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THE GILES COUNTY, VIRGINIA, SEISMOGENIC ZONE--SEISMOLOGICAL RESULTS AND GEOLOGICAL INTERPRETATIONS

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NOTE

Figures 1 and most of 2, 10, 11, 12, 13, and 14 were drafted originally for this paper, but have already been published in Bollinger (1981a). They are republished here with permission.

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SEISMOLOGICAL RESULTS AND GEOLOGICAL INTERPRETATIONS

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ABSTRACT

This paper describes and interprets a newly-recognized 40-km-long seismogenic zone, which 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. This paper attempts to aid in evaluating that 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, 10 microearthquakes ($M < 2$) have been detected. Eight of those 10 events, plus an additional 4 relocated felt earthquakes ($3.2 \leq M \leq 4.1$; 1959-1976) have a tabular distribution centered at Pearisburg, Virginia. That distribution is about 40 km long, 10 km wide, strikes N. 43° E., and has a nearly vertical extent of from 5 km to 25 km in depth. Thus, a Giles County seismogenic zone is defined presently by 12 earthquakes that span 4 orders of magnitude ($0 \leq M \leq 4$) and 2 decades of time (1959-1980). We conclude that the 1897 earthquake occurred on that seismogenic zone. From the orientation of the tabular zone, from evidence that greatest horizontal compressive stress trends east-northeasterly 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 the seismogenic 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 detachment faults. The above-mentioned reliable hypocenters for the region lie below the deepest likely detachment, 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 today's 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, detachment faults loaded the crust with several kilometers of overthrust sedimentary rocks, perhaps forming northeast-striking thrust-load faults in a brittle analogue 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 N. 43° E.-striking seismogenic 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. First, the seismogenic zone probably does not occur on a thrust-

load fault. The zone underlies detached 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 detachments would pre-date the southern Appalachian structures. That deduction contradicts stratigraphic and structural estimates of relative ages of southern and central Appalachian detachments. 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.

SEISMICITY OF THE GILES COUNTY, VIRGINIA, LOCALE

Introduction

The May 31, 1897, earthquake that occurred in Giles County, southwestern Virginia, 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, et al., 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, and
3) prior to 1897, no earthquake activity 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 (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, Virginia, the presumed epicenter of the main shock of 1897; Table 1) or perhaps in 1917 (southwestern Virginia earthquake reported on April 19, no intensity given; Bollinger, 1975; Reagor and others, 1980a). There followed a quiescent period of 4 to 6 decades that ended in 1959 with the occurrence that year of 3 felt shocks (MMI = VI, IV, IV). The following 2 decades (1960-1979) saw 6 additional felt earthquakes (MMI $\leq$ VI) reported from the Giles County locale (defined herein as the region within 50 km of Pearisburg). The largest of those 6 shocks was the m_b = 4.6, MMI = VI, Elgood, West Virginia, earthquake of November 20, 1969 (felt area = 324,000 km^2). Elgood

is a small community just north of the Virginia-West Virginia border at Giles County. Thus, there has been an apparent modern renewal of seismic activity (9 felt earthquakes in the past 22 years) in or near Giles County, Virginia (see Table 1).

This paper has three purposes:

1) It presents and interprets results of a recent seismic monitoring program in the Giles County locale. This first section of the paper achieves this purpose.
2) The paper attempts to integrate those results with what is known or reasonably inferred about local and regional geologic structure at seismic focal depths. The second and third sections of the paper achieve this purpose. Our goal of integrating results from diverse portions of seismology and geology has required us to write for two audiences. Thus, we have included material that may seem unnecessary to members of one audience or the other. As geologists and seismologists reviewed drafts of this paper, some specialists in each discipline questioned inclusion of some of the details. 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 achieved, the paper can contribute to improved zoning for seismic hazard. Throughout much of the western United States, many known or suspected seismogenic 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 zoning benefit markedly. In sharp contrast, throughout most of the East any seismogenic faults are buried beneath sediments, sedimentary rocks or thrust sheets. Thus, zoning of seismic hazard in the East often must be based mostly or entirely on the historical seismicity record. However, in most eastern areas zoning would benefit if seismicity could be associated with individual faults or classes of faults, whether buried or exposed.

Since 1962, a Worldwide Standard Seismograph Network (WWSSN) Observatory (call letters: BLA) has been in operation at Blacksburg, Virginia, 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. 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.

Terminology

Throughout this paper, we will use the terms hazard, seismogenic zone and structure. Because usage differs from one specialty to another, because usage is changing rapidly, and because in some contexts the terms have economic and legal implications, we should describe the usages that we follow in this paper.

HAZARD, RISK: These concepts have been expressed as verbal descriptions, as numerical values, and as probabilities that a specified value will be

exceeded in a specified time at a specified site. However, one distinction is common, and we 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). Because we do not use the term risk herein, that distinction will suffice for our needs.

SEISMOGENIC ZONE: "A planar representation of a three-dimensional domain in the earth's crust in which earthquakes are inferred to be of similar tectonic origin. A seismogenic zone may represent a fault in the earth's crust" (Earthquake Engineering Research Institute Committee on Seismic Risk, 1981, p. 60). We shall also use this term to refer to the three-dimensional domain itself. Because the domain that we discuss in this paper is tabular and dips vertically or nearly so, its planar representations in map and section views do not distort its true appearance too much.

SEISMIC ZONE: We will not use this term because it has two different meanings. One is "a generally large area within which absolute seismic design requirements for structures are uniform" (Earthquake Engineering Research Institute Committee on Seismic Risk, 1981, p. 60). 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. Instead of using seismic zone in its second meaning, we will use the term seismicity, as in "seismicity of the Giles County seismogenic 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 made up 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 detachment-related folds and mostly shallow-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 stress enough 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 cautious synonym for fault.

Geologic Setting of the Giles County Locale

The Giles County locale is in the westernmost Valley and Ridge province of the Appalachians with some overlap to the west into the Plateau province (Figure 2 and Rodgers, 1970, p. 5-8). Ground elevation ranges from about 0.6 to about 1.2 km above sea level but is about 1 km in most places. North-northeast-trending structures of the central Appalachians in northern Virginia and West Virginia are replaced southwestward by, and interfinger with, east-northeast-trending structures of the southern Appalachians. The transition is across a diffuse zone several tens of kilometers wide in the vicinity of Roanoke, Virginia (see bend in provincial boundary in Figure 2 herein, and in Rodgers, 1970, Plate 1A).

The Giles County locale exposes unmetamorphosed sedimentary rocks of Cambrian through Pennsylvanian ages and of various compositions. The sedimentary rocks are underlain by metamorphic and igneous basement at depths of 10,000 to 19,000 feet (3 to 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). Paleozoic deformation caused by continental collision has thrown surface and near-surface sedimentary rocks into complex folds and faults, all of which have been detached from underlying rocks and thrust many kilometers northwestward, riding on one or more soft shaly units down to and including the Lower Cambrian Rome Formation. Accordingly, structures in rocks shallower than 3-6 km are probably largely unrelated to any 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 outcropping faults or other obvious aspects of surface geology and structures at the depths of the seismicity in the Giles County locale (at least 5 km, as documented in a following section on Analysis of Network Results).

On a continental scale, most large exposed, shallow subsurface and deeper crustal structures in many parts of the Appalachians, adjacent craton, and coastal plain are probably roughly parallel to each other. That is because the Atlantic Ocean opened approximately where an older ocean opened and then closed to form the Appalachians (Wilson, 1966; later section of this paper titled "Types of faults..."). Such a continental-scale parallelism 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 continental-scale parallelism of structures in easternmost 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 are in apparent conflict with that conclusion. First, independence of deep undetached and shallow detached structure is best established farther north, in the central Appalachians (Gwinn, 1964; Rodgers, 1963; and numerous subsequent papers). The possibility remains of subtle control of Paleozoic patterns of sediment dispersal and deposition 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, and 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 detached folds 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. Dewey and D. Gordon (written commun., 1980) calculated the location of the Elgood, West Virginia, earthquake (m_b = 4.6, event J of Table

7) and obtained a depth of 2.5 km subsurface. All other reliable hypocentral depths in the Giles County locale are deeper, within the basement. The top of basement near Elgood probably is at a depth of 4-5 km subsurface, so the Elgood focal depth is apparently well within the sedimentary rocks. However, from Dewey's and Gordon's results (written commun., 1980), the vertical semi-axis 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, subsurface. Furthermore, near Elgood the deepest thrust faults probably are only from 3 to about 3.5 km subsurface (fig. 4 of Perry and others, 1979; W. J. Perry, Jr., oral commun., 1980). As much as another 3 km of undetached sedimentary rocks underlie those deepest thrust faults, separating them from the top of metamorphic and igneous basement (Perry and others, 1979). Those deeper-lying sedimentary rocks are structurally part of the basement. 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 that earthquake having occurred either in the metamorphic and igneous basement, or in the undetached sedimentary rocks below the deepest thrust faults.

Finally, Herrmann (1979) calculated a depth of 5 km for the Elgood earthquake, using surface-wave data, and Carts (1981) calculated a well-determined depth of 13.6 km (HYPO71 (Lee and Lahr, 1975) location error parameters: ERZ = 1.4 km, solution quality = B) using a locale-specific velocity model. Thus, of the three depths calculated for the Elgood earthquake, two place it below the detached near-surface 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.

Network Monitoring Program

The Giles County network is a five-station subnetwork of the Virginia Tech Seismic Network (fig. 2, table 2). It is centered at Narrows, Va. (station call letters NAV), about 6 km west of Pearisburg, Va. The subnetwork aperture 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 operational since early in 1980). Signals from all five stations are telemetered to the Virginia Tech campus where they are recorded on 16-mm film and on FM analog magnetic tape. The central station (NAV) is also recorded on a visual recorder (pen-and-ink) and as an additional, low-gain (down 30 dB) channel on the tape recorder. Station magnifications range from 30 to 300K at 1 Hz depending on station and recorder-mode (see fig. 3). The frequency passband for all recording channels is set at 1-10 Hz. Average microseismic levels, as measured on the 16-mm film records, range from 1 to 60 nanometers at 0.6 to 3.2 Hz (Sibol, 1980; Appendix A of this paper). Figure 4 shows a magnitude 1.6 event that occurred near Narrows and that was recorded on both the BLA WWSSN SPZ and the network BLA SPZ (1-10 Hz passband). The

increased efficiency of microearthquake (defined herein as earthquakes with magnitude $\leq$ 2) 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 detection and location capability of the entire Virginia network, according to A. C. Tarr (written commun., 1980; see also Tarr, 1980), is illustrated in figures 5 and 6 (threshold 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, detection is complete down to a magnitude somewhat less than 1.5. Figure 7 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 (37° N.-81° W.), 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 Tech network by Viret (1980; see Appendix B of this paper). For larger events, at distances greater than about 0.5 degree, Nuttli's (1973) m_b(Lg) formulas 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 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 associated with hypocentral locations derived from those as well as other available velocity models (Carts, 1980; Appendix C of this paper) indicated that Moore's three-layer model, TPM2 (table 3), gave the minimal values. That velocity model has been used throughout this investigation.

Analysis of Network Events - January, 1978 through December, 1980

Using the TPM2 velocity model, hypocenters were recalculated for all the events that had occurred while the network was operating. The reductions in the hypocentral error measures from their pre-TPM2 values were substantial, and 8 of the 12 epicenters (Nos. 32, 33, 35, 37, 38, 46, 58, 63) coalesced to form a northeast-trending (N. 43° E.) alignment (fig. 8; table 6). The depth distribution of those foci defines a nearly vertical tabular zone extending from 5 to 25 km (fig. 12).

Tests of the Seismogenic Zone

Because the tabular seismogenic zone is defined by so few foci, it is desirable to test its existence, using evidence and arguments that are more objective than the visual impressions that are created by figures 8 and 12. Appendix D contains discussions of statistical tests. These results 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, there is a form of testing that can be done in the field. By locating known quarry or construction blast sites from their network P and S arrival times and then comparing those calculated locations with the actual field locations, the locational capability of the Giles County network can be demonstrated. Actually, only the epicenter (the horizontal coordinates of latitude and longitude) is tested in such a procedure, because the blasts are at the surface and not at the deeper earthquake focal depths. But, if the hypocenters determined from the blast data indicate shallow focal depths, then the velocity model is judged to be 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 9 shows the blast locations (designated as A, B, C) and their HYPOELLIPSE locations. Tables 4 and 5 give the 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, was significantly smaller (less explosive) than the other two blasts (A and B). Its network signals were not as clear (smaller signal to noise ratio), and thus its calculated location was expectably not as certain as those of the larger blasts.

An additional and important corroboration of the entire northeast-striking microearthquake zone was obtained from J. Dewey and D. 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). These were all felt events ($2.1 \leq M \leq 4.6$) that occurred between 1959 and 1976 and 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. 10 through 14; note that the location of station NAV serves as a 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 ($0 \leq M \leq 4$), two decades in time of occurrence (1959-1980) and location by two different research projects. Our judgment is that this constitutes a compelling case for the existence of the zone as we have described it even though the data base is not large.

Fault Area

Conventional epicenter maps and vertical-section plots of foci are given by figures 10, 12 and 13. Figures 11 and 14, on the other hand, are special-purpose illustrations that are designed to portray specific characteristics of the hypocenter data set in the horizontal (fig. 11) and vertical (fig. 14) planes. Figure 11 presents the epicenters, scaled according to magnitude, without any geography (except the location of the station NAV) or error ellipse axes to lead or guide the eye of the viewer. Thus, the spatial definition of the seismogenic zone is presented in map view with excellent clarity.

Figure 13 is a side view of the 12 seismogenic zone foci in vertical section. Note there and in figure 12 that two of the earthquakes (D and S) do not have vertical error semi-axes indicated on the figure, which indicates that there were insufficient arrival-time data available to determine adequately a focal depth even though the data were sufficient for calculation of an epicenter. In such cases, the depth is fixed by the geophysicist performing the calculations.

Note in figures 12 and 13 that error bars in both the horizontal and vertical directions are shown. These error bars are twice the semi-axes, in the plane of the projection, for the hypocenter error ellipsoids determined by the HYPOELLIPSE program (Lahr, 1980) or by the JHD program (Dewey, 1971). The most likely location of the hypocenter is, of course, at the center of the error ellipsoid. However, there is a specific level of probability (68 percent for HYPOELLIPSE and 90 percent for JHD) that *a given hypocenter could be located anywhere within its error ellipsoid*. *That fact has important* implications with respect to the areal dimensions of the seismogenic zone. Figure 14 illustrates that significance by showing the range of fault areas allowed by the 10 hypocenters. That range, from 80 km^2 to 800 km^2, was determined by *arbitrarily* moving the hypocenters *inside* their error ellipses in the following manner:

Upper Section - All hypocenters shifted toward the centroid of the hypocentral distribution. Note the superposition of groups of two and three hypocenters. A minimal area (80 km^2) 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 km^2.

Lower Section - All hypocenters shifted away from the centroid of the hypocentral distribution and (or) restricted to a minimum focal depth of 5 km. A maximal area of some 800 km^2 (shaded area) is thus defined.

Other ways of connecting the dots in figure 14 would produce slightly different inferred fault areas, but those areas would still vary over an order of magnitude, all consistent with the locational accuracy of the hypocenters. 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 km^2) can be used to estimate the magnitude of an associated earthquake. (A variation of 10 times in the fault plane area can imply a one-unit change in the magnitude of an associated earthquake; Wyss, 1979; Singh and others, 1980; Bonilla, 1980). The magnitude estimate can in turn be used to develop an assessment of the potential hazard associated with the seismogenic zone. Recently, Bollinger (1981a) has presented such an assessment for use by governmental officials and emergency planners but not for engineering applications.

The procedure used involves two major factors: (1) Specification of fault-plane area as an estimator of potential earthquake magnitude, and (2) development of a hypothetical intensity map. The first factor takes into account the spatial distribution of the hypocenter error ellipses (as in the preceding discussion), as well as published equations relating magnitude and fault plane area.

There are four subjective aspects of 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, areas of fault planes are usually estimated from spatial distributions of main shocks and their associated aftershocks. Instead, we use here mostly the spatial distribution of microseismicity detected during an extended period of time. However, the existence, orientation, and shape of the seismogenic zone as defined by the microseismicity are supported by the distribution of felt events. The zone has had seven felt events since 1959. Four of the seven were relocated by J. W. Dewey and D. W. Gordon (written commun., 1980; see also figs. 10-13). There were inadequate instrumental data to relocate the other three felt events that occurred in 1959 and 1975 (table 1). Thus, the Giles County seismogenic zone frequently has exceeded the energy-release levels of microearthquakes ($M < 2$).

(2) The confidence ellipsoids used to estimate minimal and maximal fault-plane areas (fig. 14) 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 circles in Figures 8-13. The relocations of Dewey and Gordon were calculated using the JHD (joint hypocenter determination) program, which produces 90 percent confidence ellipsoids. These relocations are shown as open circles in figures 10-13. 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 method to all 12 Giles County events showed that the hypocenters relocated by JHD have the same general location and trend as do those presented herein (Viret and others, 1981; Bollinger and others, 1982).

(3) The confidence ellipsoids are three-dimensional shapes with various orientations in space. Figure 14 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 14. The ellipses of Figure 14 are drawn using vertical and horizontal semi-axes. Consider how the ellipses would be distorted if they were drawn using semi-axes obtained by projection of the ellipses of Figure 14 into two other perpendicular lines lying in the plane of Figure 14, say lines plunging 45° to the southwest and to the northeast. Such projected axes would allow fault plane areas not much different from those shown in Figure 14. 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 semi-axes of Figures 10 and 12-14 and after consideration of the ellipsoidal semi-axes of Table 6, we conclude that this effect is also negligible for our purposes in three dimensions.

(4) The 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 (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 seismogenic zone. However, in the arguments based on Figure 14, we have already chosen the most extreme values of the area that are 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. That is, it would then predict that the seismogenic zone could experience largest shocks ranging from one smaller than that which has already occurred in 1897, to the largest one known to have occurred in the southeast (the 1886 Charleston, South Carolina, earthquake, m_b = 6.6 to 6.9, M_S = 7.4 to 8.0; Nuttli and others, 1979, Nuttli, 1981a and b). It seems to us that such a severely cautious conclusion would be of little use to anyone.

The second factor, 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. 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 (N43°E) while the lower level isoseismals (VII and below) are to trend with the tectonic fabric of the surrounding portion of the Appalachians (N75°E).

Bollinger's (1981a) estimate of the size of the largest shock possible on the Giles County seismogenic 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. 15) 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 zone. Only 14 P-wave polarities could be obtained (six impulsive, eight emergent; see table 8), however, because of the small size (low energy level) of the individual shocks. A reasonably constrained focal mechanism could not be obtained from the data set (we easily obtained three different solutions).

Further information is available, however, that can be used to constrain the focal mechanism solution. In order to glean as much as possible from the P-wave data set, the following procedures were employed:

(1) We used the microearthquake hypocenter distribution (figs. 8 and 12) to define the preferred nodal plane as striking N. 43° E. and dipping 80° northwest. In the preceding section on "Analysis of Network Results..." and in Appendix D, we noted that regression techniques yielded a dip of 59° for the tabular seismogenic zone. However, we concluded there that the regression was not clearly significant. Further, if we had used a 59° dip for the 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 12. We then found an auxiliary plane such that it encompasses the compressional field (9 of 10 readings) in the northeast to south azimuths. Figure 15 shows that plane to strike north-south and dip 14° to the east.

The resulting CFMS (fig. 15) 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 constrain tightly the orientation of the auxiliary nodal plane (fig. 15), and partly because the hypocenters do not tightly constrain the dip of the preferred nodal plane (fig. 12, Appendix D).

(2) Two of the four Dewey and Gordon (written commun., 1980) relocations in the seismogenic zone (table 7, fig. 10) have constrained 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 15 is valid, despite being based on a small number of first motions with several inconsistencies.

(4) We compared the Giles County, Virginia, CFMS with other, nearby focal mechanism solutions. With no previous focal mechanism solutions for events in the seismogenic 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, West Virginia, shock (event J in fig. 10 and table 7) and the 1973 Knoxville, Tennessee, 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 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, Virginia, zone. Note that the 1969 Elgood, West Virginia, shock was not directly in the Giles County zone, but rather some 25 km to the northwest of that zone (fig. 10).

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./70° SE.) has an orientation roughly similar to the strike and nearly vertical dip of the Giles County, Virginia, 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, Virginia.

Thus, from other focal mechanism studies we find some supporting evidence for seismically active, northeast-striking, steeply-dipping seismogenic zones in the general area and in the same geologic-physiographic province as the Giles County, Virginia, 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 SEISMOGENIC ZONE

Introduction

One of the purposes listed at the start of this paper is to attempt to integrate the seismological results just described with what is known or reasonably inferred about geologic structure at the depths of Giles County seismicity. This section and the next (on "State of Stress...") achieve that purpose. Our approach is to evaluate in turn each of several types of basement faults, in order to select the one that is most probably responsible for activity on the Giles County seismogenic zone.

The discussion to follow is long and involved because pertinent data are sparse. However, the effort is worthwhile: If Giles County seismicity can reasonably be attributed to a particular type of fault, and if faults of that type have detectable characteristics and occur within a predictable region of North America, then evaluators of seismic hazard may be aided in their efforts to estimate where and how frequently an earthquake like that of 1897 might recur. Thus, an evaluation of the geological characteristics of Giles County seismicity may lead eventually to enhanced hazard evaluation for urban centers, critical facilities and lifelines far removed from semi-rural Giles County itself.

Previous sections of this paper have documented that the current seismicity in the Giles County locale is concentrated in a nearly vertical, tabular zone that strikes N. 43° E. and extends from 5 to 25 km in depth; have argued that that zone is probably the source zone for the 1897 shock; and have concluded that structures responsible for the seismogenic zone lie in the basement, beneath detached surface and near-surface rocks, folds and thrust faults.

The Giles County seismogenic zone involves all of the upper half of the sub-thrust continental crust. The best estimate of local crustal thickness is 51 km, from travel-time analyses of local and regional earthquakes and quarry blasts, and from an unreversed refraction survey (Moore, 1979; model TPM2 of Appendix C of this paper). That estimate is consistent with a previous one e of 50 to 55 km derived from regional analysis of seismic travel-time terms of the ECOOE experiment (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 indicate a potential for large earthquakes. That is consistent with the suggestion by Bollinger (1981a) that the Giles County seismogenic zone could generate a large shock. Also, the depth distribution of hypocenters of the seismogenic zone is consistent with a suggestion by Chen and Molnar (1981) that continental regions are characterized by aseismic lower crust. The lower crust could be

aseismic because it is too ductile to support high stresses (Meissner and Strehlau, 1982), perhaps because stress-supporting grains of common minerals recrystallize in the lower crust (Toriumi, 1982).

Because of the size of the seismogenic 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, because only such processes could stress the entire upper crust and cause it to fail. Data with which to identify clearly such deep, seismogenic faults in the Giles County locale are not now publicly available, to our knowledge. Eventually, such identification may result from interpretation of new or proprietary deep seismic reflection lines, from detailed modelling 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 usefully constrain 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 seismogenic 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 that 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 subsections 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 seismogenic zone there. Some of those ancient faults formed as an ocean called Iapetus opened. Since that time, 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 them are still weak and would be reactivated in preference to forming new faults.

The rest of this section describes and evaluates the various types of such ancient faults, mostly in the chronological order in which they might have formed.

The metamorphic and igneous basement under and near Giles County lies in that part of eastern North America commonly considered to have been deformed, recrystallized or both during the Grenville orogeny, roughly in middle Proterozoic Y time (about a billion years ago) (Ammerman and Keller, 1979, p. 344; Bass, 1960; Bayley and Muhlberger, 1968; Black and Force, 1982; Lidiak and Zietz, 1976; Lidiak and others, 1981). Pertinent data are sparse, but high-angle faults that formed during or before the local Grenville deformational or thermal peak(s) should have been sufficiently deformed, annealed or both by Grenville events that they no longer constitute important strength discontinuities. R. C. Shumaker (written and oral communs., 1978-1981; 1982) is analyzing published and unpublished structural, stratigraphic, geophysical and oil and gas production data from central and southwestern West Virginia and eastern Kentucky. He suggests 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 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. The magnetic lineament approximates the ill-defined southeastern edge of a large graben called the Rome trough (Harris, 1975 and 1978; Shumaker, 1977). Reactivated basement faults in central and southwestern West Virginia and eastern Kentucky are probably parts of the Rome trough and need not have analogs in Giles County. (For similar but more detailed discussions of the potential for reactivation of faults formed before, during and after Paleozoic thermal peaks of the Appalachian orogenies, see Odom and Hatcher (1980). Those Paleozoic thermal peaks occurred tens of kilometers southeast of Giles County, and so did not affect the rocks under consideration here.)

Since the time of the Grenville thermal peak(s), the region surrounding Giles County has experienced three deformational episodes. Each of these episodes is known or can reasonably be inferred to have produced basement faults which might include the one assumed to be reactivated in today's stress field to produce the Giles County seismogenic zone. Each of the three episodes was caused by movements of the North American and adjacent plates, and is known or suspected to have produced different types of basement faults. Each fault type has specific and predictable properties throughout much of the region now occupied by the Appalachians and the Coastal Plain. Accordingly, considerations of possible links between Giles County seismicity and one or more of those three types of basement faults may have applicability to evaluating seismic hazard over a portion of easternmost North America.

Those three opportunities for formation of basement faults occurred (1) during crustal extension at the end of the Precambrian or in the early Paleozoic, as the Iapetus Ocean began to open, (2) during crustal loading later in the Paleozoic as Appalachian thrusting reached the Giles County locale, and (3) during renewed crustal extension in the early Mesozoic as the Atlantic Ocean began to open (Wheeler and Bollinger, 1980). Most of the rest of this section considers those three fault types, in chronological order, and attempts to evaluate their applicabilities to Giles County seismicity.

Iapetan Normal Faults

This subsection considers late Precambrian or early Paleozoic normal faults as candidate sources of Giles County seismicity. Such faults formed in North American cratonic crust as an ancient ocean opened, in the manner typical of early stages of passive (Atlantic-type) continental margins. Features in the Bouguer gravity field over the Appalachians are used below 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 into the craton.

Iapetus Ocean

The predecessor ocean of the Atlantic, which eventually closed to produce the various Appalachian and Atlantic Caledonide orogens from Alabama to Spitsbergen, 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 portion of the Paleozoic ocean, which separated the Eurasian and North American cratons (see also reviews by McKerrow and Ziegler, 1972a, and 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) carry the name Iapetus 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 (Figure 16; Woollard and Joesting, 1964; Earth Physics Branch, 1974; Haworth and others, 1980). 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 that complexity of the Bouguer field in New England, and because New England lies beyond the geographic scope of this paper, we shall restrict the following discussion to the central and southern Appalachians, south of the region of New York City. However, where pertinent, we shall cite papers and observations that deal with the northern and Maritime Appalachians.

R. Simpson, M. Kane and coworkers have produced and discussed 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 computer plotted in color and show both the Bouguer 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 16, is composite, being a geometrically complex eastward rise in Bouguer values. The rise and the portions of the Bouguer field on either side of it consist mostly of numerous irregularly linear, anastamosing highs and lows. Anomalies of many wavelengths 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 referring to a single gradient like that shown in figure 16, we shall refer to the eastward rise in the Bouguer anomaly field: The rise possesses as much internal structure as do the portions of the Bouguer field that it separates, but the portion of the gravitational field east of the rise generally has more positive Bouguer values than does the portion west of the rise (R. Simpson, oral and written communs., 1981).

Simpson and co-workers 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 and 5; Simpson and Godson, 1981, their figs. 4 and 5). 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 (for example, see fig. 17).

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 in this subsection, 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. Diment (oral commun., 1981) noted that the rise could have different causes in different portions of its length. Interpretations of the rise by various authors cited above include eastward crustal thinning (by transition to buried oceanic crust or by uplift of the upper mantle and lower crust on steep faults) and eastward change to denser crust (oceanic, 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) show 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) estimate from refraction and arrival-time data that crustal thickness decreases about 13 km southeastward across the gravity rise in parts of Tennessee, the Carolinas and Georgia. They show a decrease 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 estimate of 50 km at Blacksburg, Virginia, 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 paper). Similar eastward decreases in crustal thickness across the region of the rise were suggested by James and others (1968; decrease from 45-50 km to 35-40 km, as derived from seismic travel-time terms and corroborated by Chapman, 1979) and 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).

