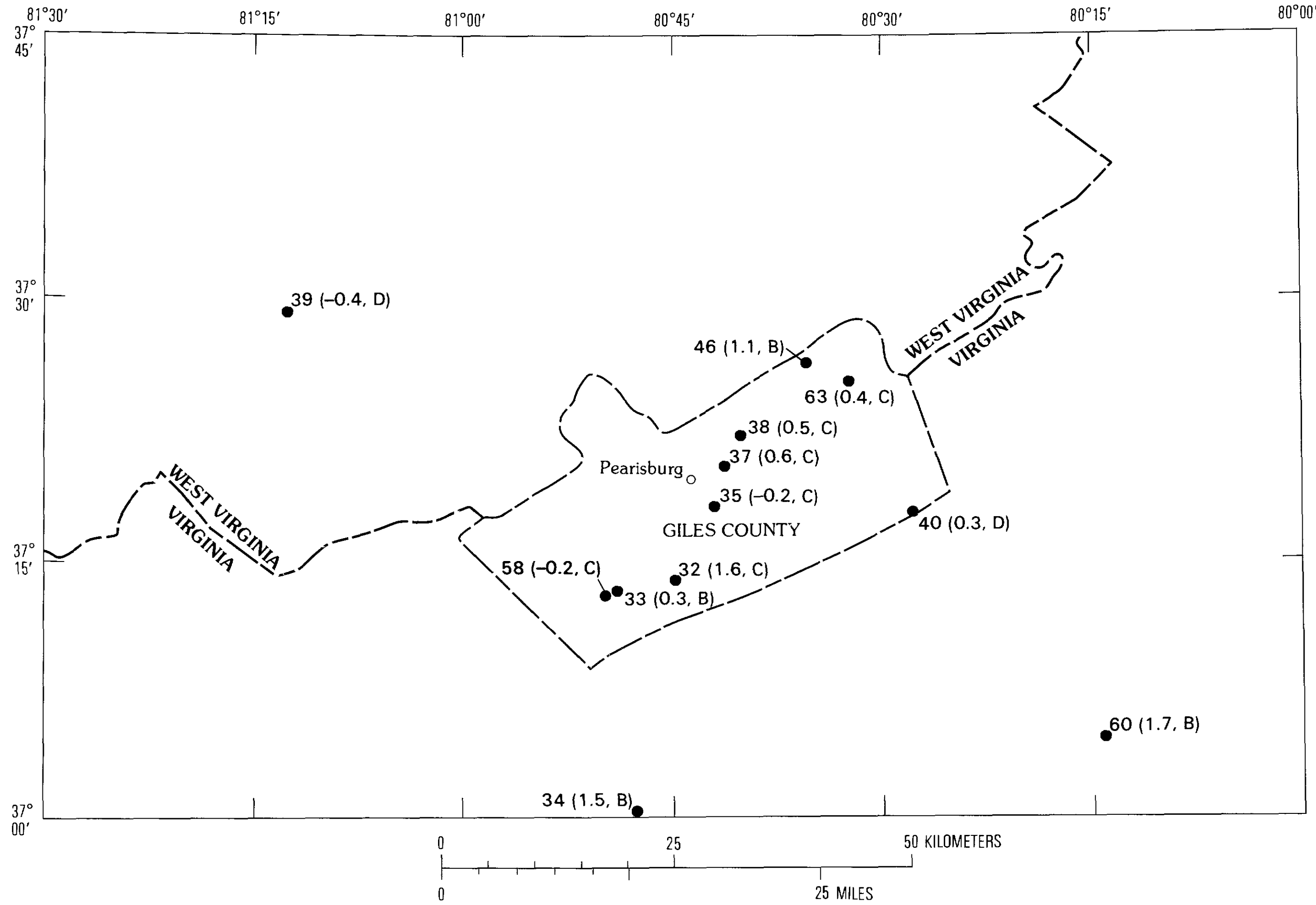


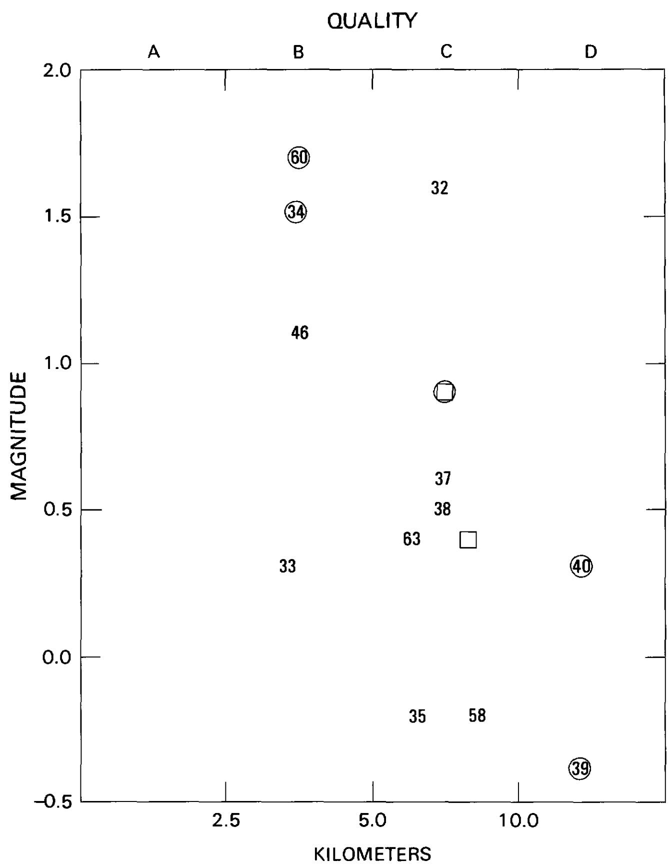
FIGURE 29.—Locations, magnitudes, and estimates of locational quality for the 12 microearthquakes that occurred in the Giles County locale in 1978–80. The locale is defined as the area within 50 km of Pearisburg. Dots locate calculated epicenters. To the right of each epicentral location appear three items: (1) the catalog number of the earthquake, (2) in parentheses, a decimal fraction giving the calculated magnitude (M_D) of the earthquake to the nearest tenth of a magnitude unit, and (3) also in parentheses, a letter indicating the quality of the calculated hypocentral location. Quality is determined from the largest of the horizontal and vertical dimensions of the 68-percent confidence ellipsoid (Lahr, 1979). A, B, and C quality locations have this largest dimension less than or equal to 2.5, 5.0, and 10.0 km, respectively; D quality solutions have this largest dimension greater than 10.0 km. Data from figure 9 and table 4 of this report.

reasonable to assume that the ability to detect and locate earthquakes is more or less constant across the Giles County locale. This assumption implies that the sample of 12 microearthquakes is representative in space and that sampling has not been biased against earthquakes occurring outside the seismic zone.

If the ability to detect and locate microearthquakes was about the same throughout the Giles County locale during 1978–80, then probably all earthquakes above some minimum size were detected and located, wherever in the locale they occurred. Judging from the magnitudes and locations of microearthquakes represented in figures 29 and 30, that minimum size is probably about magnitude -0.2, and perhaps as small as magnitude -0.4. Thus, the sample of 12 earthquakes is probably complete above a magnitude of about -0.2. Because the sample can be assumed to be representative in space and time, and because the sample is probably complete, the sample represents the population and can be used to characterize it. One aspect of the population that cannot be characterized by the 3-year sample of microearthquakes is the abundance of larger earthquakes.