Area of Expected Occurrence of Iapetan Normal Faults

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. This is presumed to be true for all crustal levels below the Appalachian detachments, including those at the depths of Giles County seismicity (5-25 km). That interpretation is made because the wavelength filtering, which separates the total field into a short-wavelength (residual) part and a long-wavelength (regional) part, can be thought of, in part, as separating anomalies caused by sources within different depth ranges. That is, the residual field from the 100-km wavelength filter is regarded as composed mostly of anomalies caused by sources at upper 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. Simpson, M. Kane and W. Diment, oral and written communs., 1980 and 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 (fig. 17). 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. Simpson, written commun., 1981). Thus, the source of the eastward gravity rise, which we and others already cited 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.

Recall that seismicity in and near Giles County occurs beneath the detached sedimentary rocks that form the local thrust complex of the Valley and Ridge province. That complex is the cratonward tip of the much thicker sheet of detached, mostly metamorphic and igneous rocks that are now suggested to 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 and 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 and may be masked by detachments on which upper crustal and shallower rocks have been transported to the west. Thus, Iapetan normal faults could exist at all undetached crustal levels 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 limit can be found for the area in which Iapetan normal faults may occur in cratonic crust. It is necessary for us to estimate regions of likely occurrence for Iapetan and Atlantic normal faults separately, even though both faulting episodes probably produced structures of comparable size, orientation and style. That need arises because the two types of normal faults are separated in time by the Appalachian detachments and metamorphisms. Thus, the Iapetan and Atlantic normal faults could differ in properties, such as degree or type of annealing, which could 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. Local exceptions are possible, because the crust east of the rise is probably a heterogeneous mixture of 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. That reworking, by deformation and thermal events during various Paleozoic subduction episodes, may have been so extensive as to destroy Iapetan normal faults as weak zones.

Although the composition, thickness and history of the crust east of the gravity rise are known only approximately and locally, 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 include Armorica (Van der Voo, 1979b, 1980a, 1982) and various pieces of Avalon (for example, Simpson and others, 1980; Skehan and Murray, 1980; see review by Rast, 1980). Hatcher (1978), Long (1979) and Hatcher and Zietz (1980) infer that various blocks of mafic, granitic and mixed deep crust comprise much of the southern Appalachians

including the region southeast of the gravity rise opposite Giles County. Osberg (1978) concludes 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) consider a possible ancient analog to be the melange 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 suggest 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, and 1978; Hayes, 1978). Hatcher (1978) suggests 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. 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 near the end of convergence as 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 through a history as complex as that suggested for the Mediterranean region and the Alpine system by Dewey and others (1973).

Because of such known and suggested complexities, we see no inconsistency between the hypothesis that the gravity rise marks the eastern edge of relatively intact Iapetan continental crust, on the one hand, and the identification of basement of Grenville age in central Virginia (Glover and others, 1978). If such Grenville basement were originally part of North America, it could have arrived east of the rise in various ways. The basement of Grenville age could have remained attached to relatively intact North American crust but linked to it by continental crust thinned by Iapetan normal faults. Alternatively, the basement terrane 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 its 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. It seems unlikely that Iapetan normal faults would survive in such an assemblage, at least not as weak zones of crustal size on which stress might be preferentially released by seismic slip. Further, any such

faults that did survive might no longer have an orientation suitable for reactivation in today's ambient stress field (northeasterly to southeasterly trending greatest horizontal compressive stress; see compilations and reviews by Zoback and Zoback, 1980, and Evans, 1979). 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 and 1979; Van der Voo and others, 1979a,b; Irving, 1979; Harland, 1980; Van der Voo, 1980a,b, 1981, and 1982a,b; Kent, 1981; Van der Voo and Scotese, 1981; Williams and Hatcher, 1982). LeFort and Van der Voo (1981) suggest 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).

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. From Massachusetts south, Mesozoic basins and their bounding faults lie on or east of the rise, with two exceptions (for example, see fig. 16). The larger exception is the western portion of the Newark-Gettysburg Basin in New Jersey and Pennsylvania (Haworth and others, 1980). 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 where a sharp offset in the rise at the Virginia-North Carolina border crosses the middle of the Dan River Basin. Thus, the Mesozoic fragmentation of this portion 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 suggest control of Mesozoic extensional faults by Paleozoic compressional structures.

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 to 100 km southeast of the locale (figs. 2 and 16). If the rise marks the 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 Iapetus to have experienced normal faulting. For example, on the edges of the Red Sea, Lowell and Genik (1972, their fig. 5) map 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) show 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 modern 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 2000 m isobath between latitudes 31° and 40° N. (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 and c) and rift-stage crust becomes abundant within 100 or 200 km of the East Coast Magnetic High (Klitgord and Behrendt, 1979). Thus, when the Atlantic opened, normal faults formed as much as about 250 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).

By analogy with normal faults formed on those and other passive continental margins, most Iapetan normal faults in eastern North America may be expected to strike northeast to north-northeast, particularly if they have not formed by reactivation of still older faults. 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 are properly oriented to have been reactivated in later compressional episodes (see following subsections), 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; later chapter of this paper on "State of Stress..."), 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 structural systems that acted to extend the continental lithosphere, and so should have dimensions comparable to the thickness of at least the brittle upper part of the crust.

Summary

As we will conclude at the end of this section, of the three types of Paleozoic and Mesozoic basement faults that reasonably could have formed under the Giles County locale and be responsible for much of its present seismicity, we consider Iapetan normal faults(s) 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 or 200 km of the Iapetan continental edge that is inferred to underlie the steep eastward rise in the unfiltered Bouguer anomaly field (fig. 16). The Giles County seismogenic 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. 15). 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 seismogenic zone.

Alleghany Thrust-load Faults

This subsection considers late Paleozoic faults of a type here named thrust-load (defined below) as candidate sources of Giles County seismicity. The likelihood that a thrust-load fault is responsible for the Giles County seismogenic 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 inferred from stratigraphic and structural observations.

Thrust-load faults are hypothesized to form just in front of recently emplaced thrust sheets, as the crust fractures under their weight in a brittle analogue 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, 1981, and oral and written communs., 1981).

A Testable Deduction from the Thrust-Load Hypothesis

Giles County and its N. 43° E.-striking seismogenic zone lie in the western part of the Valley and Ridge province of the southern Appalachians, which are characterized by east-northeast trending detached structures (about N. 70° E.: fig. 18). The locale lies near the juncture with the central Appalachians, whose detached structures trend north-northeast (about N. 30° E.). Thrust faults and detached 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 seismogenic zone are regarded as having been emplaced during the Alleghany orogeny at the end of the Paleozoic. Older detached structures are known or possible, especially farther southeast, and the detached structures now in Giles County may have begun to form and move before Alleghany times. However, such earlier events would not affect the conclusion that the detached near-surface rocks and structures of the Giles County locale probably arrived above the seismogenic zone in Pennsylvanian or Permian time.

If the seismogenic 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 reactivatable form. If thrust loading reactivated an Iapetan normal fault, the basement beneath the thrust masses might have to extend horizontally, in the direction perpendicular to the strike of the reactivated normal fault. Fleitout and Froidevaux (1982, p. 43) suggest a theoretical mechanism by which gravitational loading could produce horizontal extension. On the other hand, if the hypothesized thrust-load fault formed as a fresh Alleghany fracture, its strike should follow approximately the strike of the causative load gradient that was imposed by the emplaced thrust masses. The strike of the load gradient should follow the strikes of the causative thrust complexes. That is because depth to basement, stratigraphic levels exposed today by erosion and sedimentary facies all change gradually along strike, but do not change abruptly along strike in the region surrounding Giles County (Colton, 1970; King and Beikman, 1974).

The strike of the seismogenic zone is an unambiguous central Appalachian orientation, rather than a southern Appalachian one (fig. 18), so the hypothesized thrust-load fault would have been caused by emplacement of central Appalachian thrust sheets. However, the detached structures now exposed in and near Giles County have southern Appalachian orientations (fig. 18). Those southern Appalachian detached 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 seismogenic 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 or buried by them. Thus, the thrust-load hypothesis leads to the deduction that, in the Giles County locale, arrival of central Appalachian detached masses predates that of southern Appalachian detached masses.

That 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 all favor southern Appalachian thrust masses as having reached the vicinity of Giles County slightly before those of central Appalachian orientations. This contradiction of the relative ages deduced from the thrust-load hypothesis negates that hypothesis.

Stratigraphic Tests

The stratigraphic arguments are the strongest, although even they are not conclusive. The straightforward stratigraphic approach of determining the ages of youngest folded and oldest unfolded rocks cannot work here. Youngest folded rocks are Early Permian in the central Appalachians and Middle or Late Pennsylvanian in the southern Appalachians (King and Beikman, 1974; Van Eysinga, 1975). 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) presents 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 19. 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 Alleghany thrust sheets. At least the portions 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 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) infer 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 that hinterland. 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.

That deposition occurred in two overlapping basins separated by a wide, diffuse hinge line (fig. 20; Arkle, 1969, 1972; Horne and others, 1978). Late Mississippian (Englund and others, 1982) and Early 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. Early Pennsylvanian rocks of the Pottsville Group (figs. 19 and 20) north of the hinge line are much thinner than are correlative rocks south of it. 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. Williams and Bragonier (1974) document 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.

Thus, the sedimentary record of Alleghany tectonism indicates an older, Early and Middle Pennsylvanian (Pocahontas, New River, Kanawha, and Charleston) age in the southern Appalachians near Giles County, but a younger, Middle and Late Pennsylvanian (Allegheny, Conemaugh and Monongahela) age in the central Appalachians. That conclusion finds further stratigraphic support. First, in eastern Pennsylvania, the Lackawana syncline is at least partly a central Appalachian fold, and contains tightly folded coal measures

of Middle and Late Pennsylvanian age, all deformed by early Alleghany structures (Wood and Bergin, 1970). Thus, in that portion of the central Appalachians that is occupied by the Lackawana syncline, Alleghany deformation occurred during or after Late Pennsylvanian time.

Second, Horne and others (1978) and Cavorac and others (1964) report abrupt southward thickening of lowest Pennsylvanian (lower Pottsville) strata across the east-striking Paint Creek-Irvine fault system of eastern Kentucky and southern West Virginia. Moreover, the southeastward tilting of depositional surfaces south of the hinge line occurred in Early Pennsylvanian time. It is possible that that tilting was a crustal response to loading by advancing southern Appalachian thrust masses, and that the activity on the Paint Creek-Irvine faults 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 masses, and later, northeast-trending depression in front of advancing central Appalachian thrust sheets, would be consistent with observations of Kulander and others (1980). In the outcrop belt of Mississippian rocks that is shown in figure 18, they mapped extension fractures that strike north-northeast and east-northeast. The fractures formed later than middle Mississippian time because they occur in limestone of that age. The 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. Dean, oral commun., 1981). Thus, the evidence cited in this paragraph can be interpreted to indicate emplacement of southern Appalachian thrust masses earlier in Pennsylvanian time than emplacement of central Appalachian thrust masses.

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 post-depositional processes such as burial, tectonism and metamorphism. Accordingly, one would expect contours on rank and fixed carbon to follow Appalachian structural trends, and they do (Figure 2 of Babu and others, 1973). Values of both variables increase to the south-southeast in southeastern West Virginia, adjacent to the east-northeast-trending southern Appalachians, and 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 can record trends of detached 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 on ash and sulfur content have crude southern Appalachian trends in the southern coal field, where most coals are of Early Pennsylvanian age, and 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, pertinent data on coal composition and age are consistent with the suggestion that paleotopography that was created by growth of detached 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 result is that from the Mississippian Greenbrier Limestone in both the central and southern Appalachians, along a strike belt about 120 miles (190 km) long roughly centered on the circled numeral 1 in Figure 18 (Dean and Kulander, 1977 and 1978; Dean and others, 1979; Skinner, 1979; S. Dean, oral commun., 1981). There, stylolites formed on preexisting systematic joints, and both were folded by folds visible at scales of 1:24,000 to 1:250,000. Stylolite teeth parallel the causal greatest compressive stress, and these have slickenlines (grooves or striae on slickensided surfaces: Fleuty, 1975) parallel to the teeth. Over a large area extending about 50 miles (80 km) into the central Appalachians, teeth trend north-northwest and so are a southern Appalachian structure. In the southwestern portion of the Williamsburg anticline (locality 1, fig. 18; a central Appalachian anticline), the stylolites of southern Appalachian orientation 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. Dean, oral commun., 1981).

However, interpretations of individual larger interfering structures at several localities 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 (1973, localities 2-4 of fig. 18 of this paper). At locality 2, Perry concluded that Bick's mapping records a thrust fault of southern Appalachian orientation that has been folded by an anticline of central Appalachian orientation. Bick agrees (written communs., 1978 and 1981). At localities 3 and 4 (fig. 18) Perry interpreted Bick's maps to show other central and southern Appalachian structures interfering with each other. However, subsequent mapping indicates that the inferred interference does not or may not occur at those two localities (K. Bick, written commun., 1981). More recently, near locality 4 (fig. 18) Bick (1982) and Henika and others (1982) both interpreted an interference structure involving the Purgatory Mountain anticline and the Pulaski thrust sheet, and deduced different relative ages using different additional data. In another study in the area of the junction of 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 (fig. 18, locality 5). Olson concludes (p. 88) 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 Olson's interpretation (R. C. McDowell, 1981, and 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 portions of the central Appalachians and adjacent portions of the southern Appalachians is known to have been long and complex on both map and outcrop scales (Geiser, 1977 and 1981; Perry and de Witt, 1977, p. 39-40; Perry, 1978; Bartholomew, 1979; Nickelsen, 1979 and 1980; Van der Voo, 1979b; 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). Dahlstrom (1970) documents type examples in the foothills of the Canadian Rockies where the typical cratonward-younging sequence of relative ages of detached structures is locally reversed. Roeder and others (1978) and Witherspoon and Roeder (1981) interpret 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 Plateau provinces, it seems likely that many more local map relations will be needed before relative ages of structures become clear. Particular attention will have to be given to the size order (Nickelsen, 1963) of the structures involved, before the relative ages of central and southern Appalachian thrusting can be clearly deduced from such evidence alone. That is probable if, as seems likely from the stratigraphic evidence cited above in this subsection, the two fold and thrust provinces overlapped in time for part of their growth. Such overlap can be inferred with particular clarity from the paleogeographic maps of Donaldson and Shumaker (1981). Mapping underway by workers such as K. Bick, S. Dean, B. Kulander, and R. McDowell (oral and written communs., 1978-1982) will eventually produce the clearest and most detailed determinations of relative ages. However, now and for our purposes, the chronology of folded stylolites of Dean and Kulander, and conclusions drawn from the stratigraphic data mapped by Arkle and compiled by Donaldson and Shumaker, probably give more reliable relative ages because both approaches average results over large areas.

Summary

In summary, the thrust-load hypothesis leads to the deduction that central Appalachian detached structures entered or formed in Giles County before those of the southern Appalachians. That conclusion is contradicted by relative ages inferred from stratigraphic and outcrop-scale structural data; interference relations of map-scale structures are inconclusive. We conclude that the Giles County seismogenic zone did not form originally as a fresh Alleghany fracture or fracture zone under thrust-loading. However, recall that that line of reasoning assumes that such a fresh break would parallel the causative thrust front. Thrust-loading by southern Appalachian detached masses 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 of the two preceding subsections of this paper.

Atlantic Normal Faults

This subsection considers Mesozoic normal faults as candidate sources of Giles County seismicity. Such faults formed in Pangean continental crust as the Atlantic Ocean began to open. They bound Mesozoic grabens and half grabens that are exposed at least from Massachusetts to South Carolina (for instance, see Figure 16 and King and Beikman, 1974). They are known or inferred to bound Mesozoic basins detected geophysically and by drilling under younger sediments and sedimentary rocks as far east as the edge of the continental shelf (see compilation of Wentworth and Mergner-Keefer, 1981a, b, c and references cited there and in this subsection of this paper). The faults are high-angle, with net normal slip. Most strikes are in the range 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, some such Mesozoic faults appear seismically active (Aggarwal and Sykes, 1978), although Ratcliffe (1981a, b, c) questions that association.

Atlantic normal faults 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 basins bounded by faults such as described above are unknown as far northwest as the study locale (fig. 16; this has also been noted by Bollinger, 1981a). Second, large faults with normal slip that cut Alleghany detached structures are also unknown that far northwest. A Mesozoic normal fault of small enough extent to have escaped detection thus far would probably be too small to generate a damaging earthquake (for example, consider Figure 14), even if erosion had removed all evidence of any Mesozoic sediment that might have accumulated on the down-dropped block. Third, as noted above in the subsection on Iapetan normal faults, most Mesozoic faults and basins in the eastern United States are confined to lie on or east of the steep eastward rise in the Bouguer anomaly field. In that subsection, we suggested that Mesozoic extensional faulting was restricted to weaker, more heterogeneous crust east of the gravity rise, and should not be expected in relatively intact North American cratonic crust inferred to lie west of the rise. The gravity rise is about 50 miles (80 km) southeast of Giles County, and the nearest known Mesozoic basin is the fault-bounded Dan River Basin, about 70 miles (110 km) southeast of Giles County. Fourth, it is possible, although structurally and stratigraphically improbable, that the very last movements on Alleghany thrust faults occurred in earliest Mesozoic time, and buried very early Mesozoic or Permian basins with bounding faults beneath the Giles County locale. However, the sedimentary fillings of Mesozoic basins of the Atlantic seaboard of the southeastern United States have compressional velocities from 4.4 km/sec or less, to 4.85 km/sec, although higher velocities are possible with admixtures of basalt (Stewart and others, 1973; Daniels, 1974; Ackermann, 1977; Talwani, 1977; Behrendt and others, 1981). For the Giles County locale, Bollinger and others (1980) found no compressional velocities less than 5.[illegible]3 km/sec, and the velocity preferred for depths down to 5.[illegible] km there is [illegible].65 km/sec (table 3, this paper).

For all of the preceding 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 (for example, see review by Russ, 1981). However, we know of no data to suggest that either process operated near the Giles County locale. The nearest candidate for either is the graben known as the Rome trough of western Pennsylvania, western West Virginia and eastern Kentucky (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 of King and Zietz (1978) and Zietz and others (1980). The magnetic lineament is about 60 miles (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 detachment 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). Arguments against such a hypothesis for the Giles County locale are based on information presented in this and preceding sections. First, all well-determined hypocentral depths in the locale lie in pre-Appalachian metamorphic and igneous basement, below the deepest known Appalachian detachments. Second, a deeper Appalachian detachment, in basement, such as those found farther south and southeast and in other mountain ranges, is unknown in or near Giles County. The detachment in the shales of the Lower Cambrian Rome Formation, on which rocks of the Valley and Ridge province rode northwestward, would form part of the cratonward toe of any intrabasement detachment that might lie to the southeast, and thus could not be underlain by such an intrabasement detachment. Third, the Giles County seismogenic zone itself dips too steeply over too great a depth range, and is too tabular, to be part of a detachment complex.

There is other seismicity in the Giles County locale that is not a part of the Giles County seismogenic zone itself. It is possible that some of that other seismicity could occur on a pre-Appalachian detachment within the basement, below any Appalachian detachments. Presumably such a deeper detachment would be of Grenville 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

The Giles County seismogenic zone probably occurs by reactivation of an Iapetan normal fault, for reasons summarized at the end of a preceding subsection on such faults. The stratigraphic and structural arguments against

a seismogenic thrust-load fault are not conclusive, but neither is the existence of such faults as freshly-formed crustal fractures. 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 the ocean whose closing produces the thrusting. For the Giles County locale, the Pocahontas Basin may be such a foredeep and any such reactivated faults would be Iapetan normal faults. Finally, Mesozoic normal faults are still less likely candidates for the source of Giles County seismicity because they appear to be confined well to the southeast of the Giles County locale.

STATE OF STRESS IN THE GILES COUNTY, VIRGINIA, LOCALE

Introduction

This section uses selected stress measurements to estimate the orientation of the greatest horizontal compressive stress operating today on the Giles County seismogenic zone. We then argue that the estimated orientation is not inconsistent with the orientation and sense of present-day seismic slip on the zone, as inferred above from P-wave polarities and hypocentral distributions.

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 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 quality of stress orientations and their applicability to seismogenic depths are more important for our purposes here than is the number of sites at which the orientation is measured, we did not investigate Overbey's original sources but 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, and compiled and annotated those that they considered reliable. Two of their orientations meet all of the criteria that we discuss below. Those are measurements from their well OH-1 in southeastern Ohio, and from well 20402 in Lincoln County, West Virginia (fig. 21, table 9). Both measurements were obtained by hydraulic fracturing of wells at depths exceeding 800 m (about 2600 ft).

The other six measurements shown in Figure 21 and Table 9 are from the analyses and compilation by Evans (1979). He examined oriented cores of portions 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 applied to rocks in work funded by the U. S. Department of Energy (Kulander, Dean and Barton, 1977; Kulander, Barton and Dean, 1979). He concluded that most unmineralized, unslickensided fractures in most cores are induced by coring. Of the three types of fractures induced by the coring operation, 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 conclude that the fractures form in advance of the down-cutting drill bit, as extensional fractures in an orientation not distorted by the core or hole and 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 its orientation at all cored depths at which such fractures are observed. By inference, the vertical fractures also track the greatest horizontal compressive stress.

GangaRao and others (1979, p. 686) report 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 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. 6,128). (2) Cores can be hundreds of feet (meters) long, and so their fractures can average out variations in stress orientation between individual beds or groups of beds. (3) Such a 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. However, even cores with few petal-centerline fractures can give stress orientations that are consistent within the core and match orientations from nearby wells (table 9). (4) 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 afterward, it can be examined 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 from a well in southwestern Virginia (Wise County, well no. 20338). Evans (1979) and Wilson and others (1980) examined cored stratigraphy, vertical distributions and orientations of slickensides and slickenlines in that core, as well as regional structural, stratigraphic and drilling information. They concluded that the core bottomed in the section of rock comprising the Pine Mountain thrust fault. The stress orientation of N57°E from that core is consistent with other values obtained nearby (table 9,

fig. 21), but to use a result from rocks known to be detached would violate one of the criteria discussed in the next subsection.

Criteria Used to Select Data

Within about 500 km (300 mi) of Giles County, Zoback and Zoback (1980, their plate 2, and 1981, their fig. 1) show 27 stress-orientation measurements made using various methods. Evans (1979) examined cores from 13 wells. These 40 measurements were reduced to the eight of table 9 and figure 21 by using the following criteria.

First, 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. The arguments of the preceding section indicate that this criterion restricts us to measurements made west of the steep rise in the unfiltered Bouguer anomaly field (fig. 16). East of that rise, we and others suggest that the crust is an assemblage of pieces of various sizes, shapes, compositions, thicknesses and origins. Thus the stress field at seismogenic depths there may be both 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) divide the eastern United States (apart from the Gulf Coast) into two stress provinces. Their Atlantic Coast province is characterized by northwesterly-trending greatest horizontal compressive stress, with the least compressive stress being vertical. From this they infer 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 and c) arrive at the same conclusion. Zoback and Zoback (1980, 1981) assign the middle portion of the United States to a Midcontinent stress province, characterized by northeasterly-trending greatest horizontal compressive stress. From this they infer 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 come from a different stress province than that containing Giles County.

Second, we use only stress orientations measured near Giles County, that is, in the eastern portions of Tennessee, Ohio and Kentucky, in the western portions of Virginia and North Carolina, and anywhere in West Virginia.

Third, we use only stress orientations likely to be representative of the stress field at seismogenic depths under Giles County. This means that measurements from detached rocks in thrust masses are suspect, because such rocks may be mechanically decoupled from underlying undetached rocks. Suggestions and structural data consistent with such decoupling of Paleozoic stresses across thrust faults in the area shown in Figure 21 are given by Wheeler (1980, p. 2173-2174), Werner (1980), and Wilson and others (1980). Seeber and Armbruster (1981) suggest similar decoupling for the modern stress

field across a deeper detachment in Georgia. On the other hand, Zoback and Zoback (1980, p. 6136) point out that modern stress orientations in detached rocks of the Plateau province of western New York State and adjacent Pennsylvania are nearly parallel to those in nearby and underlying undetached rocks. Thus, the question of decoupling remains open and it seems possible that detached rocks could be partly decoupled from underlying basement in some places and not in others.

In the region shown in figure 21, deepest detachments climb stratigraphically to the northwest. Broken lines in the figure locate westernmost structures known to us to have experienced significant detachment. Small amounts of detachment, 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 21. Accordingly, we use only stress measurements made appreciably to the west of rocks likely to be detached, or well below such rocks, or both.

Fourth, we have avoided 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.

Fifth, we use only stress measurements satisfying a variety of selective, but less general considerations. Some cores examined by Evans (1979) contain no petal-centerline fractures. Some have such fractures but they exhibit no strong preferred orientation. In examining one core, early workers did not distinguish natural from core-induced fractures. The core of well 20336 in Martin County, Kentucky, 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 latter orientation because that is most representative of fracture orientations in the bottom portion of the core, below the effects of any near-surface detachment in front of the outcrop of the Pine Mountain thrust fault.

Petal-centerline fractures in the core of well 20402 in Lincoln County, West Virginia, present a problem. 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 21 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 20403, only one mile (1.6 km) away. Therefore, for well 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 20402 are not ubiquitous in the cored interval. The topmost 46 feet (14 m) of the 614 feet (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 20403 and the other wells of figure 21 and table 9. Second, both wells are within the Rome trough (fig. 21). As mentioned in a previous section of this paper, 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 portion of the rock volume cored by well 20402 was and is distorted by past movement of some underlying fault associated with the Rome trough. We suggest that the top portion of the cored volume, and all the volume cored by well 20403, sample the regional and undistorted stress field.

Such basement faults are known and inferred to be nearby, and to have affected structures in the overlying sedimentary sequence. The Warfield fault and its associated Warfield anticline are about 15 miles (24 km) south of wells 20402 and 20403. At the Midway-Extra gas field, about 30 miles (48 km) northeast of wells 20402 and 20403, 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 modeling, if the down-dropped 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 down-dropped block is on the west side of the fault (Kulander and others, 1977). Accordingly, we will use the hydrofracture result (N. 50° E.) rather than that from the petal-centerline fractures (N. 33° E.) for well 20402.

Results

Orientations of greatest horizontal compressive stress that meet all the criteria described in the preceding subsection are listed in table 9 and mapped in figure 21. They span about 250 km (170 mi) along strike and about 170 km (110 mi) across strike. They are in the western Plateau province and adjacent foreland, from about 150 to about 300 km (90 to 190 mi) west to north of Giles County.

Our selection criteria have produced a set of 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. 22a). This median orientation agrees with the east-northeasterly orientations that Zoback and Zoback (1980) find for the eastern portion of their Midcontinent stress province. Zoback and Zoback also suggest (their p. 6136) 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 absence may be attributed to two factors. First, as discussed in the introduction to this section, petal-centerline fractures can provide more valid and accurate estimators of <u>in situ</u> stress below the near-surface zone than can some other methods. Some of the transition zone 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 detached rocks. Thus, much of the transition zone may be caused by partial, local decoupling of detached rocks from underlying rocks. The detached 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 yields the results of figure 22b. The two wells in Lincoln County, West Virginia, are one mile (1.6 km) apart (Evans, 1979, p. 92). The hydrofracturing determination from well 20402 averages with the determination from petal-centerline fractures in well 20403 to give a stress orientation of N. 58° E. The two wells in Mason and Jackson Counties, West Virginia, are about 13 miles (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 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 seismogenic zone (fig. 22b).

In general, the selection criteria that we have used to produce table 9 and figure 21 have not changed the median stress orientation much, but have narrowed the range of individual site orientations considerably. Those results are shown by comparing medians and ranges between the three parts of figure 22. Figure 22c was obtained by including 14 other orientations that did not pass our selection criteria. First, we include 10 of the orientations compiled by Overbey (1976). Those 10 are mostly in the area shown in figure 21, are apparently results of hydrofracturing rather than near-surface strain-relief experiments, 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.

Second, we add three orientations compiled by Zoback and Zoback (their TN-3, OH-2 and WV-4) and one by Evans (his VA-1 from well no. 20338 in Wise County, extreme southwestern Virginia). We deleted these four measurements in our selection because OH-2 is too far north and the other three are known or suspected to have been measured in rocks that are shallow, detached or both.

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

Another encouraging conclusion can be drawn from the consistency of stress orientations over the area shown in figure 21. This is that the Rome trough apparently has little effect on present-day stress orientations, at least on the scale of figure 21. (Recall that we suggest a local effect for

all but the top portion of the interval cored by Lincoln County well 20402.) The Rome trough is a Paleozoic graben that trends east from eastern Kentucky, northeast through western West Virginia, and an unknown distance into western Pennsylvania. Its bounding and internal faults and other structures are complex and are not clearly defined (Harris, 1975, 1978; Shumaker, 1977; Kulander and Dean, 1978b; Ammerman and Keller, 1979). Some of the faults 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). Further, in central West Virginia, one of the southeast border faults of the Rome trough affected stress orientations in Pennsylvanian or later time enough to cause mappable changes in orientations of coal cleat (systematic joints in coal) over an area spanning about 100 km (70 mi) (Kulander and Dean, 1978a; Kulander, Dean and Williams, 1977, 1980).

Of the eight sites shown in figure 21, OH-1 and probably E-P no. 1 and 3 D/K are northwest of the Rome trough. The others are within its limits. Apparently this major, long-lived crustal structure has no discernible effect at the scale of figure 21 on present-day *in situ* stress orientations at depths and sites pertinent to Giles County seismicity.

Locally, basement structure may perturb the regional stress field. Such perturbations have been suggested in the preceding subsection of this paper, and by Overbey (1976) and Abou-Sayed and others (1978), but based only on widely separated results from Wetzel County (near well E-P no. 1 of fig. 21) and Lincoln County (wells 20402 and 20403 of fig. 21). 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. That conclusion encourages us in our extrapolation of the N. 64° E. orientation southeastward to Giles County and mid-crustal depths.

Consistency with Focal Mechanisms

Figure 15 gives a provisional composite focal mechanism solution for well-located microearthquakes that occurred on the Giles County seismogenic zone. As discussed in a previous section, we consider the fault or fault zone that is responsible for the seismogenic zone to strike N. 43° E., and to dip steeply, probably to the northwest. To construct figure 15, the dip was taken as 80° NW.

Figure 23 illustrates an evaluation of the consistency of stress orientations deduced from the composite focal mechanism of figure 15, with those deduced from estimates of *in situ* stress that are shown in figure 21 and table 9. The *in situ* stress estimates are strikes of vertical fractures formed by extension against least horizontal compressive stress, which is also the least compressive principal stress (S_3). If we assume that the estimates can be extrapolated southeastward and downward to the Giles County seismogenic zone, the only constraint they provide is that S_h of figure 23 is parallel to S_3. The greatest and intermediate compressive principal stresses (S_1 and S_2, respectively) are constrained only to lie in the vertical plane perpendicular to S_h (represented by the dash-dot great circle of figure 23).

In an earlier section, we concluded that the seismogenic 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. 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 23. Raleigh and others (1972) suggest a procedure for estimating the probable orientation of S_1 from a focal mechanism. That estimate is S_1' of figure 23. It lies 15° from the P axis in the direction toward the preferred nodal plane. Raleigh and others suggest 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 a small circle in figure 23.

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 23 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 15. Further, a focal mechanism published by Herrmann (1979), for event J of our figure 10, appears to be inconsistent with both of those deduced stress orientations. Herrmann's solution will be discussed below; note that event J occurred only about 25 km northwest of the Giles County seismogenic zone (fig. 10). From these apparent inconsistencies one could conclude that the stress state at seismogenic depths under and near Giles County changes markedly over horizontal distances as short as 25 km. The rest of this subsection contains arguments that 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 seismogenic zone (fig. 15), and at least some aspects of Herrmann's focal mechanism for event J, might all reflect the same stress field.

We will first compare the stress orientations estimated from the in situ measurements and from the Giles County first motions. We argue that they are roughly consistent 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 23 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 22b, 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 15 and 23. 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 arrivals from the seismogenic zone (fig. 15, table 8). If one honors only the five northeast- to southeast-plunging impulsive first motions, one could easily draw a steeper-dipping auxiliary plane that would be little worse than the shallowly-dipping one used here. If the auxiliary nodal plane were to steepen and strike more westerly in figure 23, S_1' and its enclosing small circle would move 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 23, and the two stress estimates would be consistent within the limits suggested by Raleigh and others (1972). Then B would represent the orientation of S_1.

Thus, 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. Recall that the preferred nodal plane strikes northeast and dips steeply northwest (fig. 15). Thus, the seismogenic zone is probably being reactivated in a combination of reverse and right-slip motion. That inferred motion 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 undetached 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 seismogenic 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, West Virginia, earthquake. That shock was located near but northwest of the Giles County seismogenic zone (fig. 21; Herrmann's event 11, our event J of fig. 10). 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 one. Our selected stress orientations (fig. 21), extrapolated southeast to Elgood, are not consistent with Herrmann's solution. Our estimate of the orientation of greatest horizontal compressive stress (N. 64° E./00° NE.) falls near Herrmann's dilatation (T) axis. The most likely orientation of S_1, which we have estimated in this subsection, also falls within Herrmann's T field. 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 23 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.

DISCUSSION AND CONCLUSIONS

The preceding results present the first direct instrumental evidence for a tabular seismogenic zone in Virginia. Additionally, they present the first direct instrumental evidence for a significant and seismically active basement fault or fault zone in the southeastern United States that does not parallel the trend of the host region's known tectonic fabric.

Future Work Needed for Hazard Zoning

Whether these findings are representative or atypical of a larger region than Giles County remains to be determined. In particular, 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 there a single fault or fault zone that is responsible for the Giles County seismogenic zone? Our only evidence for the existence of the seismogenic zone itself is the hypocentral distribution shown in figures 10 through 13. From that distribution, one may infer the existence of a fault or fault zone. For example, figure 14 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. 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 cumulative activity on the seismogenic zone. However, we know of only one detailed search for such rupture and that is still in progress (McDowell, 1982 and his six preceding semi-annual 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) suggests 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 seismogenic zone (5-25 km), and with the best estimate of the depth of the nearby Elgood earthquake of 1969 (table 1; depth 13.6 km: Carts, 1981). Thus, 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 seismogenic zone by repeated activity over millions of years. However, the problem is still only a potential one because it is so poorly defined. For example, the seismogenic 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 seismogenic zone might be only intermittently active. That could arise from very long recurrence intervals. It 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 thereby accumulate to be visible at the surface. Thus, 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 seismogenic 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 seismogenic 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 detached rocks, which might act to concentrate seismic release of stress on the Giles County seismogenic 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. 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 sub-thrust rock across that possible transform fault. Alternatively, Wheeler (1980) and Wheeler and others (1979) describe 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 seismogenic zone is not unique but is a presently active member of a class of similar zones that cover a portion 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 seismogenic zone, only the four that were relocated by J. W. Dewey and D. W. Gordon (written commun., 1980) exceed M = 2 (fig. 11), and reliable location of smaller events is feasible only over a small area (fig. 5).

(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 a fault at specified distances from the rise. That is expectable because such faults may not have a sharp cratonward limit but may instead decrease gradually in slip and 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 because, if the crust east of the gravity rise is weaker than that to the west (as we have suggested in a previous section), 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 and abundances 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) exhibited numerous partly balanced cross sections across the Valley and Ridge and Plateau provinces of the southern Appalachians. The sections were drawn to show numerous basement faults under the thrust masses. By the arguments of this paper, those basement faults are probably Iapetan normal faults.

Questions (2) and (3) posed in the preceding paragraphs of this section deal with the uniqueness and generalizability of Giles County seismicity. One hypothesis that bears on both questions is that of gravitationally induced stresses, which might reactivate Iapetan normal faults underlying the long Bouguer gravity low that flanks the steep eastward rise on the northwest (Woollard and Joesting, 1964; Haworth and others, 1980; fig. 16). The Giles County locale is in that long low, and the rise passes 50 to 100 km southeast of the locale (fig. 16). 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 there 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 reactivate preexisting faults.

The hypothesis of Goodacre and Hasegawa (1980) is attractive as a possible explanation for the occurrence of seismicity in Giles County, because some of the largest, steepest portions of the Bouguer rise in the central and southern Appalachians are near Giles County, and also because the locale is in or near the Bouguer low, where the hypothesis predicts the greatest shear stresses at the depths of Giles County seismicity.

However, the hypothesis needs more detailed testing before being accepted for Giles County for two reasons. First, there are several exceptionally steep portions 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, rather than in one of them as would be expected from a direct application of the results of Goodacre and Hasegawa (1980). Thus, 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; fig. 16). That low lies about along the structural axis of the Appalachian basin, where the sedimentary rocks are thickest, and the map shape of the low approximately follows the map 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 above in this section?

We suggest that the preceding questions and concepts be considered in designing future work on or near the Giles County seismogenic zone. The questions and their eventual answers will be important in zoning for seismic hazard, as well as in understanding the structural evolution of large portions 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 them.

Seismological Considerations

Note that the definition of the seismogenic zone resulted from excellent earthquake locational capability inside the Giles County seismic network. The basis of that excellence is twofold. 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, thereby allowing precise arrival-time determinations. Thus, accurate (S-P) time intervals 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).

Note also that we have implicitly assumed throughout this paper that the seismogenic 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 meizoseismal area was, 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.G.S. 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."

Finally, we mention that there were two principal reasons for demonstrating the range of allowable fault-plane areas, given the hypocenter data set to date (fig. 14). First, it conveys graphically the nature of calculated results when a given level of statistical confidence is used as an error measure for the requisite analyses. Second, in this particular case, it shows that there can be a variation of a full order of magnitude in the implied fault plane area. Such a range of 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 that (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).

Conclusions

The data presented and analyzed herein constitute a detailed instrumental description of an individual seismogenic zone in the southeastern United States. In the judgment of the authors, the evidence presented warrants the following conclusions.

1) A seismogenic zone has been defined in Giles County, Virginia with the following seismological characteristics:

a) Strike - northeast, with present data indicating N. 43° E.; Dip - near vertical; Depth range - from 5 to 25 km;
b) Horizontal Length - 40 km; centered at Pearisburg, Virginia; Horizontal Width - 10 km.

2) That zone also has the following geological characteristics:
   a) Located within the basement and beneath the Appalachian detachments;
   b) Subparallel in strike to the surface and near-surface structures of the central Appalachians to the north but at an angle of some 30° to the detached 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) At this stage, it is impossible to determine if the faulting motion is reverse or normal. High-angle reverse is more likely because the zone probably dips steeply northwest and because the region is probably under easterly-directed 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 by this study 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. 43° E. seismogenic zone has probably resulted from compressional reactivation of a late Precambrian 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 or suggested to produce large earthquakes elsewhere, that is apparently not true for the Giles County seismogenic zone. Neither is it likely for other seismicity with well-determined depths in or near Giles County.

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APPENDIX A

A NOTE ON MICROSEISMIC LEVELS FOR THE VIRGINIA TECH SEISMIC NETWORK

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Abstract

Six hundred amplitude and period measurements were made of the short-period microseismic background levels at the BLA Observatory and at eight Virginia network stations. The overall average level at BLA was 3 nanometers (daytime) and 5 nanometers (nighttime) at frequencies of 0.9-3.1 Hz. At the network sites, the average daytime level was 5 nanometers at 2.3 Hz; during the nighttime, it was 10 nanometers at 2.3 Hertz.

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 Tech 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 physiographic-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 10 different station sites. Film seismograms, using a viewer (1 sec = 10 mm) 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 two-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 two-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 measurement for WWNSS BLA.

The above 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/night time frame.

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APPENDIX B

DETERMINATION OF A DURATION MAGNITUDE RELATIONSHIP FOR THE VIRGINIA TECH SEISMIC NETWORK

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(Modified from Viret (1980))

Introduction

For the Virginia Tech seismic network, magnitudes of local and regional earthquakes are calculated using body wave magnitude equations according to Nuttli (1973) and Bollinger (1979):

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

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

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 Tech 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, page 342).

There are several sources of possible error in the preceding scheme of magnitude determinations. That is, application of the 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 $m_b(Lg)$ formula is based on waves whose periods are within 0.2 sec of 1 sec. Any observed waves whose periods depart from that range carry the potential for large error. Also, the network seismograph's passband (fig. 3) 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 are clipped, 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) estimates that a variation of only about ± 0.1 magnitude unit 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 above empirical observation can be described (for a review, see Whitehurst, 1977, pp. 9-16). Thus, using the above equations 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:

$$M_D = A + B \log (D) \tag{3}$$

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 above relationship. 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

[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).

the end of duration at the time the trace amplitude returns to a predetermined arbitrary peak-to-peak amplitude (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 (Lg)) measured on the Virginia Tech seismic network. The other two sources used durations and magnitudes (M_L or m_b(Lg)) measured on the Phase I (P1) and Phase II (P2) networks used for seismic monitoring at the North Anna site (see Dames and Moore, 1976). The instruments used at that site were similar, and in some cases identical, to those now in use at Virginia Tech. 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 Tech 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 deviation 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 pints:

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

Figure 24 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.

APPENDIX C

VELOCITY MODEL TEST FOR GILES COUNTY LOCALE

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A microearthquake was detected at 04^h UTC, February 18, 1980, by seismograph stations in 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 (TPM1, TPM2) were recently published (Bollinger and others, 1980). The models are:

Model	V_P (km/sec)	Depth(km)	V_P/V_S
GAB	6.24	0.0	1.70 (regional; Bollinger, 1970)
	8.22	45.0	
VPI	5.7	0.0	1.70 (regional; hybrid, unpublished)
	6.24	10.0	1.70
	8.22	45.0	1.70
MCC	6.34	0.0	1.67 (regional; Chapman, 1979)
	8.18	45.0	1.73
TPM1	5.63	0.0	1.64 (local; Moore, 1979)
	6.53	10.0	1.70
	8.18	49.0	1.71
TPM2	5.63	0.0	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 HYPO 71 (Lee and Lahr, 1975). Initial runs were made to eliminate arrival times with large residuals. Next, each model was tried with one or more different compressional to shear velocity (V_P/V_S) ratios. All runs had trial focal depth (TFD) set equal to zero. The results of the eight runs are tabulated:

Model	V_P/V_S	RMS (sec)	ERH (km)	ERZ (km)	Quality
GAB	1.70	.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

Error measure from HYPO71 (Lee and Lahr, 1975):

RMS - Root-mean-square error of the travel-time residuals, in sec

ERH - Standard error of the epicenter, in km

ERZ - Standard error of the focal depth, in km

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

Stability of Focal Depth Estimated with Changes in TFD

The TPM2 velocity model with $V_P/V_S = 1.70$ was used with several trial focal depths (TFD) and the stability of the estimated focal depth observed. TFD's were chosen to be in each layer and near some layer boundaries. Results are as follows:

Trial Focal Depth (km)	Origin Time (0358 +) (sec)	Lat N (37° +) (min)	Long W (80° +) (min)	Focal Depth (km)	RMS (sec)	ERH/ERZ (km/km)
00	55.28	25.68	35.37	13.1	0.21	0.8/1.5
03	55.34	25.72	35.36	12.1	0.20	0.8/1.5
04	55.27	25.70	35.37	13.2	0.21	0.8/1.5
05	55.35	25.63	35.34	12.1	0.20	0.8/1.5
06	55.34	25.57	35.42	12.6	0.20	0.8/1.5
10	55.24	25.62	35.44	13.9	0.20	0.7/0.8
12	55.21	25.60	35.44	14.4	0.19	0.6/0.8
13	55.21	25.56	35.45	14.6	0.20	0.6/0.6
14	55.18	25.73	35.41	14.7	0.19	0.6/1.9
15	55.17	25.74	35.36	15.0	0.19	0.6/0.8
18	55.17	25.66	35.31	14.9	0.19	0.6/1.8
20	55.17	25.71	35.21	14.5	0.19	0.6/0.6
25	55.18	25.78	35.42	14.7	0.19	0.6/1.9
30	55.16	25.75	35.20	14.4	0.19	0.6/0.6

It is seen that regardless of the initial depth estimate, a final focal depth near 14 km depth is obtained. All runs had B-quality solutions, RMS values of 0.20 $\pm$ 0.01 and ERH values of 0.7 $\pm$ 0.1. Runs with TFD near a boundary 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 SEISMOGENIC ZONE

Note that the nearly vertical, tabular seismogenic zone is defined by only eight microearthquake foci. Thus, it is possible that the zone is only an artifact of the small sample size. To test this, one should ask whether one could choose 12 random earthquakes from the hundreds or thousands that have occurred or will occur in the network locale, and have a reasonable probability of obtaining from 8 of the 12, by chance alone, a tabular distribution as well defined as, or still better defined than, the one shown in figures 8 and 12. By convention, if that probability is less than 0.05, we can assume that the nearly vertical tabular zone reflects a real feature, and is not a chance occurrence. Properly chosen statistical tests can answer than question, by providing the desired probability as their significance values, because such tests include the effect of sample size. In particular, tests lose power as sample size decreases. Here, this means that for small samples the tests used below produce valid but conservative results, and may fail to detect associations that are only weakly significant. The tests will not overemphasize the significance of marginal associations. We will examine strike (fig. 8) and dip (fig. 12) of the tabular zone separately.

Our evaluation of the significance of the tabular zone will proceed in two steps. First, we shall test the significances of the epicentral alignment of figure 8 and the hypocentral alignment of figure 12. Each alignment is defined by 8 of the 12 microearthquakes located by the Giles County network in or near Giles County. Second, if those alignments are significant, we shall test whether the eight locations define a zone that is tabular rather than equidimensional or of indeterminate shape. These two steps overlap somewhat,

but both are necessary. That is, the first will demonstrate that the tabular zone exists, and the second will estimate its orientation.

For the first step, we know of no statistical test that is generally appropriate for detecting and evaluating alignments of points in a plane. For example, one of the problems is the difficulty of constructing mathematical expressions of such perceptual concepts as alignment, the maximum allowable gap between points that comprise an alignment, and the effect of the mean areal density of all points in the sample. However, for the present case only, we argue that we can safely ignore such problems of quantifying perceptions. That is because we have shown figures 8 and 12 to many geologists and seismologists and have encountered no objection to our suggestion that if there is a significant tabular zone of microearthquakes, it is the one described in the first paragraph of the subsection on "Analysis of network events..." and shown in figures 8 and 12. It remains only to determine whether we are all wrong together in assuming that that tabular zone does not arise randomly.

We do this with a carefully constructed randomization test (Conover, 1971, p. 357-364; Mosteller and Rourke, 1973, p. 12-15). The descriptive level of significance of the epicentral alignment of figure 8 is expressed as a fraction. The denominator is the number of ways in which eight epicenters may be chosen from the 12 located, or 495. The numerator is the number of those choices that produce an alignment of eight epicenters at least as extreme as that observed in figure 8. By inspection of figure 8, the numerator is clearly 1, so the descriptive significance level is 0.002. Similarly, the hypocentral alignment of figure 12 is also significant at 0.002.

Thus, the alignment is significant rather than a randomly occurring pattern. In the second step of its evaluation, we determine whether the hypocentral concentration is sufficiently elongated in the northeasterly and vertical directions to justify calling it tabular.

We use simple linear regression to estimate the strike of the tabular zone, by fitting a stright line to the eight epicenters shown in figure 8. This operation assumes that the microseismicity occurs on a vertical plane, which is reasonable (figs. 8, 12). Similarly, we estimate dip from the eight network located foci (solid circles) shown in figure 12. For each of the 2 regressions, we calculate and test for significance 2 correlation coefficients, the regular Pearson's product-moment correlation coefficient r and Spearman's rank-correlation coefficient r_s. Simple linear regression and testing r for significance require 10 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, ch. 7 and 8; H. Rauch, R. Lamb, and P. Lentz, oral communs., 1973 throug 1979). All assumptions are partly or wholly valid for our data. In our case, the effects of small departures from assumptions are that the regression lines and values of r remain valid, but that the tests of significance of r may produce significance values that are slightly too large or too small. Therefore, we also calculate r_s and test it for significance, because that does not require the assumptions that cause problems with the significance of r.

In map view (fig. 8), the regression line trends N. 43° E., which is thus the estimated strike of the zone of the eight foci. $r = 0.95$, which has a significance value less than 0.005, and is supported by a significance value for r_s of less than 0.01. Thus the alignment of the eight epicenters is real,

in that there is less than one chance in 100 that it was produced by chance, despite the small sample size. (The strikes of N. 36° E. and N. 37° E. cited by Bollinger and Wheeler (1980a,b), Wheeler and Bollinger (1980), Bollinger (1981a,b) and Hamilton (1981) were calculated before the occurrences of events 58 and 63. The change of strike is not large enough to alter any conclusions.)

In section view (fig. 12), the regression line slopes 59° NW., which is thus the estimated dip of the zone. r_s has a significance value of 0.028 and so is significant. Here, r_s is probably a better guide to the significance of the dip than is r because the calculation of r is heavily influenced by the three extreme foci (events 32, 33 and 58, the shallowest and deepest in figure 12), whereas calculation of r_s weights all eight foci equally.

However, the four central foci actually hint at a steep southeasterly dip (which is, however, not significant). Further, r is only 0.56; its significance value is 0.079, so r is not significant. Thus we conclude that there is no clear association between horizontal (epicentral) distance along the NW-SE section of figure 12, and focal depth. This lack of detected association could arise in one of three ways, all consistent with the signficant epicentral alignment shown in figure 8. (1) The earthquakes could be occurring in a linear or cylindrical source zone oriented horizontally, or the source zone could be (2) tabular but elongate in a NE-SE direction and nearly horizontal, or (3) tabular and nearly vertical. Figure 12 favors the third interpretation. We conclude that the eight foci occur on a nearly vertical tabular zone of unknown dip, but that a steep northwest dip is more likely than a steep dip to the southeast.

APPENDIX E

STATISTICAL TESTS OF THE COMPOSITE FOCAL MECHANISM

In terms of this problem, the binomial test requires that each first motion shown in figure 15 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 bias, 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 figure 15 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 CFMS is nearly horizontal (fig. 15), that motion is predominately dip-slip. As argued in the main text, the movement is more likely to be high-angle reverse than high-angle normal.

Any statistical test produces biased results if the null and alternative hypotheses are designed by first inspecting the data on which the test will be performed. Such 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 biased test may appear to find significance where an unbiased test would not. The standard protection against such bias 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. 8 and 12) and not from inspection of the first motions of figure 15. 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 affect results of the tests.

A binomial test, with $P = 0.5$ and using all 14 first motions, gives a significance value 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, they suggest that there is no

more than one chance in nine or ten 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, that is, with three or fewer inconsistent first motions of all kinds, or one or no inconsistent impulsive first motions.

Figure Captions

Figure

Number

1. Intensity maps for the May 31, 1897, Giles County, Va., earthquake. (A) Law Engineering Testing Company (1975); (B) 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.
2. Virginia Tech seismic network. Individual stations shown by solid circles with 3- or 4-character station codes. Dashed line divides Plateau (on northwest) and Valley and Ridge (on southeast) provinces, along outcrops of Clinchport and St. Clair faults and Allegheny Front (northwest limb of Wills Mountain anticline) (Rodgers, 1970, plate 1A). Solid lines are state boundaries. Shaded area defines Giles County.
3. Magnification curves for the Giles County subnetwork of the Virginia Tech Seismic Network. (a) Visual (pen-and-ink) recorders; (b) Develocorder. Magnifications at film viewer magnification of 20X; (c) Analog tape recorder-playback system. Input level: 2v p-p, galvoamplifiers: 0.2v. Amplifier gains at all stations are 84 dB except for PWV which is at 78 dB. See Table 2 for general network information.
4. Two different, short-period, vertical seismograms for the same microearthquake that occurred near Narrows, Va. (January 28, 1978;

Event #32. Magnitude (M_D) = 1.6; minute marks every 60 mm on original seismograms. Both transducers located on the same pier at Blacksburg, Va. (A) BLA WWSSN: magnification is 50,000 at 1 Hz and 4,500 at 10 Hz. (B) BLA network visual: magnification is 28,000 at 1 Hz and 65,000 at 10 Hz (see Figure 3A). Note the increase in signal-to-noise ratio achieved by the increased magnification of the higher ground frequencies by the network station.

5. Detection and location capability by any 5 stations of the Virginia Tech Seismic Network. (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 1/4 degree latitude and longitude grid, for events detected by five or more stations. Ellipses are not plotted if their semi-major axes are greater than 100 km or if their 95 percent confidence interval on the focal depth is greater than 100 km. After Tarr (1980). Interpolate only between adjacent grid points; do not extrapolate to undefined grid points.

6. Detection and location capability by any 15 stations of the Virginia Tech Seismic Network. (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 1/4 degree latitude and longitude grid, for events detected by 15 or more stations. Ellipses are not plotted if their semi-major axes are greater than 100 km or if their 95 percent confidence interval on the focal depth is greater than 100 km. After Tarr (1980).

Interpolate only between adjacent grid points; do not extrapolate to undefined grid points.

7. Ninety-percent confidence location ellipses, on a 1/2 degree latitude and longitude grid, for magnitudes m_b = 2.0 (upper figure) and m_b = 3.0 (lower figure) events detected by five or more Virginia Tech Seismic Network Stations. Ellipses not plotted if their semi-major axes are greater than 100 km or if their 95 percent confidence interval on focal depth is greater than 100 km. After Tarr (1980). Interpolate only between adjacent grid points; do not extrapolate to undefined grid points.

8. Epicenter (solid circles) map for microearthquakes located with data from the Giles County, Va., subnetwork. Event identification numbers refer to the listing given in table 6. Sixty-eight percent confidence-ellipsoid axes plotted at each epicenter (Lahr, 1980). Network seismic stations shown by open triangles with 3-letter codes. Inset map shows area of this figure (shaded portion) and locations for the Narrows seismic station (NAV; open triangle) and Richmond (R).

9. Actual locations of Blasts A, B and C shown by stars. Computed locations shown by solid circles with 68 percent confidence ellipses. Location of Narrows, Va., seismic station (NAV) shown by open triangle with center dot symbol.

10. Epicenter map of figure 8 with addition of the JHD relocated epicenters of J. W. Dewey and D. W. Gordon (written commun., 1980). They are shown by open circles and 90 percent confidence ellipsoidal axes. Letter designators refer to table 7. The

locations of vertical profiles A-A' (northeasterly-striking) and B-B' (northwesterly-striking) are also indicated on the figure. Inset map same as for figure 8.

11. The same epicenter map as figure 10 with all geography (except location of seismic station NAV, open triangle) and confidence ellipsoidal axes deleted to provide a particularly uncluttered definition of the seismogenic zone. Additionally, the epicenters are scaled according to magnitude in the figure and are separated according to locational authority: Open circles for epicenters according to J. W. Dewey and D. W. Gordon (written commun., 1980), see table 7; solid circles for epicenters according to this study, see table 6. Inset map shows area of this figure (shaded portion) and NAV station (open triangle).

12. Vertical distribution of the hypocenters along a northwest-striking plane B-B' (see figure 10 for location). Solid circles, this paper; open circles, J. W. Dewey and D. W. Gordon (written commun., 1980) with focal-depth control; open stars, J. W. Dewey and D. W. Gordon (written commun., 1980) without focal-depth control, therefore depths shown were arbitrarily fixed during calculations. Left side of the figure shows error ellipse axes and event numbers; in right side of the figure, hypocenter symbols are scaled according to magnitude (<1 to >4) of the individual earthquakes. Event numbers and letters refer to tables 6 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 profiles. The inset map shows the profile location, the NAV station (open triangle) and Richmond (R).

13. Vertical distribution of the hypocenters along a northeast-striking plane A-A' (see figure 10 for location of A-A'). Solid circles, this paper; open circles, J. W. Dewey and D. W. Gordon (written commun., 1980) with focal-depth control; solid stars, 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; 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 6 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, the NAV station (open triangle) and Richmond (R).

14. Examples of hypocenter (solid circles; each with an associated error ellipse) distributions and interpretations of fault plane areas that can be derived by arbitrarily moving the hypocenters of figure 13 to various positions inside their error ellipses: Upper portion - 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. Lower portion - Maximal area (800 km^2, shaded region). Inset map shows profile

location, the NAV station (open triangle) and Richmond (R). Note that events S and D have unknown focal depths (see figures 12 and 13) and so cannot be used here.