THE MOST APPROPRIATE ALIGNMENT

I know of no statistical test that is generally appropriate for detecting and evaluating single alignments



EXPLANATION

34 Magnitude and quality of earthquake no. 34

○ Earthquake occurred outside the Giles County seismic zone

□ Point with the coordinates of the median quality and median magnitude of a group of earthquakes

FIGURE 30.—Relationship between earthquake magnitude and locational quality for the 12 earthquakes represented in figure 29. Uncircled box shows median of coordinates of 8 earthquakes inside the seismic zone and circled box does the same for 4 earthquakes outside the zone.

of points in a plane. For example, one problem is the difficulty of constructing mathematical expressions of such perceptual concepts as alignment, the maximum allowable gap between aligned points, and the effect of the mean areal density of all points in the sample. However, for the present case only, such problems of quantifying perceptions can be ignored. This is because many geologists and seismologists have examined figures 9 and 13. None have objected to the suggestion that if there is a significant tabular zone of microearthquakes, it is the one defined by earthquakes 32, 33, 35, 37, 38, 46, 58, and 63 in figures 9 and 13. It remains to determine whether we are all correct in assuming that the tabular zone did not arise randomly.

SIGNIFICANTLY CONCENTRATED EPICENTERS

In the discussions of representativeness and of appropriateness of the alignment, the sample of 12 microearthquakes was examined. Any subsequent statistical tests that are based on the findings of this examination would produce distorted results. Accordingly, the procedure summarized next does not use those findings and would be performed in the same way whether or not the sample had been examined.

The area represented in figure 29 can be divided into 18 rectangular cells, each measuring in minutes 15×15. The 12 epicenters fall into various cells. We wish to determine which cells contain significantly large numbers of epicenters. Such significantly overpopulated cells will occur independently of each other. If significantly overpopulated cells cluster along the alignment of eight epicenters in Giles County, then the alignment can be considered as a real and reliable feature of the population.

Cell significance is evaluated with the Poisson distribution and Chi-squared test, following procedures and examples of Lewis (1977, p. 76–78, 228) and Feller (1957, p. 146–154). The cell boundaries are drawn along the lines of latitude and longitude that are indicated around the boundaries of figure 29. To avoid distortion peculiar to any one choice of cell boundaries, the calculations are repeated for three other sets of cells, with boundaries moved successively 7.5 minutes east, 7.5 minutes south, and both.

Results vary among the four sets of cell boundaries (fig. 31). Three choices of cell boundaries found significantly many epicenters in one cell per choice; these three cells are hachured in figure 31 and overlap in the center of Giles County. The fourth choice of cell boundaries found significantly many epicenters in each of the three adjacent 15′×15′ cells that together cover most of the area inside the dashed lines of figure 31. For all choices of cell boundaries taken together, most cells contained either no or one epicenter, and the largest concentration of epicenters in any cell was six. Concentrations of four or more epicenters in a cell were significant and, for one choice of cell boundaries, so were concentrations of three or more epicenters in a cell. This variability can be overcome by summing results for all four choices of cell boundaries. Figure 31 does this.

Patterned areas in figure 31 show cells or parts of cells that contain significant concentrations of epicenters for two or more of the four different choices of cell boundaries. These cells overlap the alignment of eight epicenters that define the Giles County seismic zone, and align northeast-southwest along the zone.

The observations of the preceding paragraph are not subject to a numerical test of significance, but they