15. Provisional composite focal-mechanism solution (lower-hemisphere equal-area plot) for events in the Giles County, Va., seismogenic zone. Symbols: solid circles for definite compressions; plus signs for doubtful compressions; minus signs for doubtful dilatations; P, T for pressure and tension axes at the source, respectively; boxes with X's for nodal plane poles; dashed lines for nodal planes; C, D for quadrants about the source where the P-wave arrivals show compressional (away from source) and dilatational (toward source) first motions, respectively.

16. Map showing gravity rise in exposed portions of central and southern Appalachians near Giles County locale. Broken lines show state boundaries. Vertical ruling shows Giles County. Solid lines interrupted by signed (negative) and unsigned (zero and positive) numerals are isogals, selected and traced from Woollard and Joesting (1964). Contour interval is 20 mgal. Dotted patterns identify areas of Mesozoic and Cenozoic sedimentary rocks: solid line ornamented with dots on its southeast side shows simplified edge of Coastal Plain onlap from the southeast onto Paleozoic and older rocks of the Appalachians; other dotted areas show Mesozoic fillings of extensional basins, mostly bounded by high-angle normal faults. The bounding faults themselves are shown by heavy solid lines. Geology from King and Beikman (1974). For more details of the geology and gravity field over larger areas, compare map of King and

Beikman with that of Woollard and Joesting or see maps of Haworth and others (1980) or of Simpson, Bothner and Godson (1981) and of Simpson and Godson (1981).

17. Map showing positions of eastward gravity rise in wavelength-filtered Bouguer anomaly fields. Light solid lines show state borders and outlines of Washington, D.C. (W) and Giles County (G). Selected isogals show position and approximate form of rise in unfiltered Bouguer field: (1) light solid lines show -50 mgal and -10 mgal isogals, chosen as the most extreme isogals that define the bottom (northwest side) and top (southeast side) of the rise, respectively, along its entire length in Virginia and North Carolina; (2) heavy dashed line shows -30 mgal isogal, chosen because it is halfway between the -50 mgal and -10 mgal isogals in value. Horizontal ruling shows position of 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 on that map. 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 on that map.

Gravity data simplified and traced from unpublished maps supplied by R. Simpson (written commun., 1981), which combine the maps of Simpson, Bothner and Godson (1981) and Simpson and Godson (1981). Refer to those maps for details that cannot be reproduced in this line drawing.

18. Approximate orientations of Giles County seismogenic zone and of central and southern Appalachian detached structures. Thick solid

line trending about N. 43° E. shows approximate center of the seismogenic zone. Broken line shows West Virginia-Virginia border. Orientations of detached southern Appalachian structures are indicated by traces of outcrops of main thrust faults (solid lines with sawteeth on upper plates). Orientations of detached central Appalachian structures are indicated by main folds as outlined by traces of systemic boundaries (P, Pennsylvanian rocks; M, Mississippian; D, Devonian; S, Silurian and older). Geology and structure simplified from compilation of Willden and others (1968). Circled numbers show localities discussed in text.

19. Permian and Pennsylvanian stratigraphy of West Virginia coal fields. Sources: Englund and others (1979) and Arkle (1974); also Berryhill and Swanson (1972) and Cardwell and others (1968), unless there is a conflict with the two newer sources. Nondeposition and erosion followed Charleston and Dunkard deposition. In both coal fields, basal Pennsylvanian strata are conformable on Mississippian beds southeast of the hinge line of figure 20 and unconformable northwest of it. Numbers refer to isopachs of figure 20. Note that "Allegheny Formation" is spelled with an e, whereas "Alleghany orogeny" takes an a (Rodgers, 1970, p. 30).

20. Maps showing distributional patterns of Pennsylvanian units in West Virginia and parts of adjacent states. See figure 19 for stratigraphy.

(a) Broken hachured lines show outcrops of base (single hachures) and top (double hachures) of Pennsylvanian System, greatly simplified. Hachures point inward toward center of late Paleozoic

Dunkard basin. Double broken line shows position of hinge line of Arkle (1969, 1972), separating northern and southern coal fields of central Appalachians and figure 19. Location of hinge line is approximate: Donaldson (1974) gives its width as 25 to 50 miles (50-80 km).

(b) Distributional patterns of Lower and Middle Pennsylvanian units, which entered the southern coal field of the Pocahontas basin from a southeastern source. Heavy lines show isopachs selected from the maps of Arkle (1974). Circled numerals at ends of isopachs refer to units numbered in figure 19 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 figures 20b-20d 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.

(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 figure 20b. Sediment entered the northern coal field of the Dunkard basin mostly from a northeastern source, but with influx from the

southeast in western Pennsylvania (Williams and Bragonier, 1974). That transitional nature between the patterns of figures 20b and 20d is indicated by the northeastward thinning in Pennsylvania.

(d) Distributional patterns of Middle and Late 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.

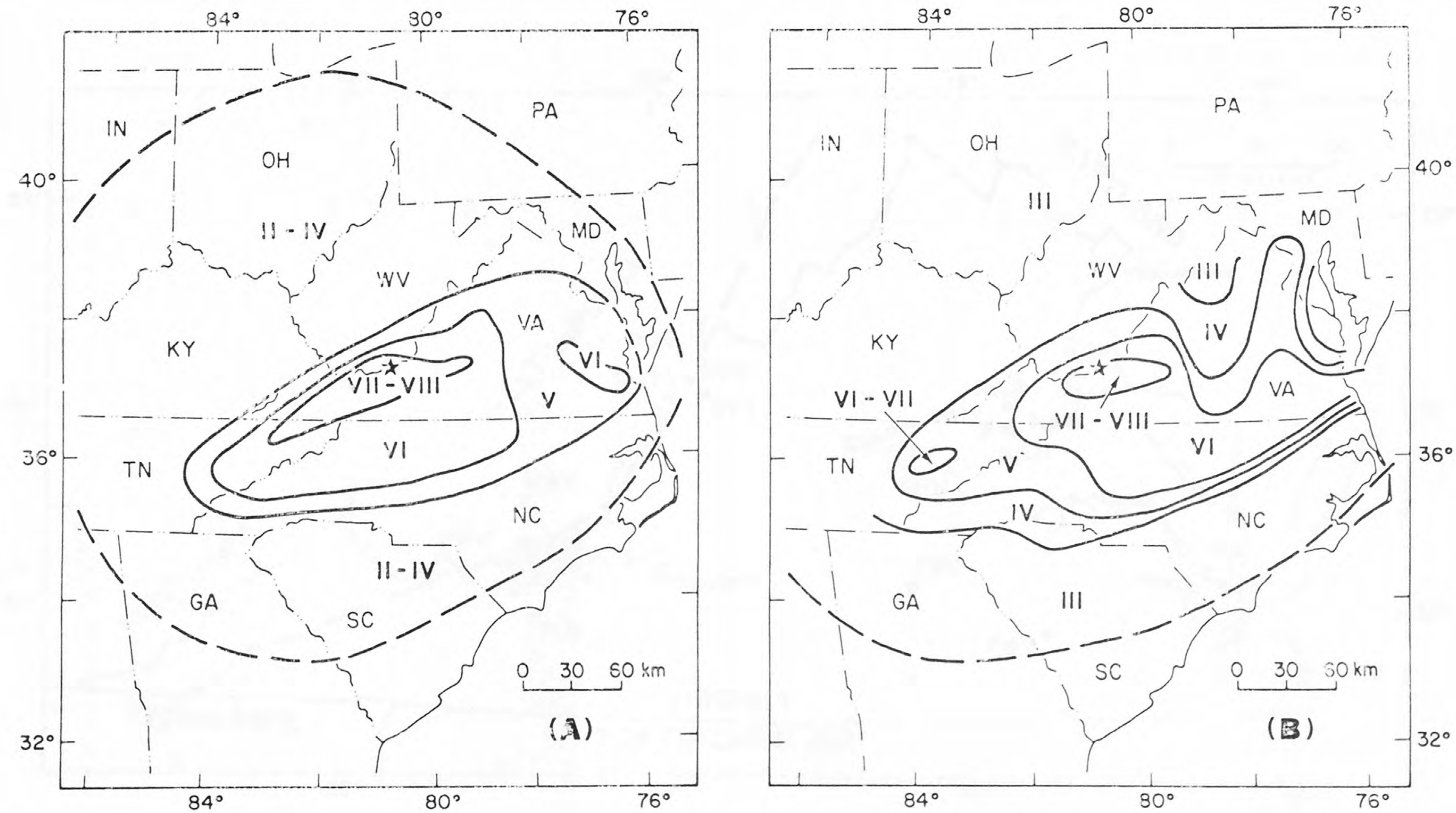
21. Orientations of maximum horizontal compressive stress. Solid lines show state and county boundaries. Lined pattern shows Giles County. Solid circles and lines through them show locations and orientations of selected stress determinations (see text, table 9). Dashed lines show approximate locations of westernmost structures known to us to show significant detachment: C, Chestnut Ridge anticline; B, Burning Springs anticline; M, Mann Mountain anticline; P, outcrop of Pine Mountain thrust fault. E shows approximate location of Elgood earthquake (J of figure 10). 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), Shumaker (1977).

22. 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 figure 22a, 10 from Overbey (1976), 3 from Zoback and Zoback (1980) and 1 from Evans (1979).

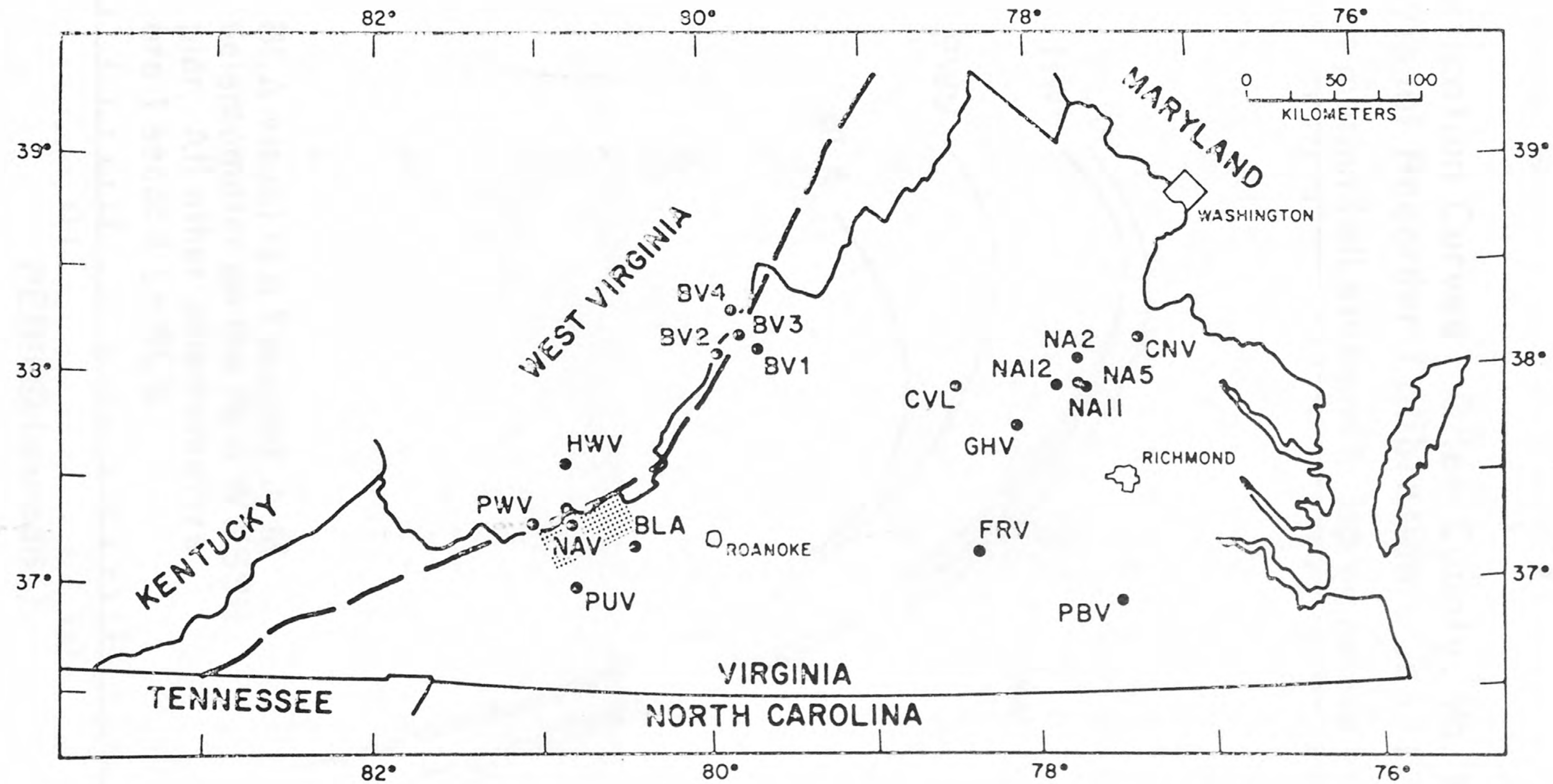
23. Consistency of _in situ_ stress orientation with orientation deduced from composite focal mechanism. Lower-hemisphere equal-area projection. Elements of focal mechanism of figure 15: solid curves show nodal planes, and boxed X's, their poles; F identifies preferred nodal plane, assumed to represent the orientation of the seismogenic fault or fault zone (N. 43° E./80° NW.); P and T locate compressional and tensional axes, respectively, at the seismic source. Elements of _in situ_ stress field of figure 21 and table 9: S_H shows orientation of greatest horizontal compressive stress (N. 64° E./00° NE.); S_h shows orientation of least horizontal compressive stress (N. 26° W./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'. Points A and B are defined in text.

24. Plot of average coda duration versus magnitude for earthquakes recorded by the Virginia Tech Seismic Network.



84°
80°
76°
40°
36°
32°
IN
OH
PA
MD
WV
VA
KY
TN
NC
GA
SC
II - IV
VII - VIII
VI
V
0 30 60 km
(A)
III
IV
VI - VII
(B)

Figure 1



82°
80°
78°
76°
39°
38°
37°
MARYLAND
0 50 100
KILOMETERS
WASHINGTON
WEST VIRGINIA
BV4
BV3
BV2
BV1
NA2
CNV
NA12
NA5
NA11
CVL
GHV
HWV
RICHMOND
PWV
BLA
NAV
ROANOKE
FRV
KENTUCKY
PUV
PBV
VIRGINIA
TENNESSEE
NORTH CAROLINA

Figure 2

Magnification Curves - Giles County, Va. Network
Visual Recorder Calibration - Jan. 1979
Direction of motion (all stations): Up on record = Up on ground

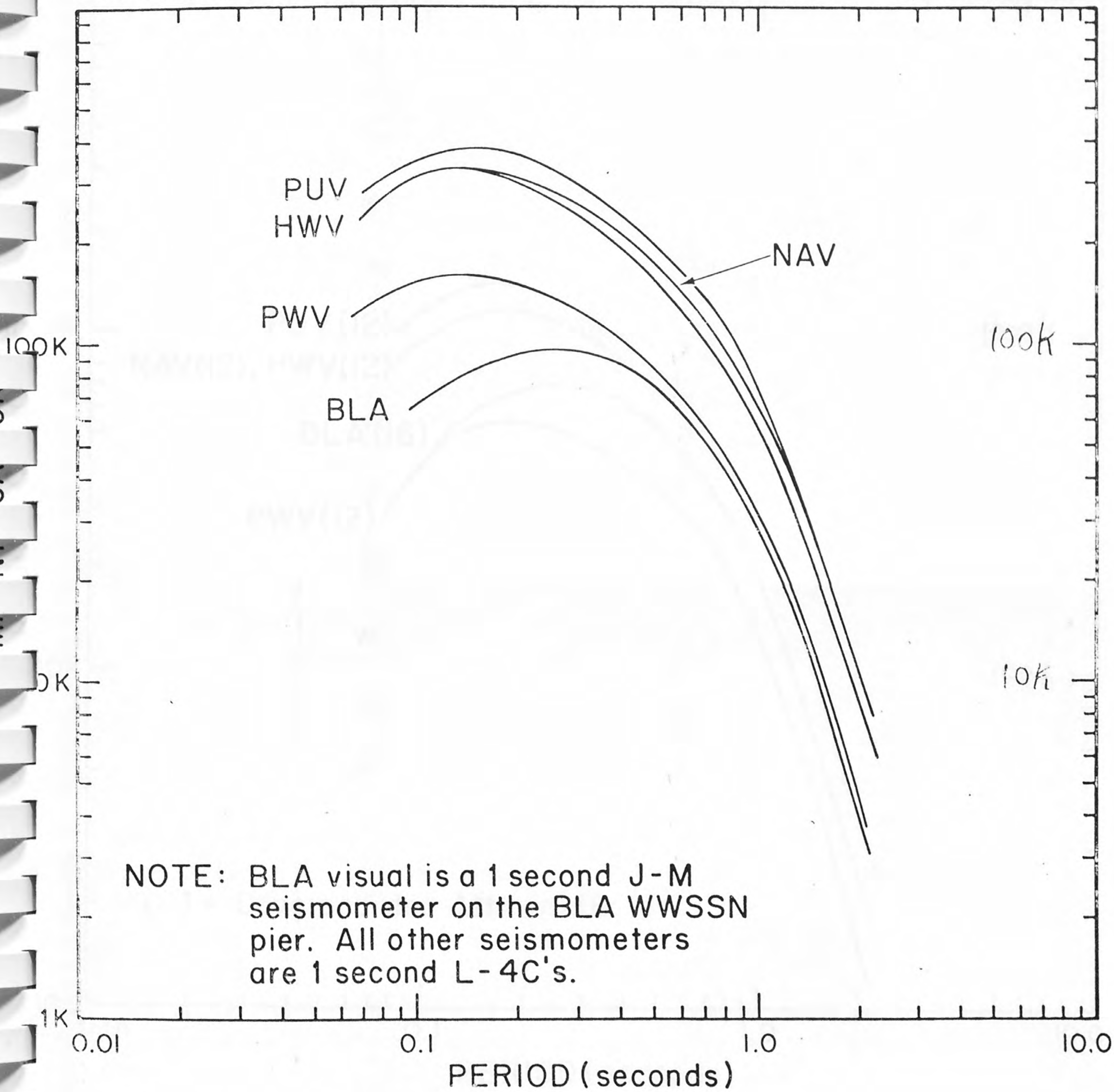


Figure 3a

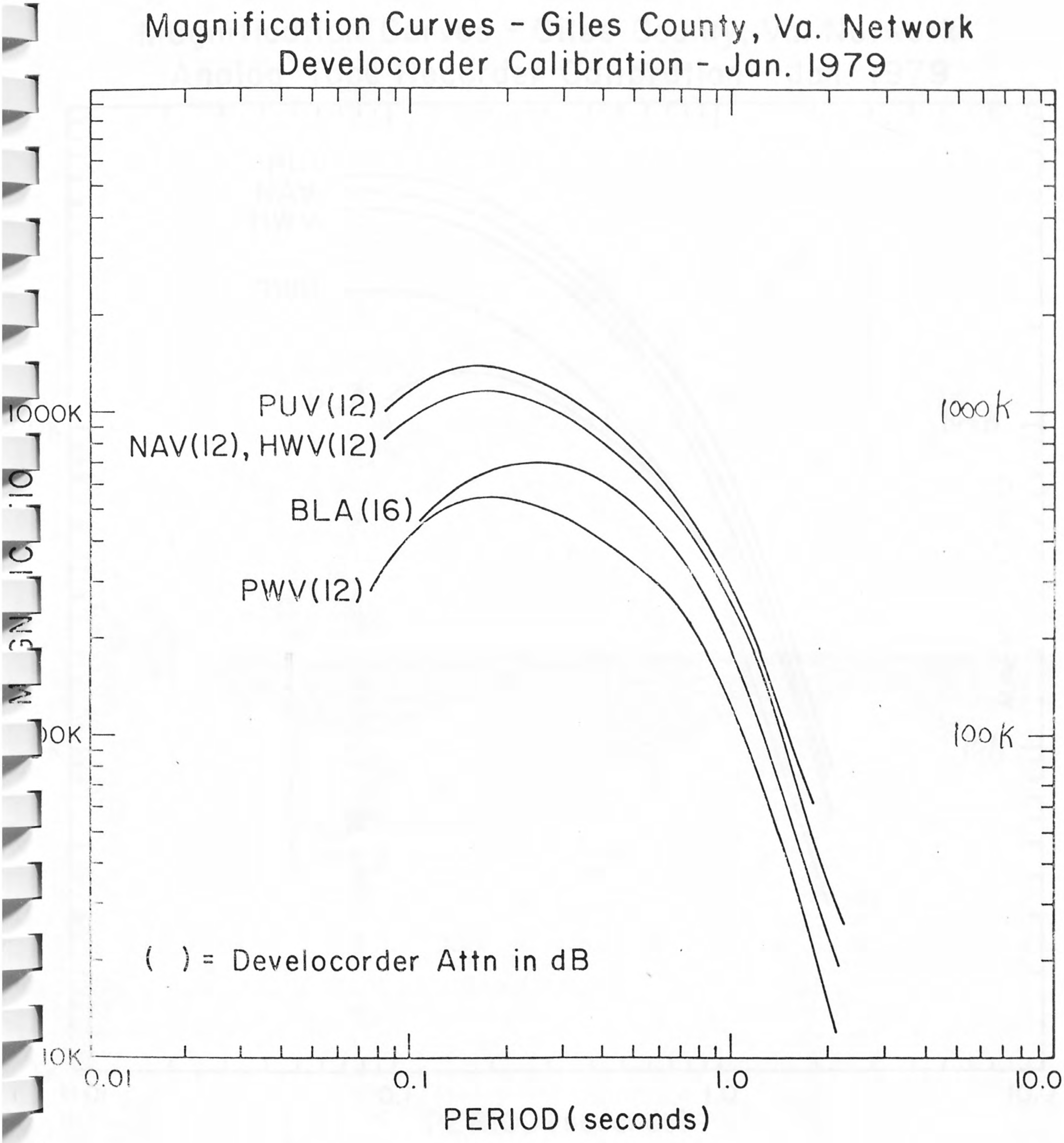

Magnification Curves - Giles County, Va. Network
Developcorder Calibration - Jan. 1979
PUV(12)
NAV(12), HWV(12)
BLA(16)
PWV(12)
1000K
100K
10K
1000 K
100 K
( ) = Develocorder Attn in dB
0.01
0.1
1.0
10.0
PERIOD (seconds)

Figure 3b

Magnification Curves - Giles County, Va. Network
Analog Tape Recorder Calibration - Jan. 1979

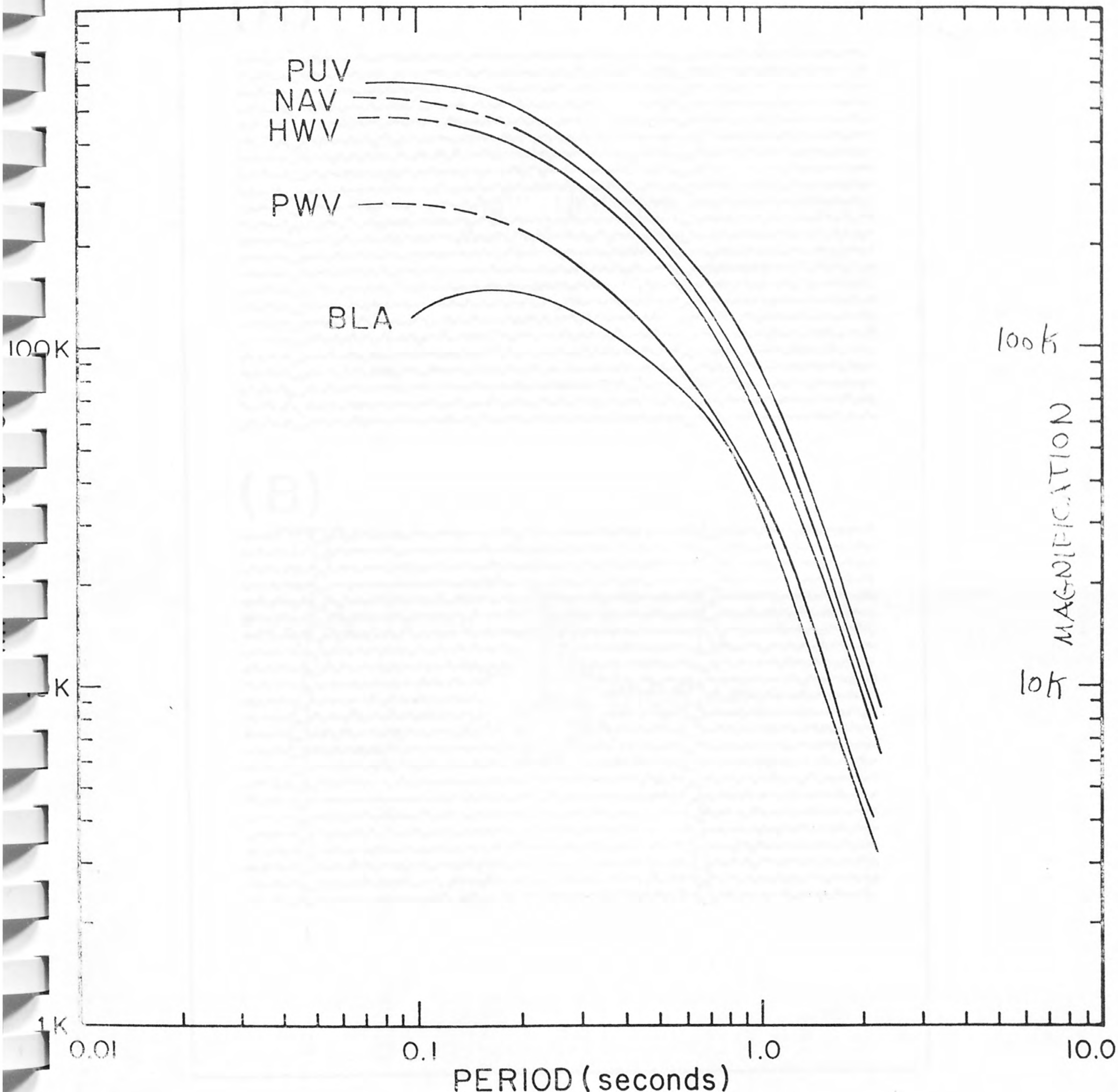


Figure 3c

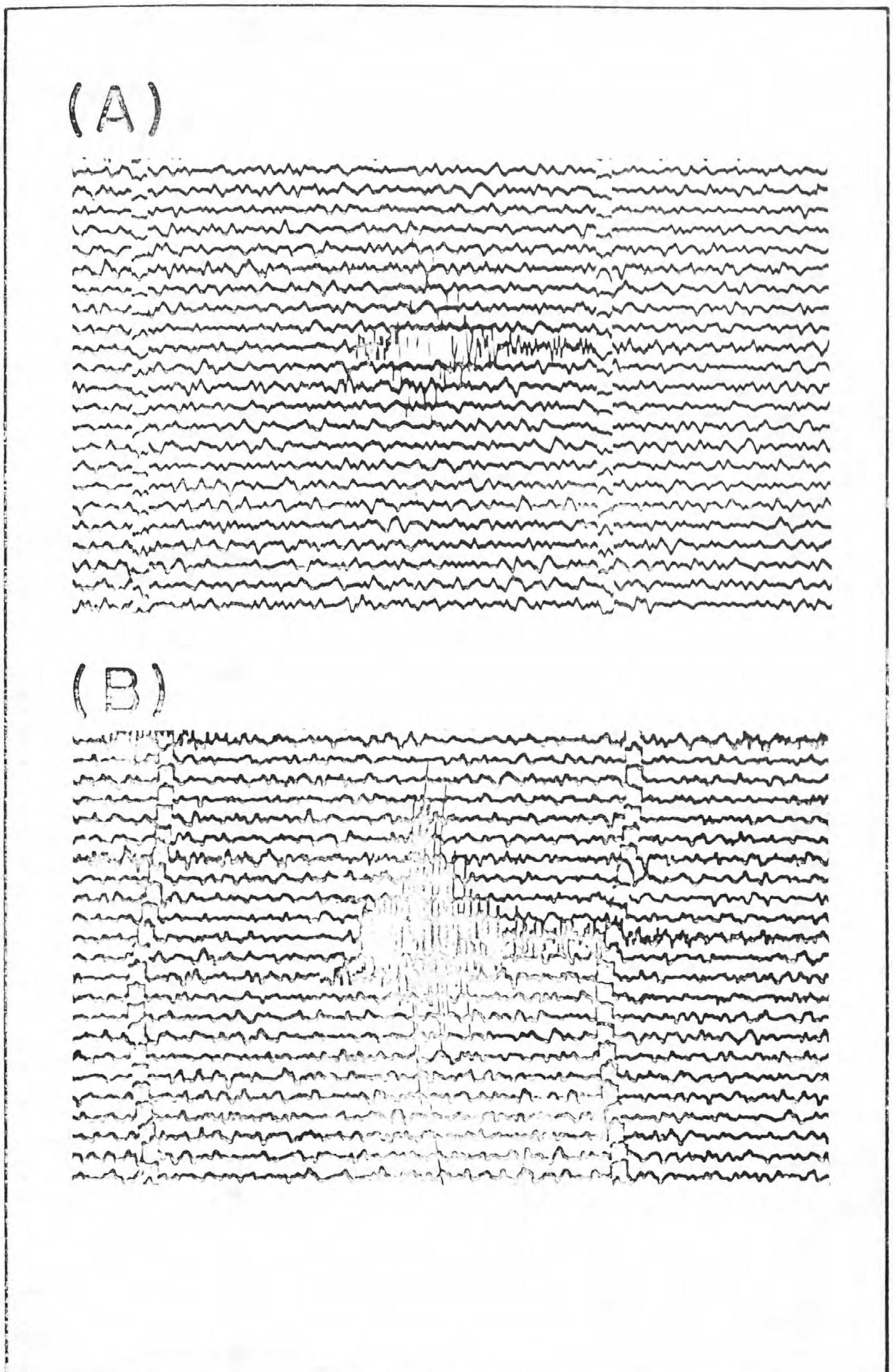


Figure 4

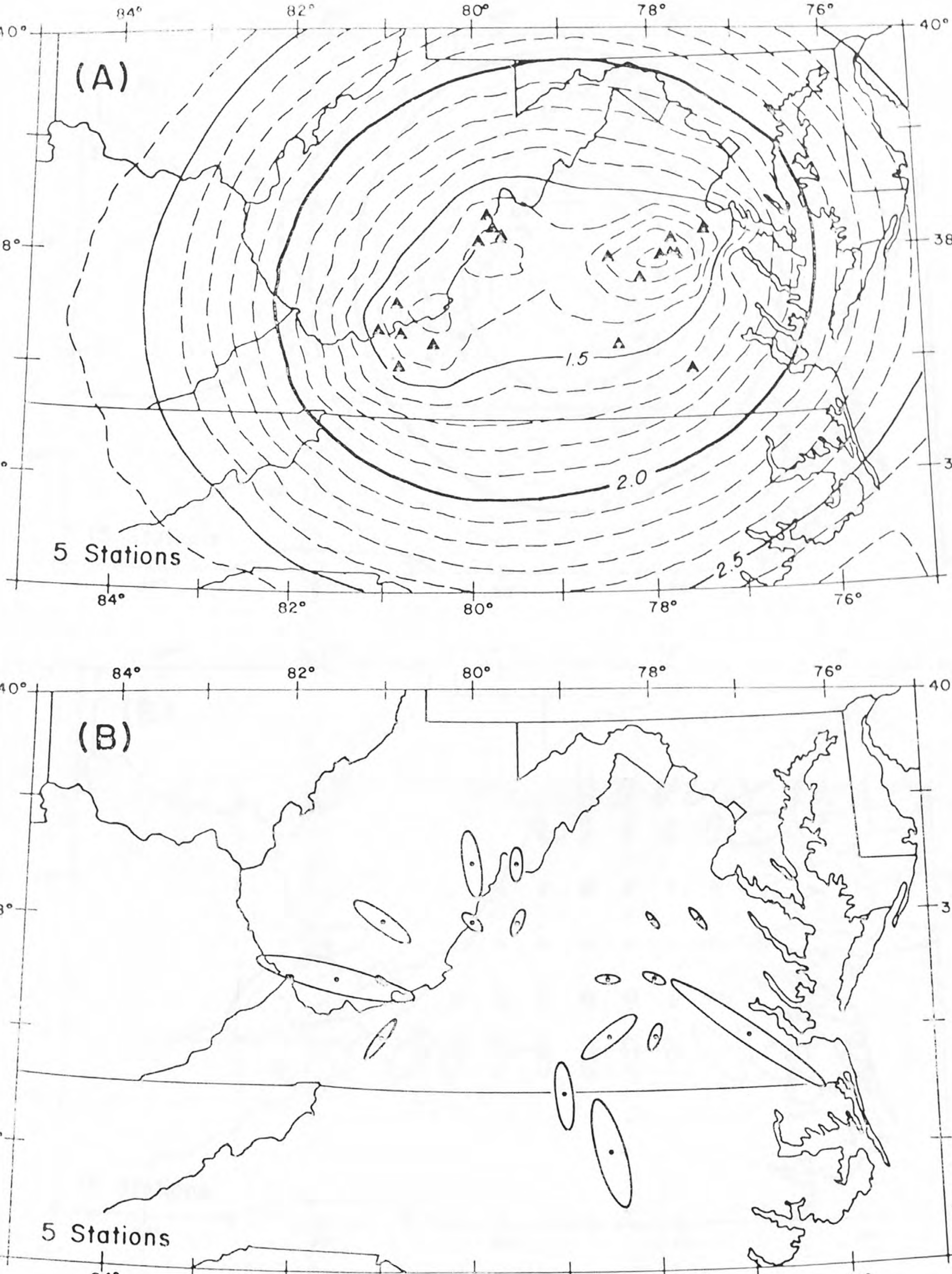


Figure 5

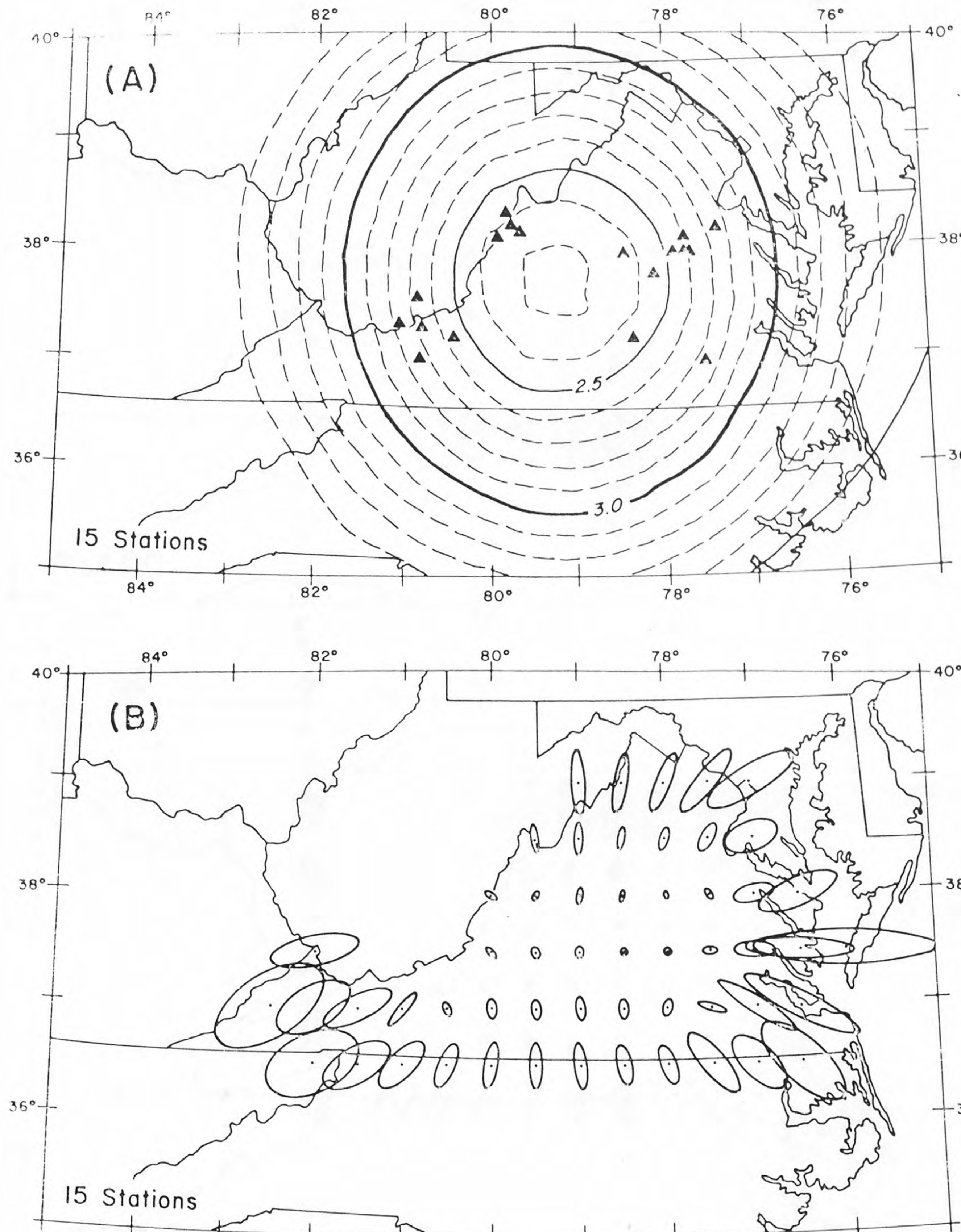
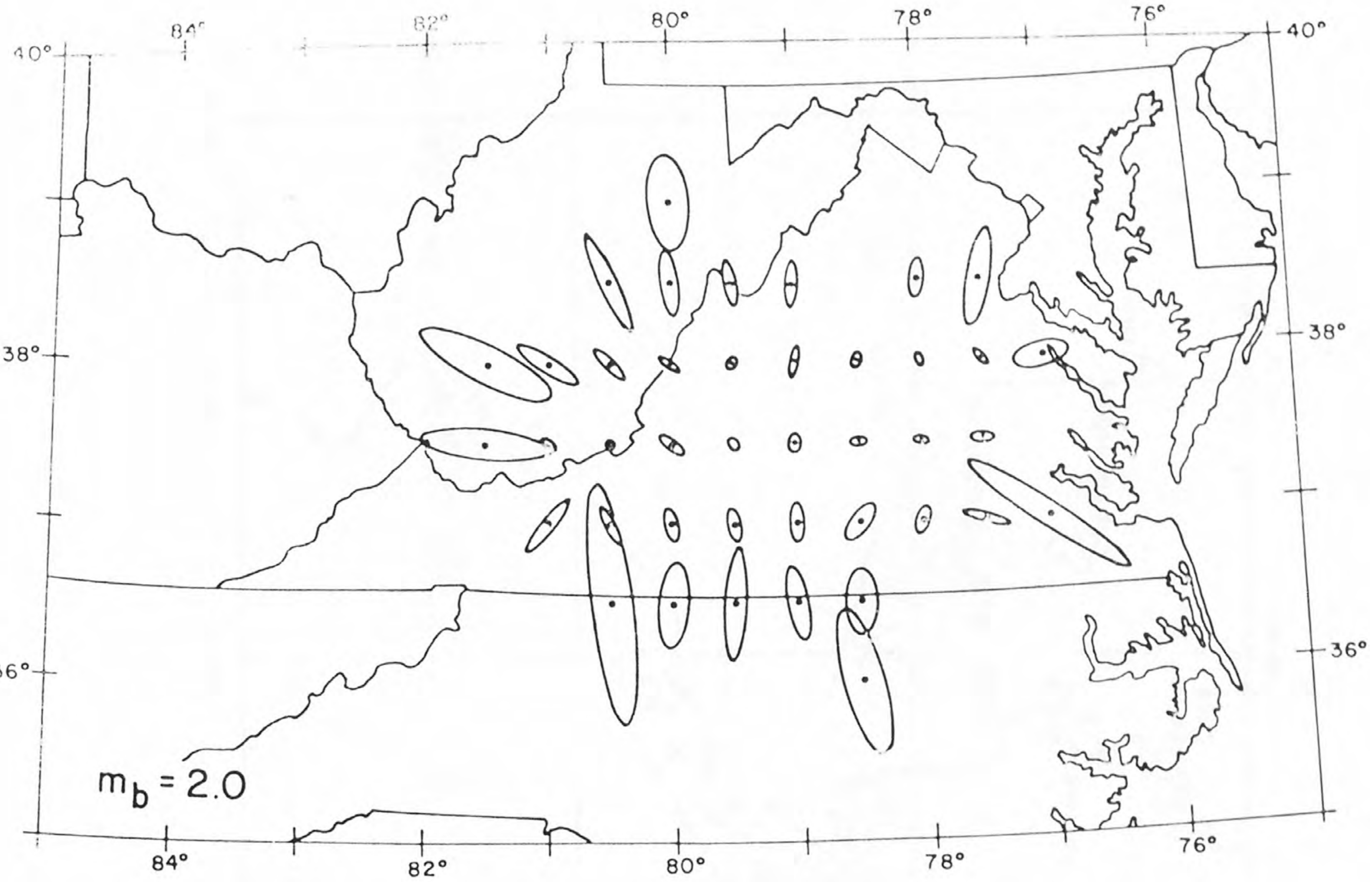


Figure 6



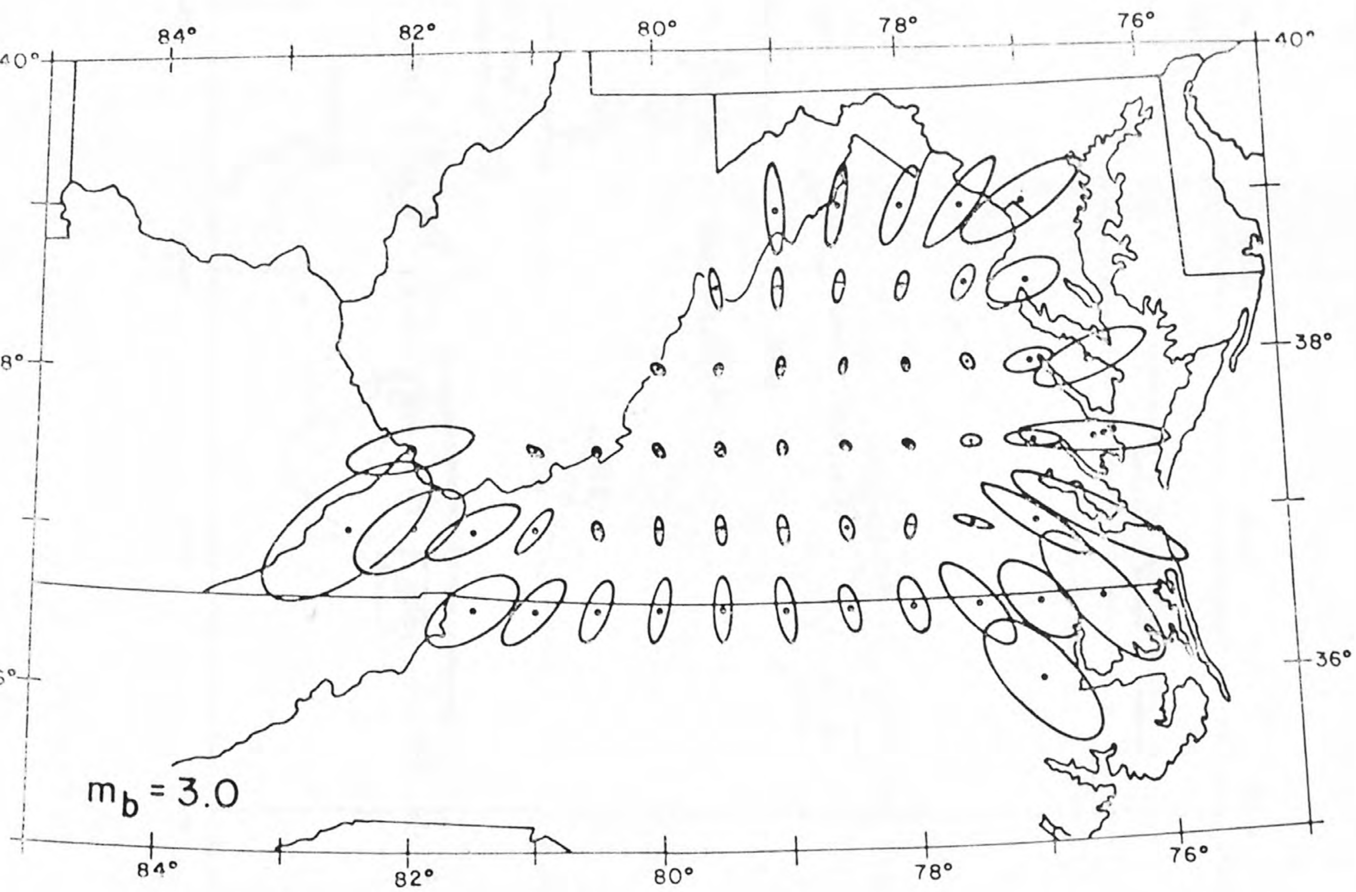


Figure 7

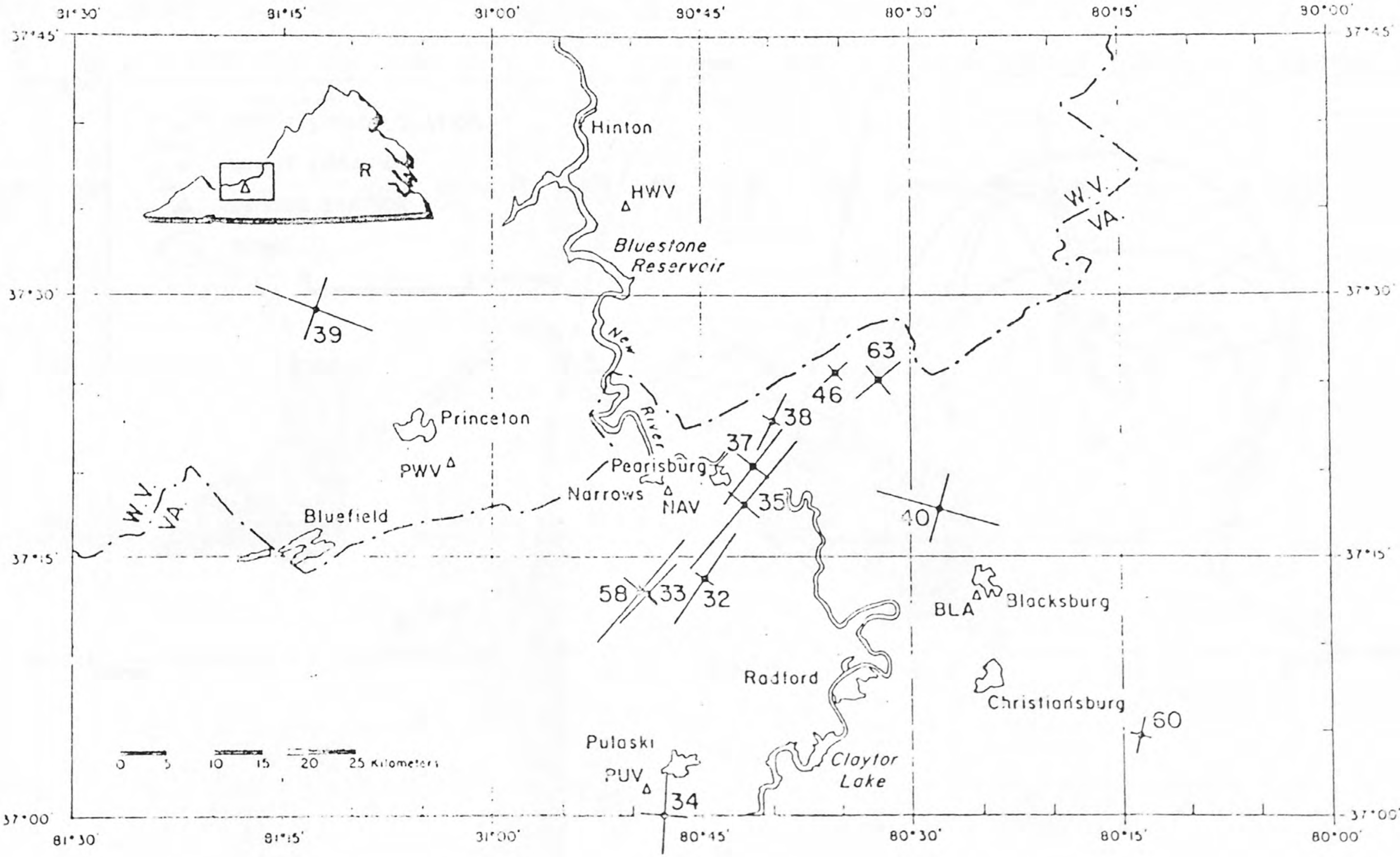

81°30'
81°15'
81°00'
80°45'
80°30'
80°15'
80°00'
37°45'
37°30'
37°15'
37°00'
R
Hinton
HWV
Bluestone
Reservoir
New
River
W.V.
VA
Princeton
PWV
Pearisburg
Narrows
NAV
Bluefield
39
63
46
38
37
35
40
58
33
32
60
34
BLA
Blacksburg
Radford
Christiansburg
Pulaski
PUV
Claytor
Lake
0 5 10 15 20 25 Kilometers

Figure 8.

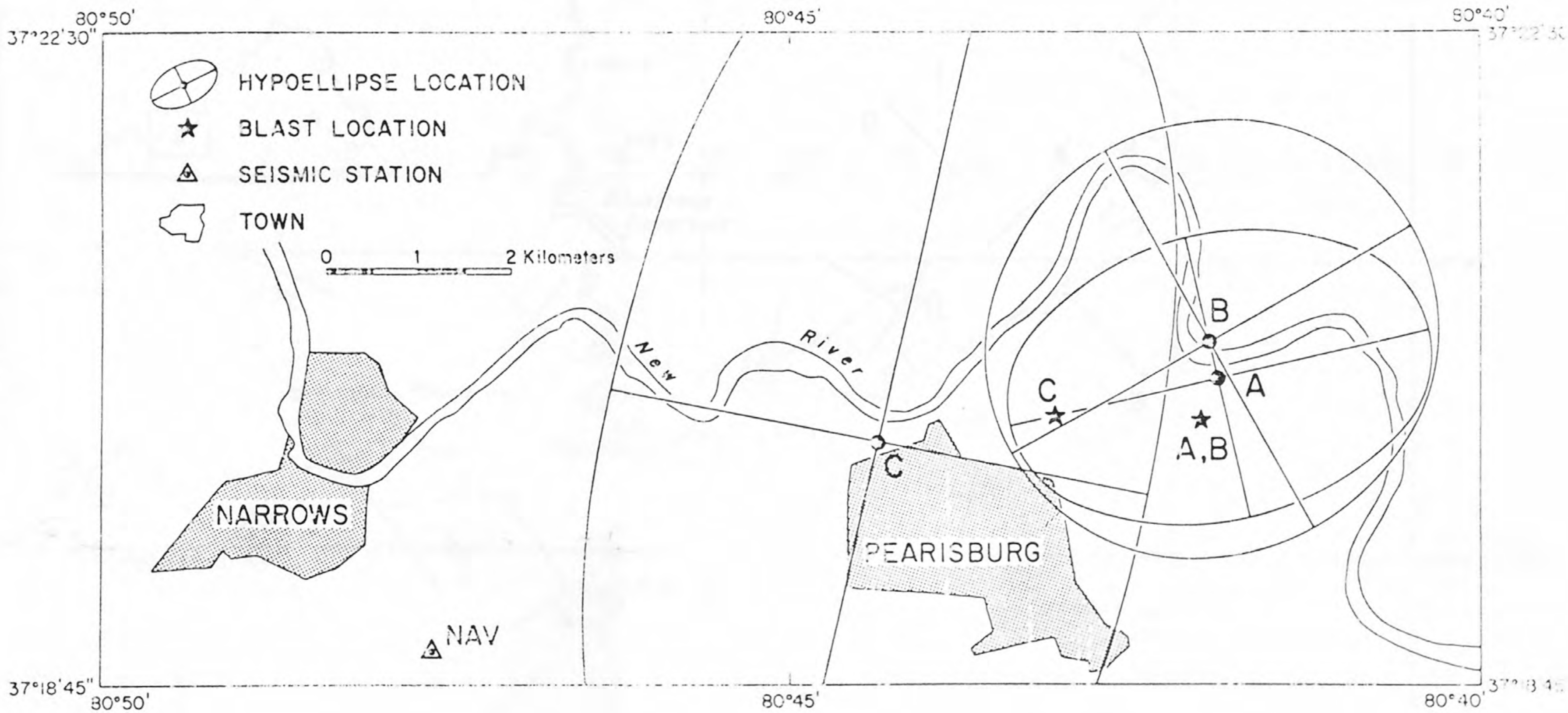
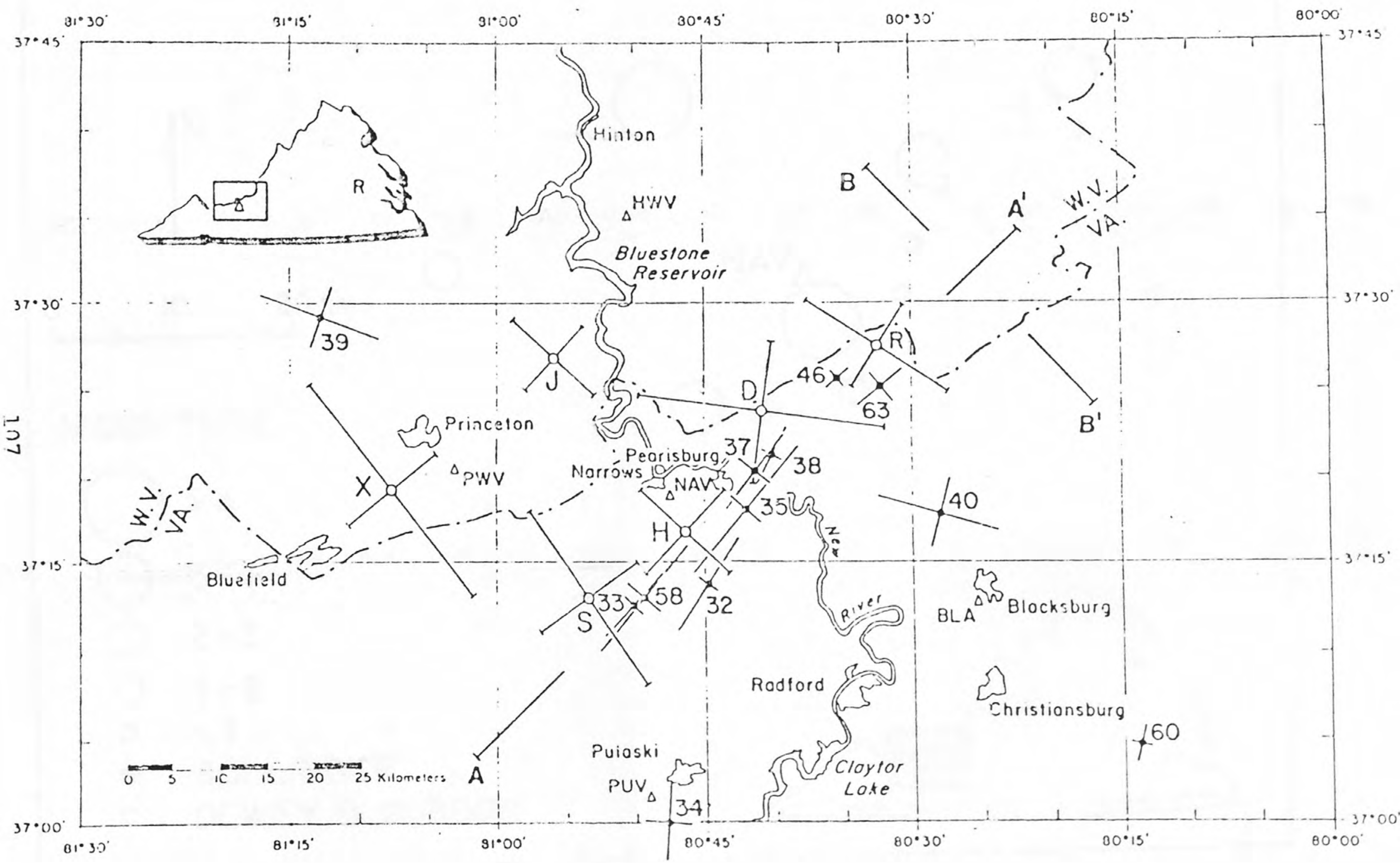

80°50'
37°22'30"
80°45'
80°40'
HYPOELLIPSE LOCATION
BLAST LOCATION
SEISMIC STATION
TOWN
0
1
2 Kilometers
New
River
B
A
C
A,B
NARROWS
PEARISBURG
NAV
37°18'45"
80°50'
80°45'
80°40'

Figure 9



81°30'
81°15'
81°00'
80°45'
80°30'
80°15'
80°00'
37°45'
37°30'
37°15'
37°00'
R
Hinton
HWV
Bluestone
Reservoir
B
A'
W.V.
VA.
B'
39
J
D
46
R
63
Princeton
PWV
Pearisburg
Narrows
NAV
37
38
35
40
X
W.V.
VA.
H
Bluefield
33
58
32
S
River
BLA
Blacksburg
Radford
Christiansburg
60
Pulaski
PUV
34
Claytor
Lake
A
0 5 10 15 20 25 Kilometers

Figure 10.