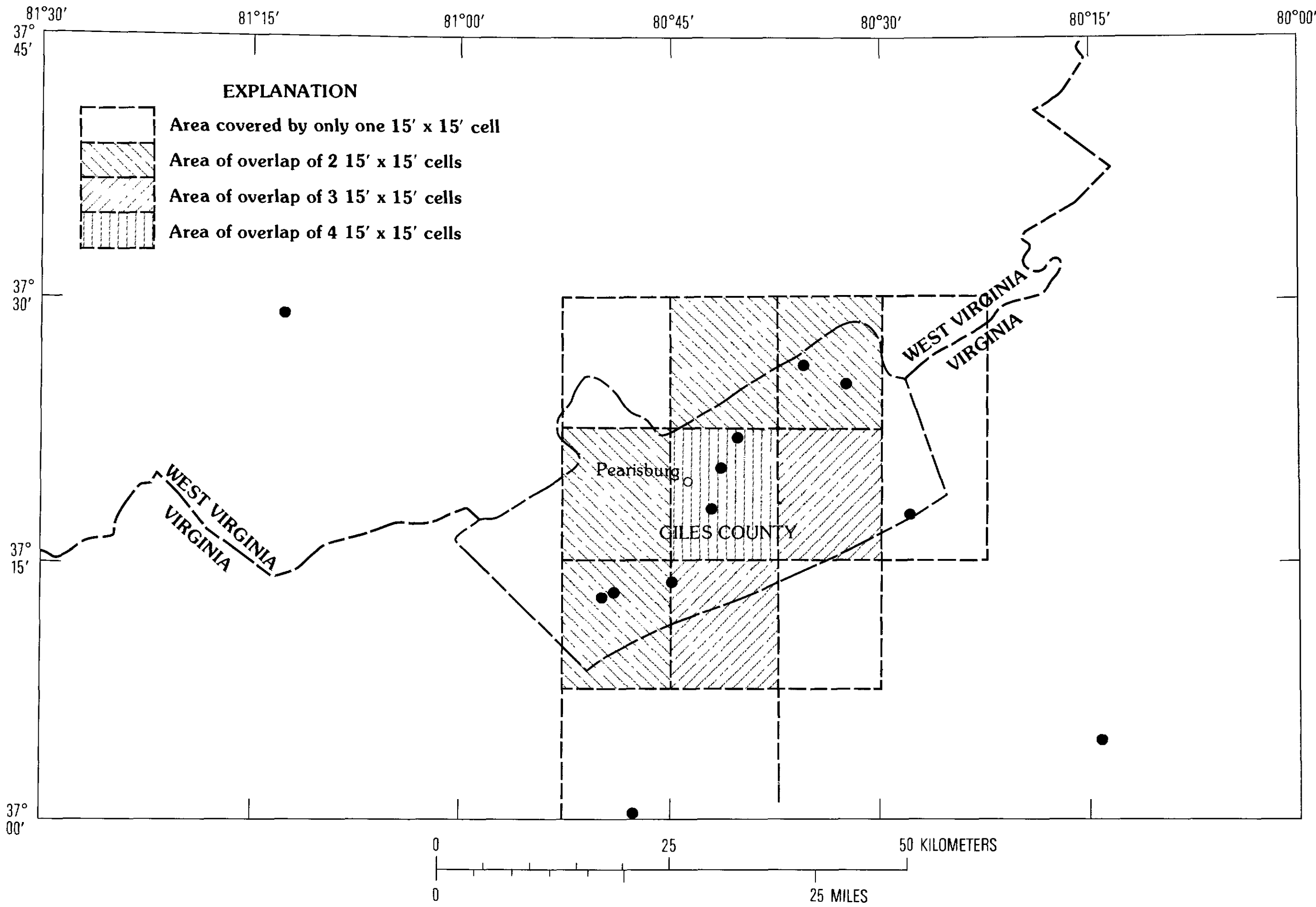


FIGURE 31.—Significant concentrations of the 12 epicenters represented in figure 29. Dots locate the epicenters. Dashed lines outline the 15′×15′ cells that contain significantly many epicenters.

provide support for concluding that (1) the earthquakes of figure 29 are concentrated into Giles County, (2) the earthquakes in Giles County define an elongated zone that trends northeast, and (3) the concentration, its elongation, and the trend of the elongation characterize a representative sample, and so also characterize the population of earthquakes from which the sample of 1978–80 was drawn.

STRIKE AND DIP OF THE SEISMIC ZONE

The eight hypocenters in the seismic zone define a tabular zone. The zone is presumed to reflect a fault or fault zone, and its strike and dip can be estimated. Earthquakes 34, 39, 40, and 60 lie outside the zone. These four earthquakes are presumed to have occurred on some other fault or faults, and will not be considered further.