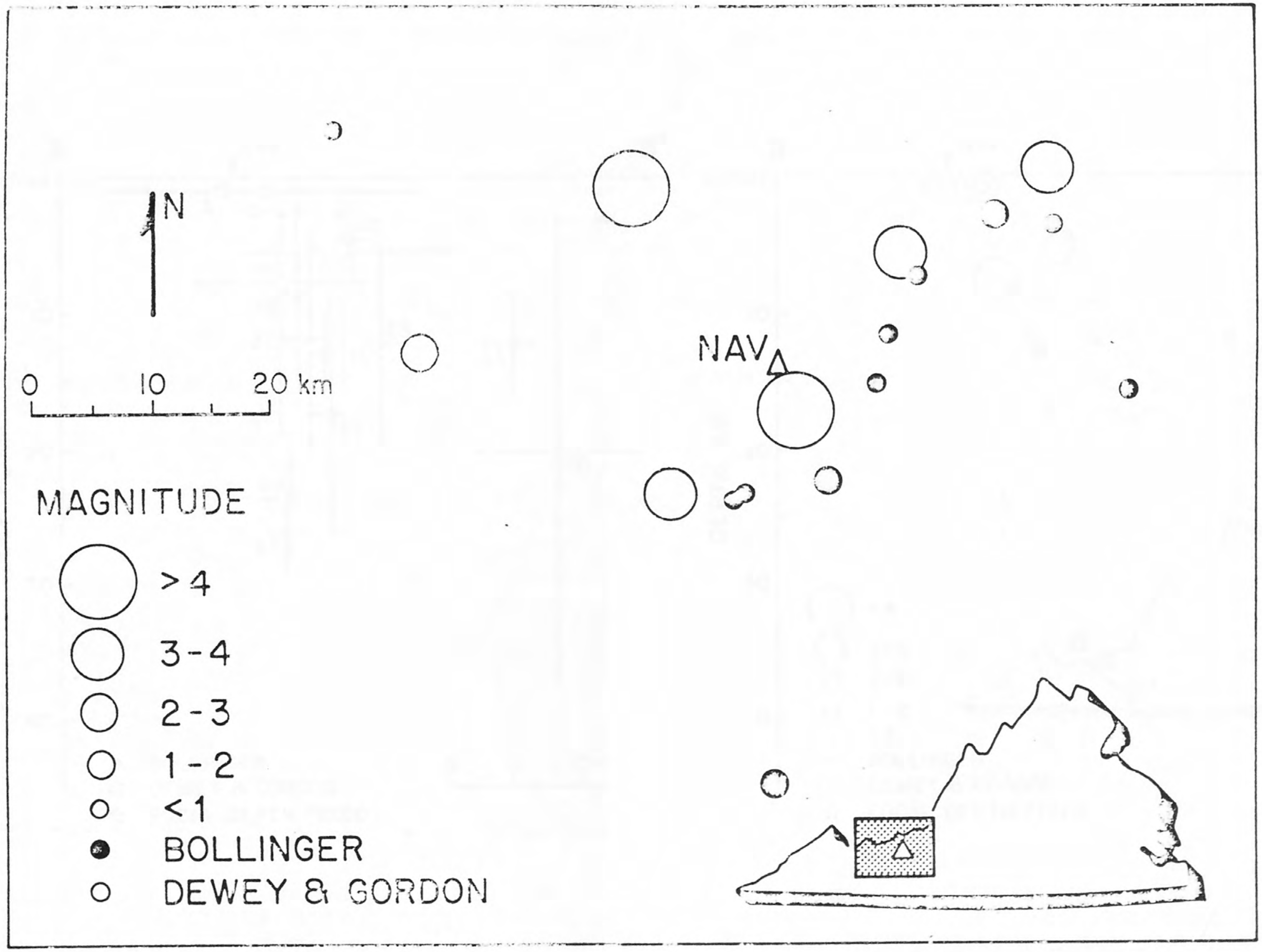


Figure 11

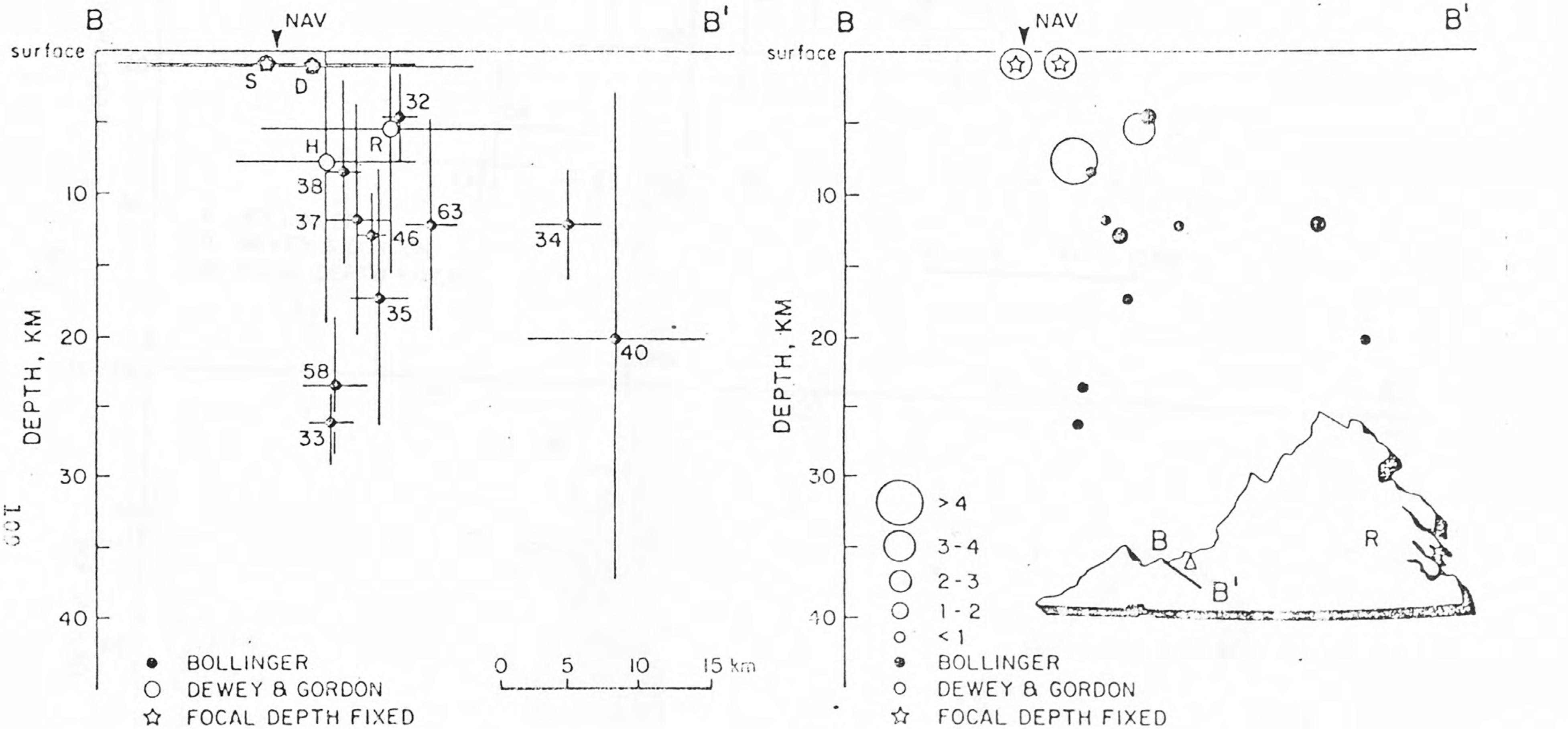
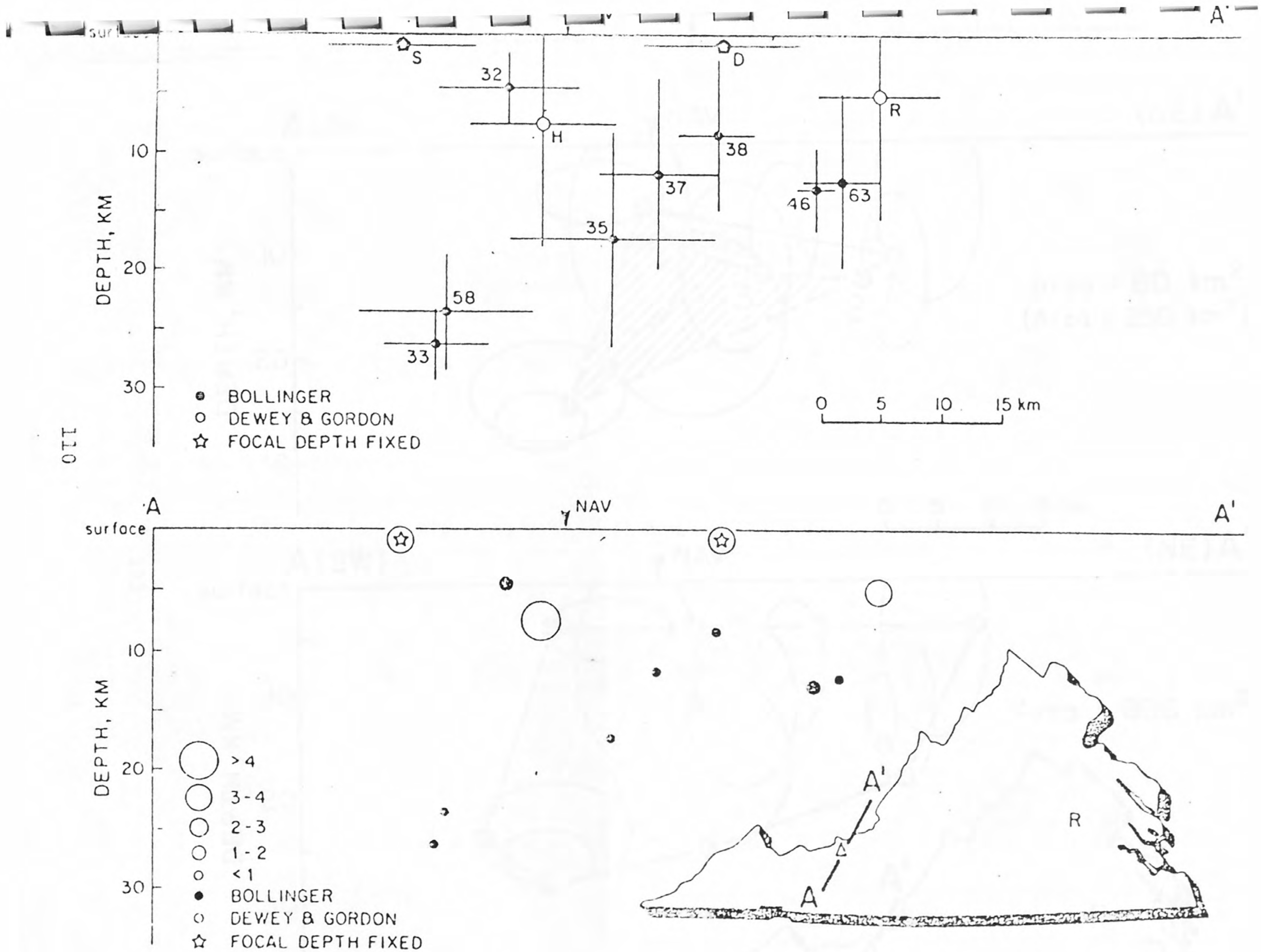

B
NAV
B'
surface
S
D
H
R
32
38
37
46
63
34
35
40
58
33
DEPTH, KM
10
20
30
40
BOLLINGER
DEWEY & GORDON
FOCAL DEPTH FIXED
0
5
10
15 km
B
NAV
B'
surface
DEPTH, KM
10
20
30
40
> 4
3 - 4
2 - 3
1 - 2
< 1
BOLLINGER
DEWEY & GORDON
FOCAL DEPTH FIXED
B
B'
R

Figure 12.



A'
surface
S
D
32
H
R
38
37
35
46
63
58
33
DEPTH, KM
10
20
30
BOLLINGER
DEWEY & GORDON
FOCAL DEPTH FIXED
0 5 10 15 km
A
NAV
A'
surface
DEPTH, KM
10
20
30
>4
3-4
2-3
1-2
<1
BOLLINGER
DEWEY & GORDON
FOCAL DEPTH FIXED
A'
A
R

Figure 13.

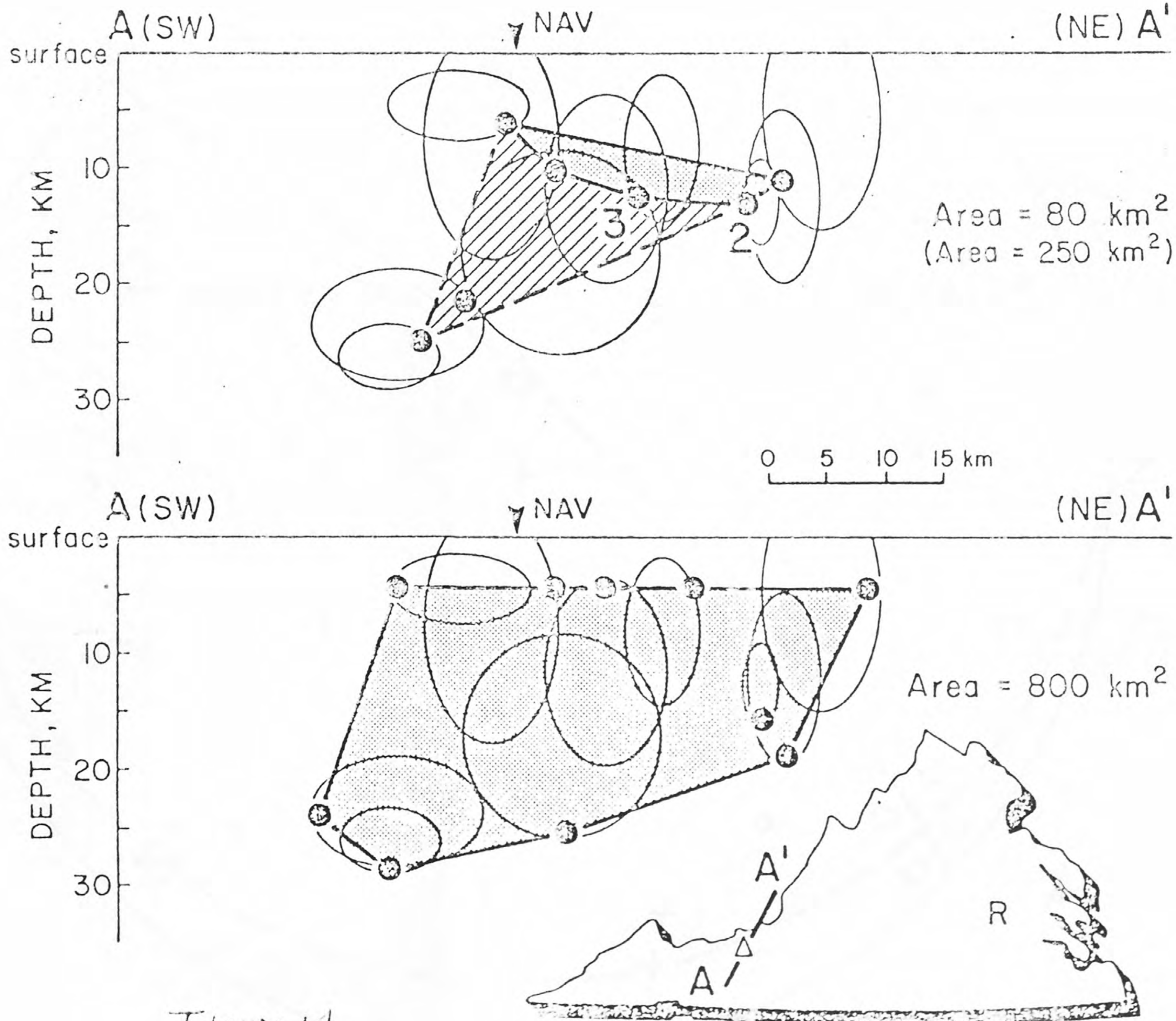

A(SW)
NAV
(NE) A'
surface
DEPTH, KM
10
20
30
3
2
Area = 80 km²
(Area = 250 km²)
0 5 10 15 km
A(SW)
NAV
(NE) A'
surface
DEPTH, KM
10
20
30
Area = 800 km²
A'
A
R

Figure 14

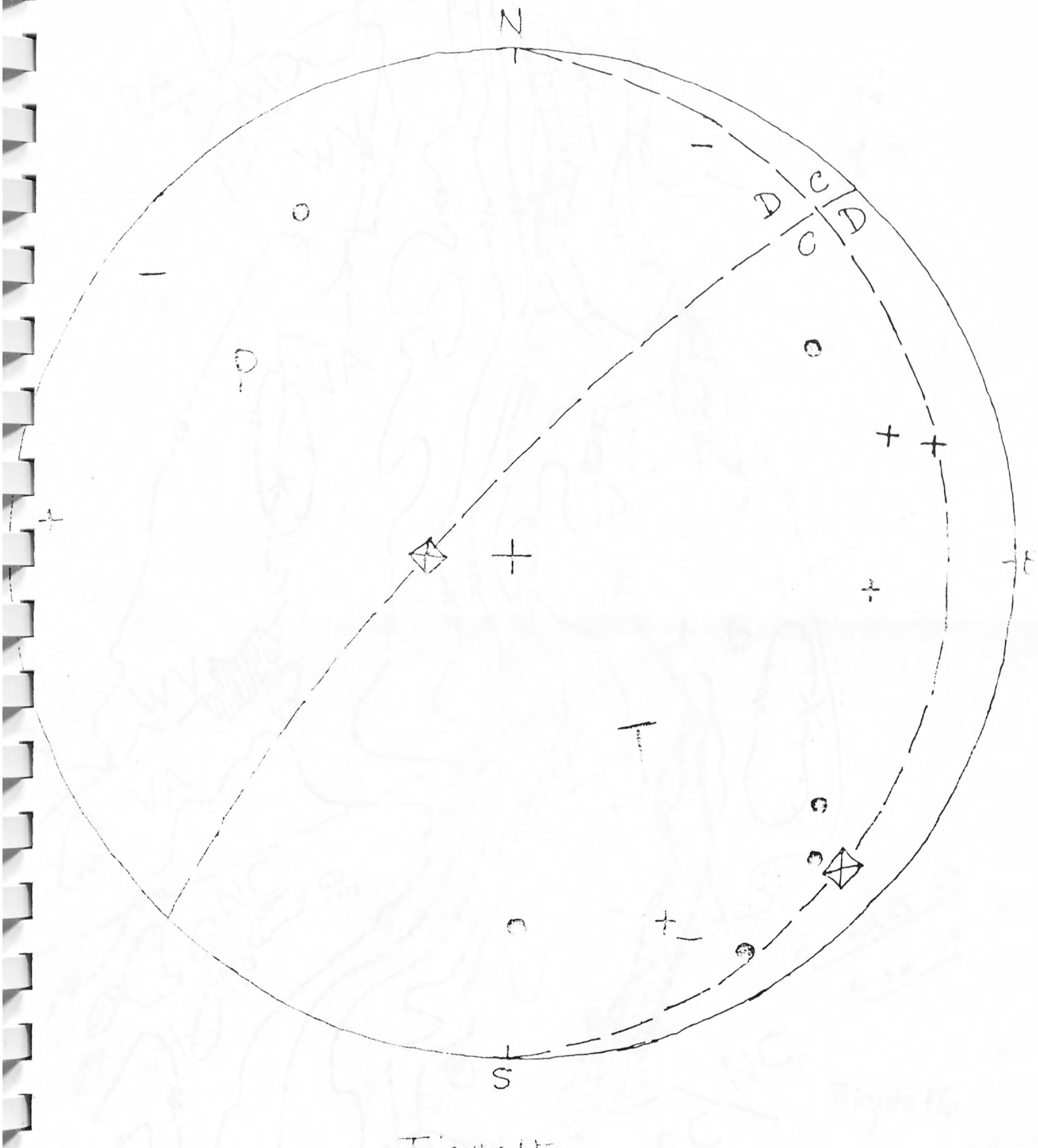


Figure 15

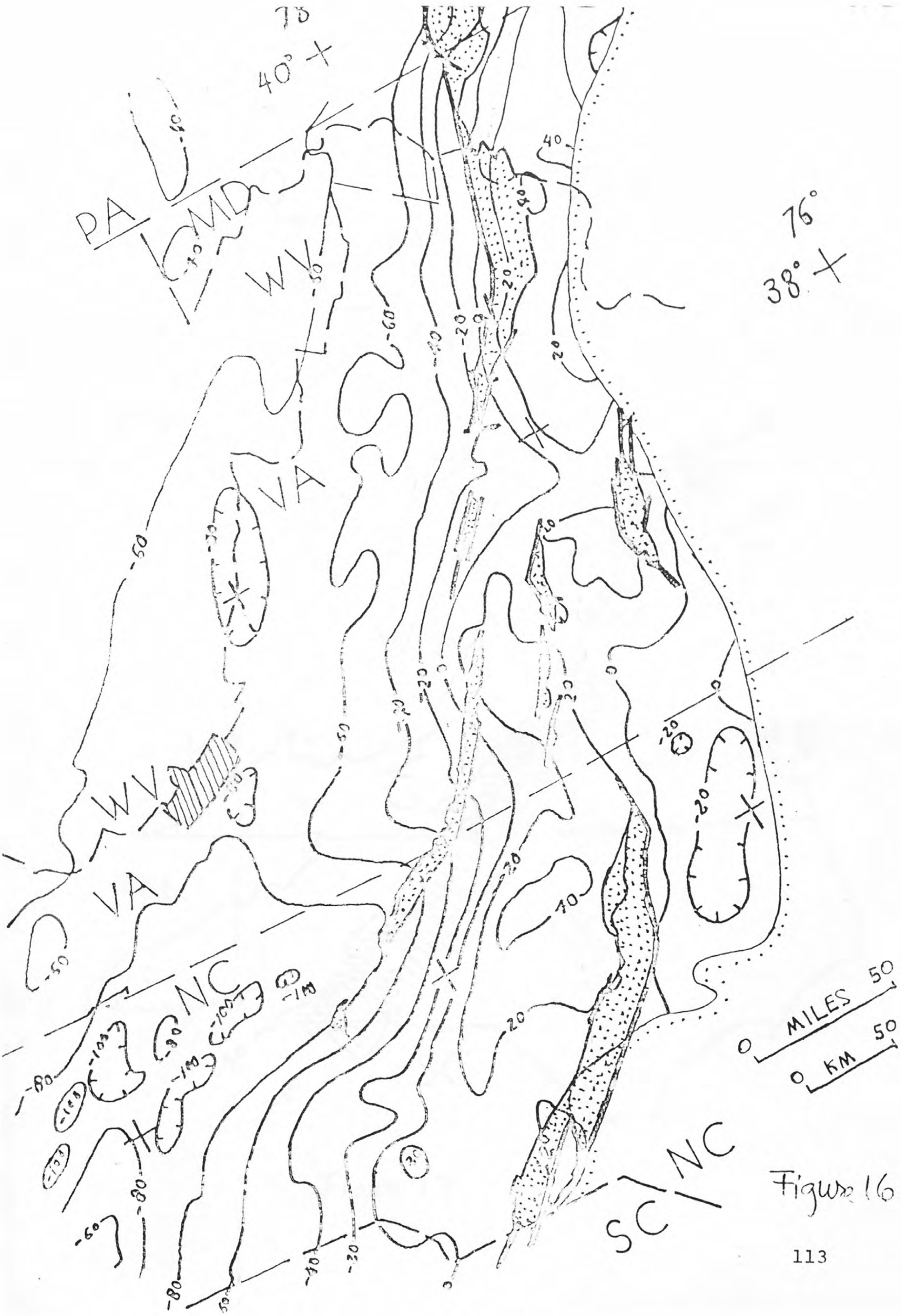


Figure 16

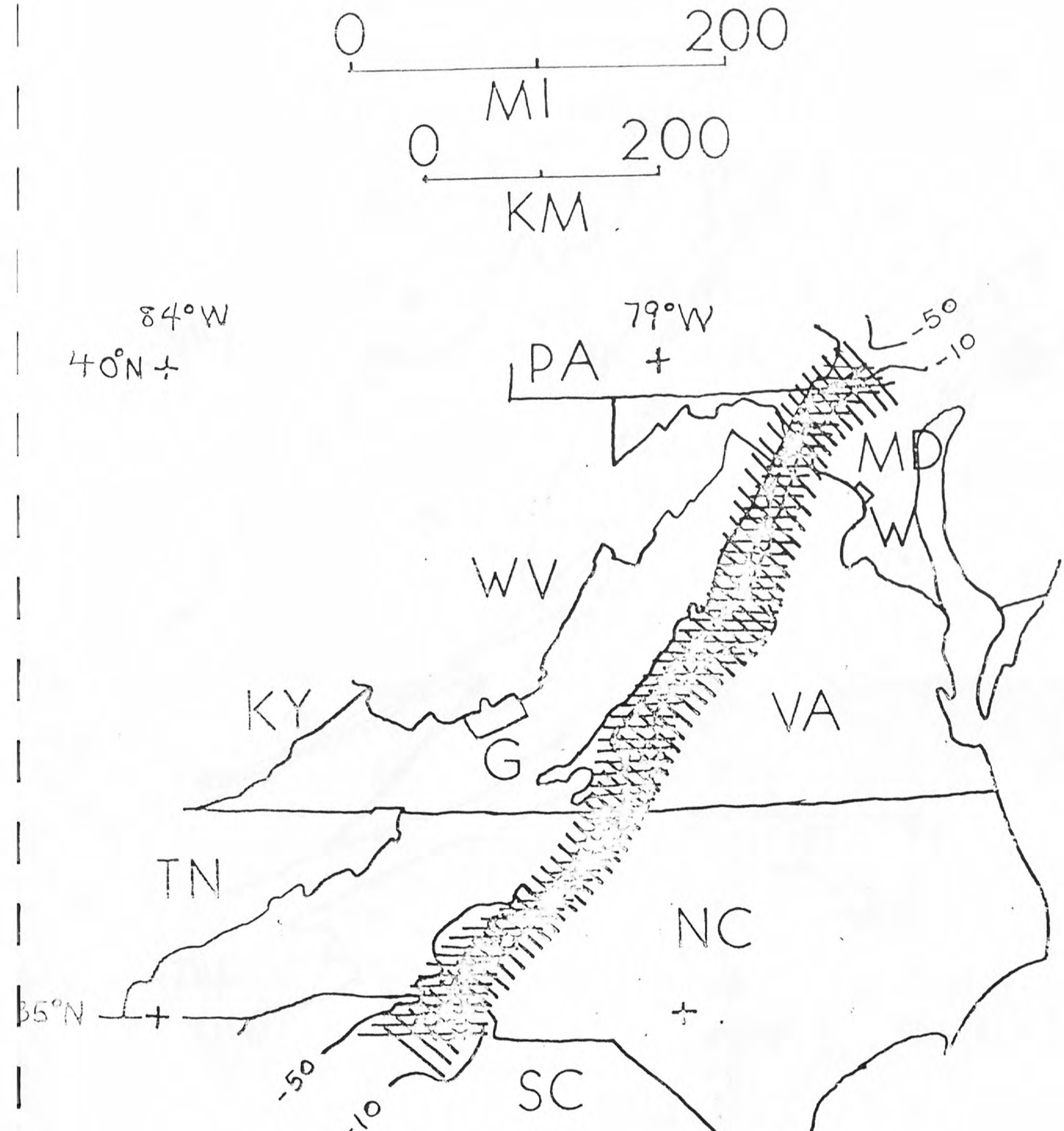

0 200
MI
0 200
KM
84°W
79°W
40°N
PA
-50
-10
MD
W
WV
KY
VA
G
TN
NC
35°N
-50
-10
SC

Figure 17

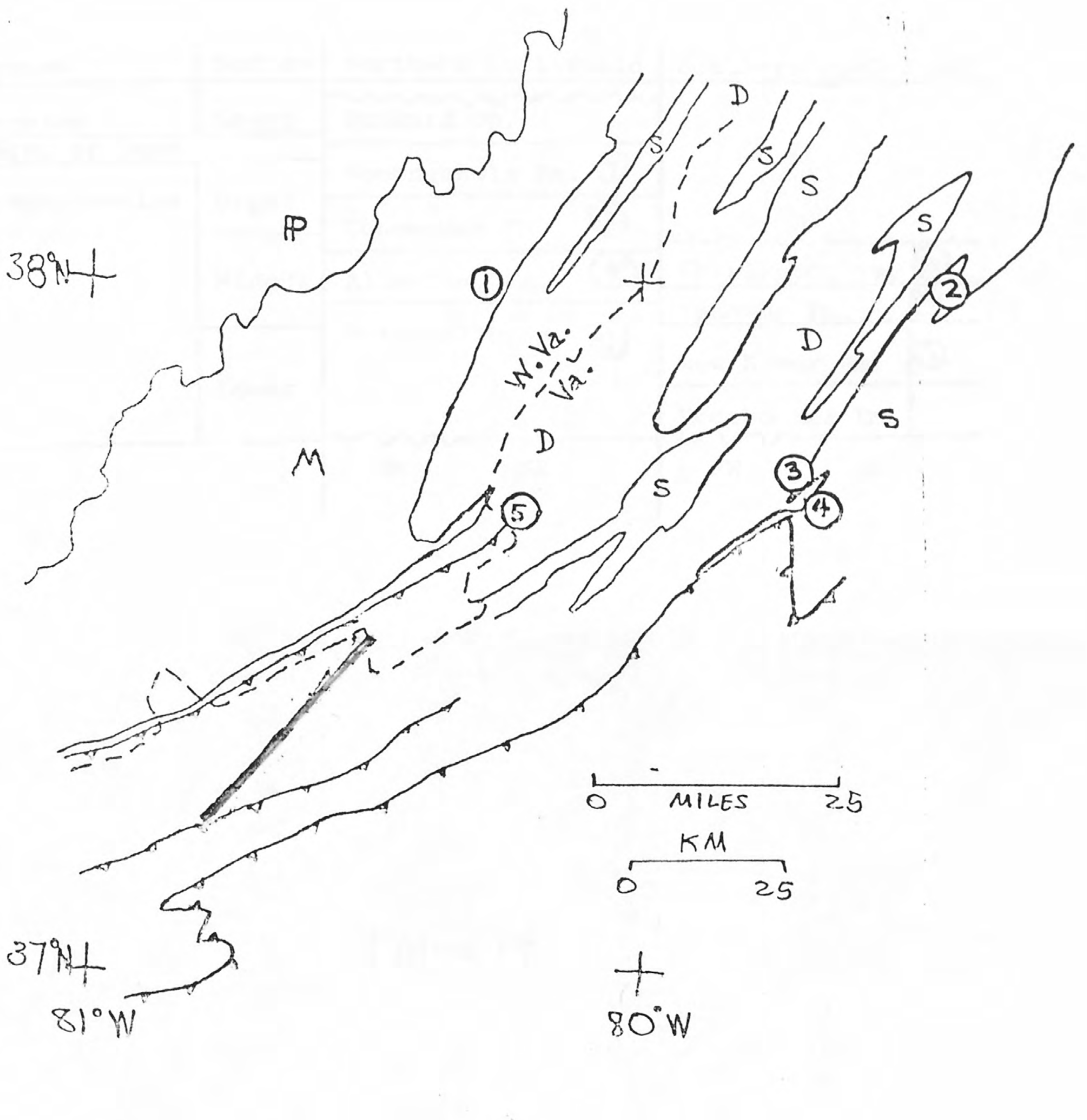


Figure 18

System	Series	Northern coal field	Southern coal field
Permian	Lower	Dunkard Gp.	
Perm. or Penn.			
Pennsylvanian	Upper	Monongahela Fm. 7	
		Conemaugh Fm. 6	
	Middle	Allegheny Fm. 5	Charleston Ss. 3
			Kanawha Fm. 2
	Lower	Pottsville Gp. 4	New River Rm. 1
			Pocahontas Fm.
		NW SE	NW SE

Figure 19

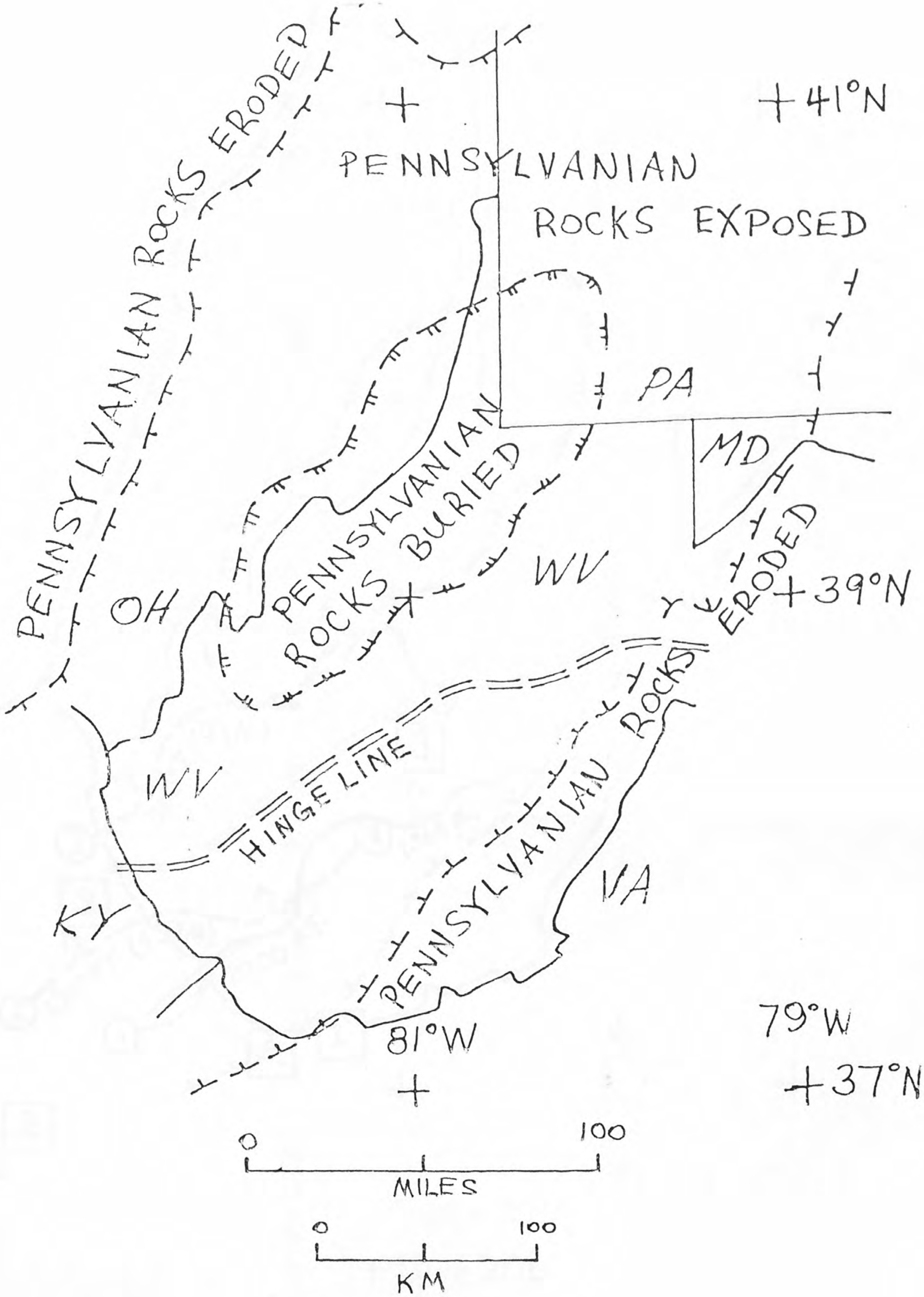

PENNSYLVANIAN ROCKS ERODED
PENNSYLVANIAN
ROCKS EXPOSED
41°N
PA
MD
OH
WV
PENNSYLVANIAN
ROCKS BURIED
39°N
ROCKS ERODED
HINGE LINE
WV
PENNSYLVANIAN
VA
KY
81°W
79°W
37°N
0
100
MILES
0
100
KM

Figure 20a

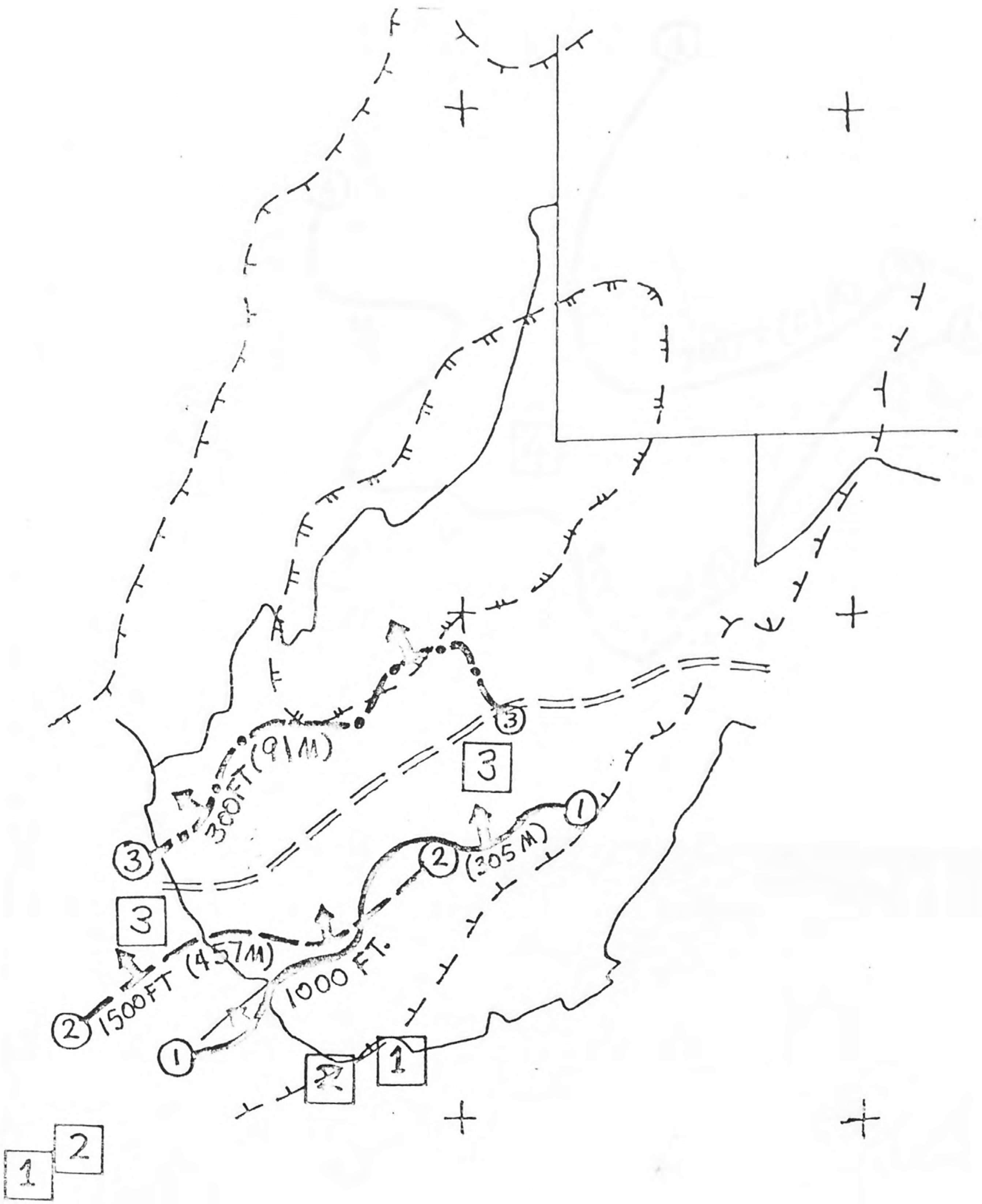


Figure 20b

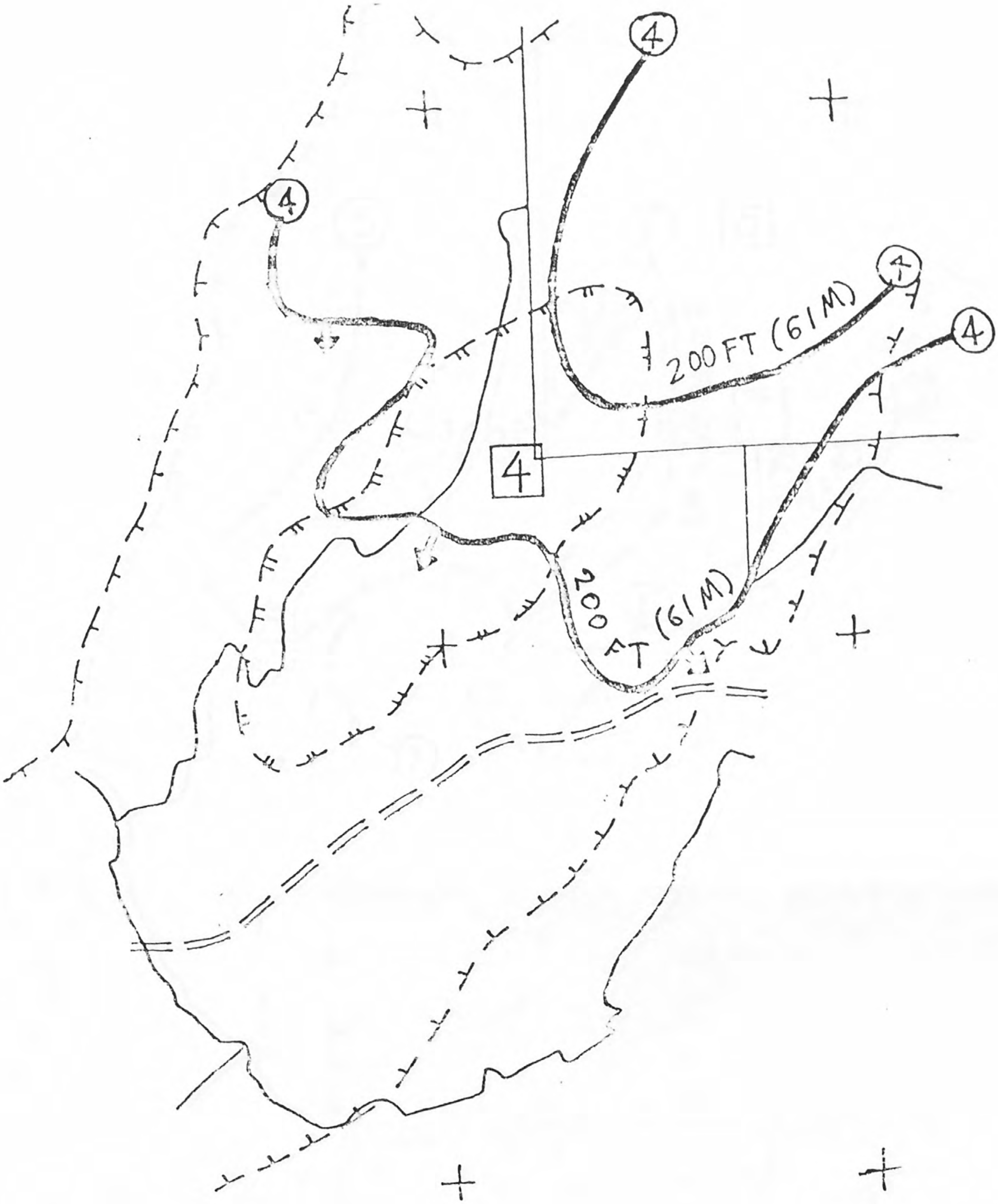


Figure 20a

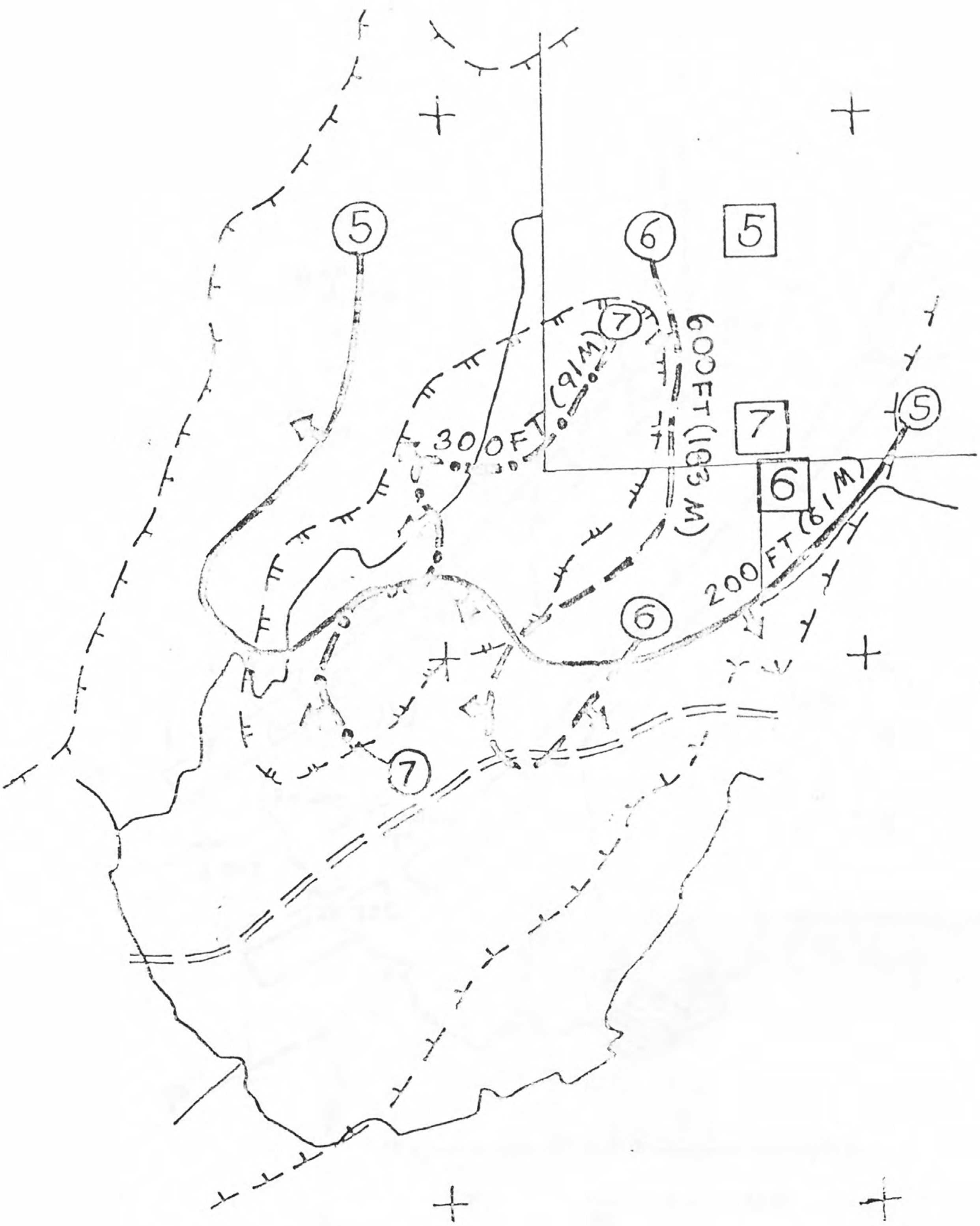


Figure 20d

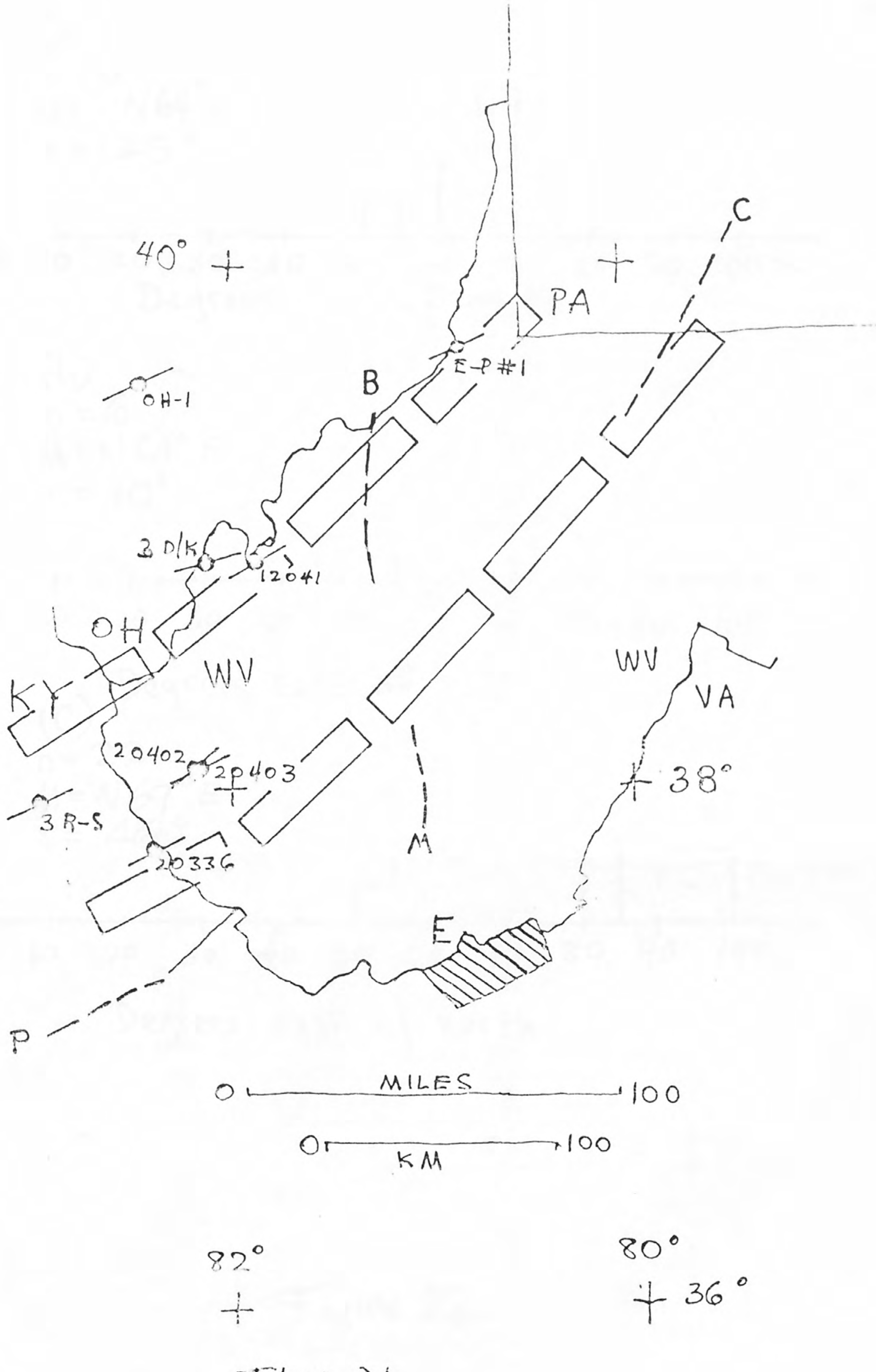

C
40°
PA
E-P #1
B
OH-1
3 D/K
12041
OH
WV
KY
WV
VA
20402
20403
38°
3 R-S
M
20336
E
P
MILES
0
100
KM
0
100
82°
80°
36°

Figure 21

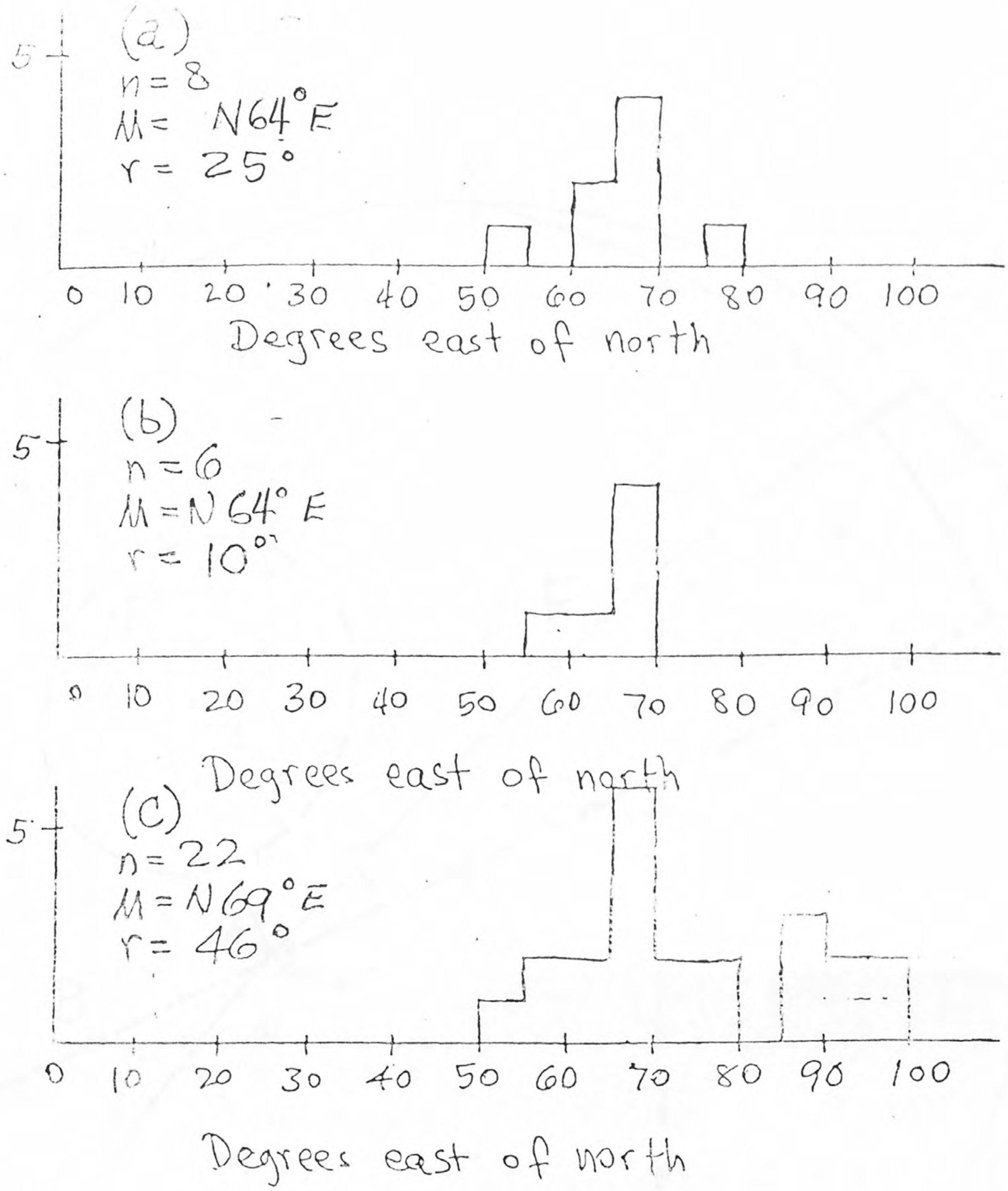

(a)
n = 8
M = N64°E
r = 25°
5
0 10 20 30 40 50 60 70 80 90 100
Degrees east of north
(b)
n = 6
M = N 64° E
r = 10°
5
0 10 20 30 40 50 60 70 80 90 100
Degrees east of north
(c)
n = 22
M = N69°E
r = 46°
5
0 10 20 30 40 50 60 70 80 90 100
Degrees east of north