To measure the northeasterly strike of the seismic zone, the eight hypocenters are projected up into a horizontal map at ground level (fig. 32*A*). Although deep hypocenters tend to occur farther northwest than do shallow hypocenters (fig. 32*B*), this upward projection will not distort the strike of the seismic zone much because the dip is steep. A straight line fitted to the epicentral locations will allow measurement of a numerical value of the strike of the seismic zone. To measure the dip of the seismic zone, the eight hypocenters are projected into a vertical plane that lies about perpendicular to the strike of the tabular zone (fig. 32*B*). Within the vertical plane, a straight line fitted to the hypocentral projections will allow measurement of a numerical value of the dip of the seismic zone.

In three dimensions, the eight hypocenters are somewhat scattered. No single straight line in figure 32*A* can pass through all eight earthquake locations, and the same is true of figure 32*B*. The regression coefficients that are calculated below allow qualitative estimates of whether that scatter is small enough that

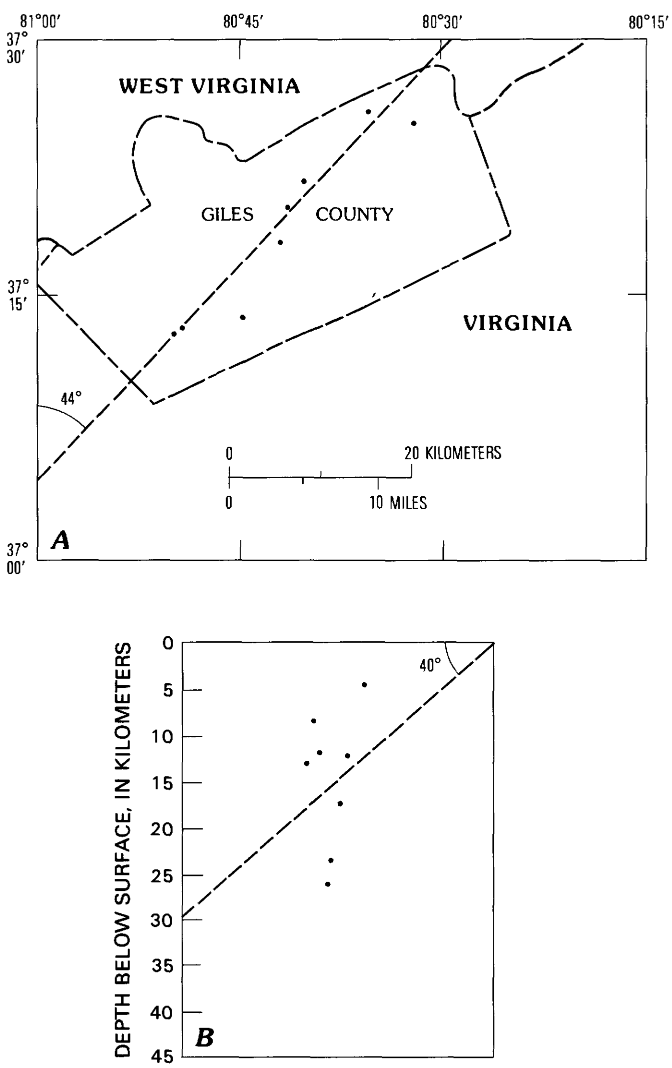


FIGURE 32.—Illustrations of strike and dip of Giles County seismic zone. Modified from figures 11 and 13. Dots locate epicenters and hypocenters of the eight numbered microearthquakes that define the seismic zone. *A*, Locations of hypocenters projected vertically up into epicenters. Dashed line locates regression line fitted to epicenters. Regression line trends 44° east of north. *B*, Locations of hypocenters as projected into vertical plane that strikes northwest, about perpendicular to trend of seismic zone, parallel to section *B–B'* of figure 11. No vertical exaggeration. Dashed line locates regression line fitted to hypocenters. Regression line plunges 40° to northwest.

it can be ascribed to random factors, such as uncertainty in calculated locations of hypocenters and irregular shapes of the faults that underlie Giles County.