Figure 22

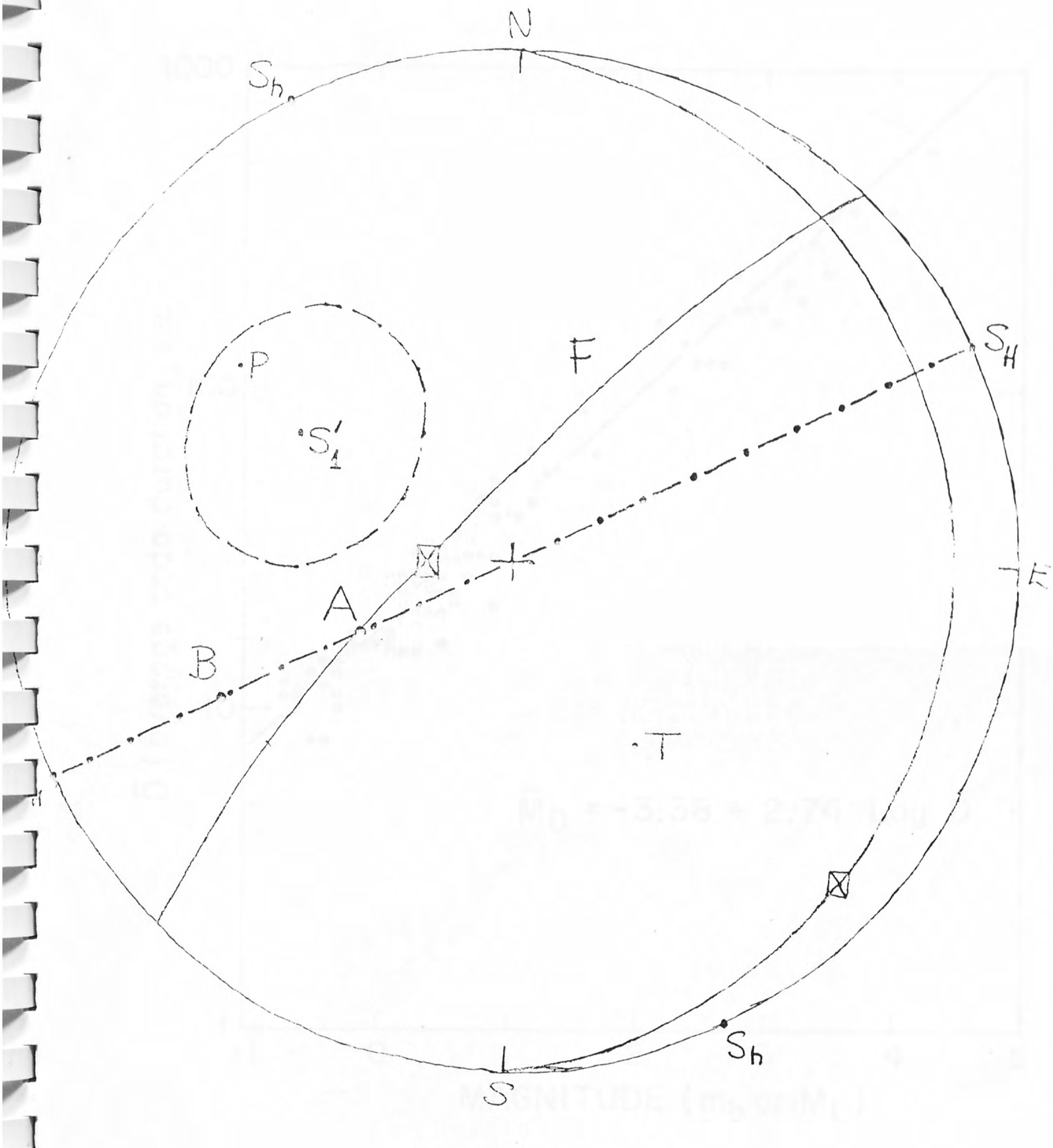


Figure 23

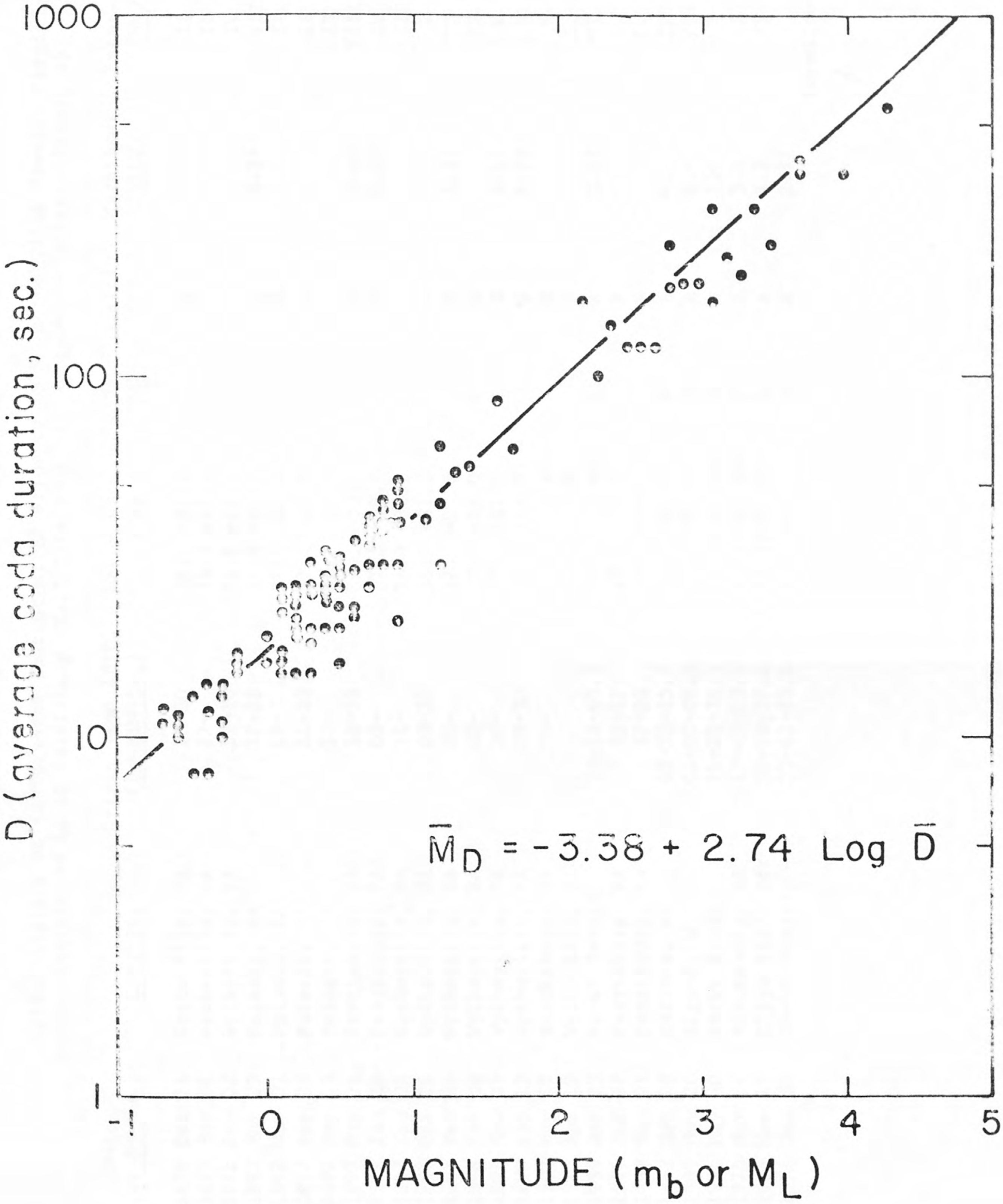

1000
100
10
1
-1
0
1
2
3
4
5
$\bar{D}$ (average coda duration, sec.)
MAGNITUDE (m_b or M_L)
$\bar{M}_D = -3.38 + 2.74 \text{ Log } \bar{D}$

Figure 24

TABLE 1

Chronological Listing of Earthquakes that Occurred Prior to 1978 in the Giles County, Virginia, Locale (within 50 km of Pearisburg, Va.) Data Sources: Reagor and others, 1980a, b)

Date (Yr-Mo-Day)	Locality	Origin Time (UTC) (Hr-Min-Sec)	Lat - Long (N-W)	DPTH (KM)	Quality (*)	Magnitude (mbLg)	Intensity (MM)
1876 Dec 21	Wytheville, VA	15-30	36.9 -81.1		G		II
1879 Sep 01	Wytheville, VA	12-	36.9 -81.1		G		II
1885 Feb 02	Wytheville, VA	12-10	36.9 -81.1		G		IV
1897 May 03	Pulaski, VA	17-18	37.1 -80.7		G	4.3+	VII
1897 May 03	Pulaski, VA	19-	37.1 -80.7		G		III
1897 May 03	Pulaski, VA	21-10	37.1 -80.7		G		III
1897 May 03	Pulaski, VA	23-	37.1 -80.7		G		III
1897 May 31	Pearisburg, VA	18-58	37.3 -80.7		G	5.8+	VIII
1897 Jun 29	Pearisburg, VA	03-	37.3 -80.7		G	4.0+	IV
1897 Sep 04	Wytheville, VA	11-	36.9 -81.1		G		III
1897 Oct 22	Wytheville, VA	03-20	36.9 -81.1		G		V
1898 Feb 05	Wytheville, VA	20-	37.0 -81.0		G	4.3+	VI
1898 Feb 06	Wytheville, VA	02-	37.0 -81.0		G		II
1898 Nov 25	Wytheville, VA	20-	37.0 -81.0		G	4.6+	V
1899 Feb 13	Wytheville, VA	09-30	37.0 -81.0		G	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	VA-WV Border	20-53-40.2	37.40-80.68	1	A	3.8?	VI
1959 Jul 07	Pearisburg, VA	23-17	37.3 -80.7		F		IV
1959 Aug 21	Pearisburg, VA	17-20	37.3 -80.7		F		IV
1968 Mar 08	Narrows, VA	05-38-15.7	37.28-80.77	8	A	4.1	IV
1969 Nov 20	Elgood, WV	01-00-09.3	37.45-80.93	3	A	4.6	VI
1974 May 30	VA-WV Border	21-28-35.3	37.46-80.54	5	A	3.6	V
1975 Mar 07	Blacksburg, VA	12-45-13.5	37.32-80.48	5	A	3.0	II
1975 Nov 11	Giles-Bland Bdr	08-10-37.6	37.22-80.89	1	A	3.2	VI
1976 Jul 03	VA-WV Border	20-53-45.8	37.33-81.13	1	A	2.1!	

(continued)

TABLE 1
(continued)

*The letter code in the QUALITY column is defined below:
A. Determination of instrumental hypocenters is estimated to be accurate within the ranges of latitude and longitude listed below; 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
B. 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

mbLg - Body wave magnitude according to Nuttli (1973) and Bollinger (1979).
MM - Modified Mercalli intensity rating (Roman numerals) according to Wood and Newmann (1931).
+ - Magnitude (mbLg) according to Nuttli and others (1979).
# - Magnitude (mbLg) according to J. W. Dewey and D. W. Gordon (written commun., 1980).
! - Magnitude (mbLg) according to Bollinger, G. A. (unpublished data, 1976).

TABLE 2

Site, Instrumentation and Operation Information
for the Giles County, Va., Subnetwork of the
Virginia Tech Seismic Network

Site Information

Code	Station Name	Lat (Deg N)	Long (Deg W)	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, WV	37.5905	80.8408	521	4/78	Mississippian Clastics
PWV	Princeton, WV	37.3348	81.0488	820	3/78	Mississippian Clastics
BLA	Blacksburg, VA	37.2114	80.4211	634	1962	Cambrian Carbonates

Instrumentation

Code	Seismometer#	T_o (sec)	T_g (sec)*	Type Recording**	Magnification at T_o	Maximum Magnification
NAV	SPZ (L4-C)	1.0	0.1	V,F,T	65K	310K @ 0.15 sec
PUV	SPZ (L4-C)	1.0	0.1	F,T	75K	390K @ 0.15 sec
HWV	SPZ (L4-C)	1.0	0.1	F,T	53K	320K @ 0.15 sec
PWV	SPZ (L4-C)	1.0	0.1	F,T	32K	160K @ 0.15 sec
BLA	SPZ (J-M)	1.0	0.1	V,F,T	30K	97K @ 0.30 sec

Operation***

Code	Years of Operation	Total Down Days	Percent Downtime
NAV	2.60	33	3.5
PUV	2.24	29	3.5
HWV	2.13	4	0.5
PWV	2.21	27	3.3
BLA	2.60	1	0.1

Comments

Timing System: Systron-Donner Time Code Generator 8120
Direction of Motion of Records: Up on record for up on ground
System Response Curves: See Figure 3
Two horizontal sensors added at PUV early in 1980.
Magnifications listed are for the visual recorders.
#SPZ = short-period vertical seismometer
L4-C = Mark Products design
J-M = Johnson-Matheson design
*High-cut filter setting
**V = visual; F = 16 mm film; T = FM magnetic tape
***Through June 1, 1980

TABLE 3

Velocity Model (TPM2)
Developed for the Giles County, Virginia, Locale
by Moore (1979)

Depth (km)	P Velocity (V_P, km/sec)	S Velocity (V_S, km/sec)	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

TABLE 4

HYPOELLIPSE EPICENTER LOCATION ERRORS FOR GILES COUNTY, VIRGINIA, BLASTS

Blast ID	Date of Blast	Difference: Actual and Calculated Epicenter (km)	ERH* (km)
A	December 3, 1979	0.5	2.2
B	December 6, 1979	0.9	2.4
C	May 20, 1980	2.0	5.7

*ERH = standard error of the epicenter (Lahr, 1979)

TABLE 5

HYPOELLIPSE DETERMINATION OF FOCAL DEPTHS FOR GILES COUNTY, VIRGINIA, BLASTS

Blast ID	Date of Blast	Trial Focal Depth (km)	Solution Focal Depth (km)	ERZ* (km)
A	December 3, 1979	4.0	0.5	57.7
		10.0	0.2	99.0
B	December 6, 1979	0.0	0.0	99.0
		5.0	2.5	16.7
C	May 20, 1980	4.0	2.2	14.3

*ERZ = standard error of the solution focal depth (Lahr, 1979)

TABLE 6

Chronological Listing of Earthquakes that Occurred Subsequent to 1977
in the Giles County, Va., Locale (within 50 km of Pearisburg, Va.)
and Were Located Using Network Data and the HYPOELLIPSE Program
(USGS Open-File Report 79-431, Lahr, 1979)

Map ID	Date (Yr-Mo-Day)	Origin Time (UTC) (Hr-Min-Sec)	Lat - Long (N-W)	Depth (Km)	MAGN* (MD)	RMS! (sec)	Error Ellipsoid Semi-Axes+ (Horiz.: Km, Deg);	(Vert.: Km)	Quality#
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	04-19-09.6	37°-12.80' 80°-49.82'	26.2	0.3	0.09	1.5,-46;4.4,44	3.0	B
34	1978 May 25	08-30-25.1	37°-00.01' 80°-47.65'	12.1	1.5	0.23	2.7,-86;4.3,4	3.8	B
35	1978 Jun 01	01-33-01.0	37°-17.99' 80°-41.98'	17.3	-0.2	0.17	2.1,-49;8.8,41	9.1	C
37	1978 Jul 28	08-39-40.7	37°-20.22' 80°-41.41'	11.8	0.6	0.27	2.2,-51;4.9,39	8.1	C
38	1978 Aug 30	02-19-38.2	37°-21.71' 80°-40.06'	8.4	0.5	0.09	1.0,-62;3.1,28	6.4	C
39	1978 Sep 14	19-37-06.6	37°-29.22' 81°-12.80'	9.9	-0.4	0.17	3.6,20;6.6,-70	17.4	D
40	1978 Oct 14	01-50-51.0	37°-17.68' 80°-28.03'	20.1	0.3	0.06	3.8,16;5.3,-74	17.2	D
46	1980 Feb 18	03-58-55.3	37°-25.78' 80°-35.54'	13.0	1.1	0.25	1.2,-41;1.7,-131	3.6	B
58	1980 Oct 09	01-47-01.1	37°-13.01' 80°-49.32'	23.5	-0.2	0.25	2.3,-50;7.2,40	4.9	C
60	1980 Oct 14	01-20-04.6	37°-04.69' 80°-13.82'	11.0	1.7	0.35	1.1,-77;2.0,13	3.1	B
63	1980 Dec 02	07-47-33.2	37°-25.08' 80°-32.25'	12.2	0.4	0.34	2.0,-39;3.2,-129	7.4	C

* - Average network magnitude: MD = -3.38 + 2.74 LOG (D) where D = average duration (sec) at network stations from the onset of the P-wave until return of vibrations to background microseismic level.

! - Root-mean-square error of the travel-time residuals (observed seismic wave travel-time minus calculated seismic wave travel-time).

+ - Projection onto the earth's surface (horizontal) and onto a vertical plane of the 68% confidence ellipsoid on the hypocentral coordinates. This projection is specified by giving the lengths, in km, and the trend, in degrees (plus-and-minus from north) of the semi-major and semi-minor axes along with the length, in km, of the vertical semi-axis.

\# - Quality factor according to HYPOELLIPSE (Lahr, 1979):
The lengths and azimuths of the axes of this ellipse are calculated as described above.
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

TABLE 7

Chronological Listing of Earthquakes that Occurred Prior to 1978 in the Giles County, Va., Locale (within 50 km of Pearisburg, Va.) and Were Relocated Using Joint Hypocenter Determination Techniques (J. W. Dewey and D. W. Gordon, written commun., 1980)

Map ID	Date (Yr-Mo-Day)	Locality	Origin Time (UTC) (Hr-Min-Sec)	Lat - Long (N-W)		Depth (Km)	MAGN (mbLg)	90% Conf. Ellipsoid Proj.* (Trend, Deg)	(Semi-Lengths, km)
D	1959 Apr 28	VA-WV Border	20-58-40.2	37°-23.70'	80°-40.92'	5 +	3.8	98.1	12.9, 7.7
H	1968 Mar 08	Narrows, VA	05-38-15.7	37°-16.86'	80°-46.44'	7.7	4.1	133.5	6.5, 6.1
J	1969 Nov 20	Elgood, WV	01-00-09.3	37°-26.94'	80°-55.92'	2.5	4.6	132.7	6.2, 4.4
R	1974 May 30	VA-WV Border	21-28-35.3	37°-27.42'	80°-32.40'	5.4	3.7#	122.7	8.6, 5.1
S	1975 Nov 11	Giles-Bland Bdr	08-10-37.6	37°-13.02'	80°-53.52'	1.0+	3.2	144.8	11.6, 6.7
X	1976 Jul 03	VA-WV Border	20-53-45.8	37°-19.26'	81°-07.62'	1.0+	2.1!	141.3	13.7, 6.5

* - Projection onto the earth's surface of the 90% confidence ellipsoid on the hypocentral coordinates. This projection is specified by the trend, in degrees, of the semi-major axis and the lengths, in km, of the semi-major and semi-minor axes, respectively.

\+ - Focal depth fixed.

\# - Reagor and others (1980a) give a value of 3.6.

! - Magnitude according to G. A. Bollinger (unpublished data, 1976).

NOTE: Some of these same data appear in Table 1.

TABLE 8

P Wave Polarity Data for Giles County, Virginia, Earthquakes

Event No	Date	Station	AZM	AIN	P Wave Polarity
32	780128	BLA	274	81	eC
		NAV	157	67	eC
33	780510	HWV	179	63	iC
37	780728	HWV	155	73	eD
38	780830	BLA	307	80	eD
		HWV	149	81	iC
40	781014	NAV	95	61	eC
		HWV	135	75	iC
46	800218	BLA	328	69	iC
		NAV	55	62	iC
		HWV	129	68	iC
		PWV	75	76	eC
		[illegible]UV	24	79	eD
		CVL	72	68	eC

NOTES: Event No. refers to the listing in Table 6.
Date is given as year-month-day.
AZM is the epicenter-to-station azimuth, in degrees.
AIN is the angle-of-incidence (measured from the down-going vertical) at the focus, in degrees.
P Wave Polarity - C for compression, D for dilatation, e for emergent beginning of P wave and i for impulsive beginning of P wave.

TABLE 9

Locations, Sources and Values of Selected Stress Orientations

State (County)	Well Code	Source & Code	Method	Stress	Depth	n	References
OH (Hocking)	?	Z(OH-1)	HF	N64°E	808 m (2650 ft)	NA	Haimson (1974)
KY (Martin)	20336	E(KY3)	PCL	N63°E	758-1038 m (2486-3404 ft)	1573	Evans (1979), p. 244-260; Wilson and others (1980)
KY (Johnson)	3 R-S	E(KY4)	PCL	N65°E	290-457 m (950-1500 ft)	3	Evans (1979), p. 261-276
WV (Wetzel)	E-P No. 1	E(WV7)	PCL	N67°E	1859-2027 m (6100-6650 ft)	11	Evans (1979), p. 161-176
WV (Mason)	3 D/K	E(WV5)	PCL	N75°E	826-1042 m (2711-3420 ft)	1268	Evans (1979), p. 125-146
WV (Jackson)	12041	E(WV2)	PCL	N60°E	981-1125 m (3220-3690 ft)	738	Evans (1979), p. 67-84
WV (Lincoln)	20402	Z(WV-1)	HF	N50°E	835-839 m (2738-2752 ft)	NA	Evans (1979), p. 106-124; Haimson (1977); Abou-Sayed and others (1978)
WV (Lincoln)	20403	E(WV3)	PCL	N65°E	829-1227 m (2720-4025 ft)	1215	Evans (1979), p. 85-105

NOTES: Well codes are usually permit numbers. Sources are Zoback and Zoback (1980, Z in table) and Evans (1979, E in table). Codes shown in parentheses after Z or E are well designators used by those authors. HF means stress orientation was obtained by hydrofracturing the well and then determining strike(s) of vertical cracks in hole wall. PCL means orientation was determined by measuring strikes of vertical portions of n petal-centerline fractures (defined in text) in an oriented core. Depths are below ground level. References cited give further information about the wells.

TABLE 10

AVERAGE MICROSEISMIC AMPLITUDE LEVELS (NANOMETERS)

Province	Stations	January Day/Night	March Day/Night	June Day/Night	September Day/Night	December Day/Night
Plateau	HWV				3/-	18/-
Plateau	PWV	11/15	3/3	2/-	3/3	4/5
Valley & Ridge	NAV			-/1		
Valley & Ridge	PUV	23/60	2/3	2/1	-/9	-/32
Valley & Ridge	BLA	5/8	2/2	2/1	2/2	3/12
Piedmont	CVL	11/-	3/2	-/1	2/-	11/8
Piedmont	GHV	-/12				
Piedmont	FRV	13/50		1/-	-/1	
Coastal Plain	PBV		2/2	2/2	1/1	3/4
Averages		13/29	2/2	2/1	2/3	8/12

AVERAGE MICROSEISMIC FREQUENCIES (HERTZ)

Province	Stations	January Day/Night	March Day/Night	June Day/Night	September Day/Night	December Day/Night
Plateau	HWV				2.3/ -	0.8/ -
Plateau	PWV	1.1/1.0	2.9/3.1	3.1/ -	3.0/2.9	3.1/1.7
Valley & Ridge	NAV			- /3.2		
Valley & Ridge	PUV	0.8/0.6	2.8/1.4	2.5/3.4	- /0.9	- /0.7
Valley & Ridge	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
Piedmont	GHV	- /1.0				
Piedmont	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
Averages		1.1/0.9	3.1/3.5	2.9/3.0	2.7/2.7	1.9/1.3

TABLE 11

Data Set Used in the Determination of:

$M_D = -3.38 + 2.74 \text{ Log } (D) \qquad n = 102$

D (sec)	M_L	m_b	D (sec)	M_L	m_b	D (sec)	M_L	m_b
15	0.3		16	-0.2		188	3.3	
16	0.5		16	-0.3		209	3.2	
21	0.9		16	0.1		232	3.5	
30	0.8		17	-0.2		361	4.0	
30	0.9		18	0.3		361	3.7	
30	1.2		19	0.0		8	-0.4	
40	1.1		19	0.2		8	-0.5	
44		1.2	20	0.5		10	-0.6	
120		2.5	21	0.2		10	-0.6	
120		2.6	24	0.4		10	-0.3	
15	0.1		24	0.1		11	-0.3	
16	0.0		25	0.3		11	-0.7	
16	0.1		26	0.4		11	-0.6	
20	0.3		26	0.5		30	0.3	
20	0.4		26	0.1		31	0.5	
21	0.2		26	0.2		32	0.4	
22	0.1		26	0.2		35	0.6	
23	0.2		26	0.3		36	0.9	
23	0.5		27	0.4		37	0.8	
23	0.6		29	0.5		39	0.8	
23	0.6		34	0.7		42	0.8	
25	0.4		36	0.7		44	0.9	
26	0.7		37	0.7		51	0.9	
28	0.5		39	0.8		86	1.6	
29	0.6		39	0.7		117	2.7	
29	0.5		42	0.8		137	2.4	
30	0.7		51	0.9		160	2.2	
12	-0.4		52	1.3		179	3.0	
12	-0.7		53	1.4		180	2.9	
13	-0.5		62	1.7		232	2.8	
13	-0.3		64	1.2		287	3.4	
13	-0.3		100	2.3		294	3.1	
14	-0.4		159	3.1		392	3.7	
15	0.2		174	2.8		545	4.3	

TABLE 12

Average Network Duration Magnitudes

Recalculated (October, 1980) Using $M_D = -3.38 + 2.74 \text{ Log } D$

Map #	Date	Time (UTC)	Average Duration (sec)	M_D	m_b (Lg)	M_L
21	9/13/76	18:54	193	2.9	3.1	
32	1/28/78	23:13	66	1.6	(2.9)	(2.4)
32A	3/17/78	18:26	152	2.6	2.8	
33	5/10/78	04:19	22	0.3		0.6
34	5/25/78	08:30	59	1.5		1.0
35	6/1/78	01:33	14	-0.2		(0.5)
36	6/22/78	06:42	43	1.1	(2.27)	1.5
37	7/28/78	08:39	28	0.6		0.6
38	8/30/78	02:19	25	0.5		0.7
39	9/14/78	19:37	12	-0.4		(0.7)
40	10/14/78	01:50	22	0.3		(1.0)
41	10/29/78	12:22	44	1.1		1.6
42	11/15/78	08:33	72	1.7	2.1	2.1
43	11/06/79	03:04	70	1.7	1.3	
44	11/12/79	07:21	44	1.1	1.2	
45	1/6/80	13:50	66	1.6	(1.0)	
46	2/18/80	03:58	42	1.1		0.9
47	4/10/80	22:33	30	0.7		0.9
48	4/22/80	03:14	110	2.2	(2.8)	
49	4/26/80	03:59	58	1.4	(3.0)	1.7
50	5/18/80	03:21	36	0.9		1.2
51	5/18/80	22:33	13	-0.3		(1.2)
52	7/7/80	17:02	20	0.2		(1.4)
53	8/4/80	10:13	30	0.7		1.0
54	9/18/80	01:28	22	0.3		0.7
55A	9/21/80	10:02	56	1.4	(2.6)	1.5
56	9/24/80	06:41	42	1.1	(2.0)	1.4
57	9/26/80	01:31	89	2.0	(3.5)	2.2
57A	9/26/80	05:04	19	0.1		(1.0)
58	10/9/80	01:47	14	-0.2		0.4
59	10/11/80	22:40	[illegible]	0.7		1.1
60	10/14/80	01:20	71	1.7		1.9
61	10/16/80	03:48	44	1.1		1.5

NOTE: () = unacceptable value, probably due to inappropriate wave period or phase; delete.

Map # refers to Table 6 and G. A. Bollinger (1980, unpublished data).