The two straight lines that have been fitted to the points in figures 32*A* and 32*B* are regression lines. Although regression is a statistical technique, it is used here merely as a method of fitting lines to points. Several considerations govern the applications of regression methods to the points represented in figure 32: (1) The straight lines are fitted to the points of figure 32 by regression rather than by eye and ruler, because regression is more objective and its results are more reproducible. (2) The regression calculations provide values for Pearson's product-moment correlation coefficient r. Consideration of these values can provide qualitative estimates of the confidence that can be placed in the regression lines and of the degree to which the fitted points adhere to the regression lines. Such qualitative estimates are aided by visual examination of figure 32. (3) The values of r cannot be tested for significance, so standard numerical estimates of the goodness of fit of the points by the regression lines cannot be made. Significance testing is ruled out by the one-sample problem. Inspection of the distribution of the hypocenters in map and section views led to the observation that the seismic zone is tabular. This observation led to the desire to fit straight lines to the hypocenters as they are represented in figures 32*A* and *B*. Therefore, results of any significance tests would be distorted. (4) Full quantitative use of the simple linear regression used here is inappropriate for points that have uncertainty in all coordinates (Bolt, 1978). Use of the full panoply of regression techniques may also be inappropriate if it is not clear which coordinate should vary as a function of the other (Williams, 1983). Both problems occur here. The hypocentral locations of figure 32 have small uncertainties in all directions. In both parts of figure 32, it is unclear which coordinate ought to depend on the other. (5) Simple linear regression and testing r for significance require ten assumptions, which range from representativeness of the sample to statements about statistical properties of the scatter of data points about the regression lines (Kmenta, 1971, chaps. 7, 8; H. W. Rauch, R. F. Lamb, and P. A. Lentz, oral communs., 1973–79). For the data represented by the points in figure 32, the net effect of most of these assumptions is that the regression lines and the values of r remain mostly valid, but any tests of significance will produce unreliable results. Spearman's rank-correlation coefficient r_s does not require most of the assumptions that cause problems with r, so values of r_s are also calculated.

In map view (fig. 32*A*), the regression line trends N. 44° E., which is therefore the estimated strike of the seismic zone. r=0.95, and r_s=0.98. For both coefficients, values may range from −1 to +1, with values near either extreme indicating reliable fits of the data to the dashed lines of figure 32, and values near zero indicating little evidence for such fits. Thus, the alignment of the eight epicenters of figure 32*A* appears to be strong, despite the small sample size. Scatter of the eight points about the regression line may be attributed to random effects. The seismic zone strikes about

N. 44° E., although inspection of figure 32*A* suggests that this value could vary a few degrees either way with more data. For example, Bollinger and Wheeler (1980a, b), Wheeler and Bollinger (1980), Bollinger (1981a, b), and Hamilton (1981) cited strikes of N. 36° E. and N. 37° E. These strikes were calculated before the occurrences of earthquakes nos. 58 and 63. Similarly, added data caused the strike to increase one degree from the N. 43° E. of Bollinger and Wheeler (1982). The changes of strike are not large enough to alter any conclusions of those reports.

In section view, the regression line plunges 40° NW., which would therefore be the estimated dip of the seismic zone. However, r=-0.23 and r_s=-0.12. Neither coefficient has a value close to -1, so there is no strong relationship between distance to the northwest in figure 32*A* and hypocentral depth. These equivocal values of r and r_s could arise in one of three ways, all consistent with the strong epicentral alignment of figure 32*A*: (1) The earthquakes could be occurring in a linear or cylindrical zone that is oriented horizontally; (2) the source of the earthquakes could be tabular but elongate in a northeast-southwest direction and nearly horizontal; or (3) the source of the earthquakes could be tabular and nearly vertical. Figure 32*B* favors the third interpretation and indicates that the dip is steeper than the calculated 40°. The eight hypocenters occur on a nearly vertical tabular zone of unknown dip, but a steep northwest dip is more likely than a dip to the southeast.

SUMMARY

Results of statistical tests and other statistical procedures, when applied to the hypocenters of the eight microearthquakes of the Giles County seismic zone, allow the following conclusions. The sample of microearthquakes collected during 1978–80 is probably representative of the larger population of earthquakes that have occurred in and near the seismic zone over the past decades to millenia—and which will occur there for a similar time into the future. Accordingly, these microearthquakes may be used to characterize expected future seismicity of the seismic zone. Earthquakes larger than those sampled can be characterized by probable location but not by probable abundance.

Earthquakes concentrate in the seismic zone and are distributed tabularly. The tabular zone strikes N. 44° E. and dips steeply, probably to the northwest. The strike may vary a few degrees with more data but is not expected to change any large amount.

APPENDIX E

STATISTICAL TESTS OF THE COMPOSITE FOCAL MECHANISM SOLUTION

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The composite focal-mechanism solution (fig. 16) is based on a small number of first motions. We wonder whether the observed pattern in the data could have arisen by chance. The statistical test that can determine the significance of the observed pattern is the binominal test (Lewis, 1977, p. 59–64; Mosteller and Rourke, 1973, p. 24–25). The binomial test requires that each first motion shown in figure 16 and listed in table 8 be independent of the others; that each first motion be either consistent or inconsistent with the solution, but not both; and that each first motion has the same probability p of being consistent under the null hypothesis.

These three requirements are met. The first motions are independent because no seismograph influences records produced at another. Each first motion is either consistent or inconsistent, and first motions that cannot be classified as compressions or dilatations have not been used. Under the null hypothesis discussed in the next paragraph, $p=0.5$ for each first motion.

The null hypothesis must be carefully worded in order to eliminate distorted results as explained below. Our null hypothesis is that compressional and dilatational first motions are equally distributed over the focal hemisphere, so that they show no preference for the southeast side of the fault that corresponds to the preferred nodal plane of figure 16 having moved either up or down with respect to the northwest side. Then, the one-sided alternative hypothesis is that first motions reflect reverse motion on the steeply dipping nodal plane in which the northwest side moved up relative to the southeast side. Because the neutral axis of the composite focal-mechanism solution is nearly horizontal (fig. 16), that motion is predominantly dip slip. As argued in the main text (p. 24), the movement is more likely to be high-angle reverse than high-angle normal.

Any statistical test produces distorted results if the null and alternative hypotheses are designed by first inspecting the data on which the test will be performed. Such biased hypotheses will reflect structure in the sample that may not be present in the population from which the sample was drawn. That danger persists even if sampling is rigorously representative, because the characteristics of the sample will always differ from those of the population by some (usually small) random amount. In practice, a test of such biased hypotheses produces anomalously low significance values, and the test may appear to find significance where an undistorted result would not. The standard protection against such distortion is to run the test on a second, uninspected sample. Clearly, no such second sample is available here.

We argue that such protection is unnecessary because the steeply dipping nodal plane is unbiased, and the shallowly dipping one is biased in a way that does not affect our results. The steeply dipping plane was determined by inspection of hypocentral locations of several earthquakes (figs. 9, 13) and not from inspection of the first motions of figure 16. The shallowly dipping plane was determined by (1) the constraint that the two planes be orthogonal, which does not introduce bias of the type under discussion and (2) further adjusting the plane's orientation to minimize or eliminate inconsistent first motions. Step 2 introduces bias, but because the shallowly dipping plane is not involved in the null or alternative hypotheses as we have worded them, that plane is not involved in the binomial tests and its bias does not distort results of the tests.

A binomial test, with $p=0.5$ and using all 14 first motions, gives a level of significance of 0.029. A conservative test using only the six impulsive first motions gives 0.109. The conservative result is not significant at the habitual level of 0.05, but both results provide some support for our conclusion of high-angle, mostly dip-slip motion, probably reverse, on the steeply dipping nodal plane. In particular, the tests suggest that there is no more than 1 chance in 9 or 10 that first motions located randomly on the focal hemisphere would produce P (pressure or compression) and T (tension or dilatation) fields as well defined as, or better defined than, those observed.

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