NUREG/CR-5123

Studies of the Pattern and Ages of Post-Metamorphic Faults in the Piedmont of Virginia and North Carolina

Prepared by L. Glover, III, J.K. Costain, C. Coruh

Department of Geological Sciences
Virginia Polytechnic Institute and State University

Prepared for
U.S. Nuclear Regulatory
Commission

NOTICE

NOTICE

Availability of Reference Materials Cited in NRC Publications

Most documents cited in NRC publications will be available from one of the following sources:

1. The NRC Public Document Room, 1717 H Street, N.W. Washington, DC 20555
2. The Superintendent of Documents, U.S. Government Printing Office, Post Office Box 37082, Washington, DC 20013-7082
3. The National Technical Information Service, Springfield, VA 22161

Although the listing that follows represents the majority of documents cited in NRC publications, it is not intended to be exhaustive.

Referenced documents available for inspection and copying for a fee from the NRC Public Document Room include NRC correspondence and internal NRC memoranda; NRC Office of Inspection and Enforcement bulletins, circulars, information notices, inspection and investigation notices; Licensee Event Reports; vendor reports and correspondence; Commission papers; and applicant and licensee documents and correspondence.

The following documents in the NUREG series are available for purchase from the GPO Sales Program: formal NRC staff and contractor reports, NRC-sponsored conference proceedings, and NRC booklets and brochures. Also available are Regulatory Guides, NRC regulations in the *Code of Federal Regulations*, and *Nuclear Regulatory Commission Issuances*.

Documents available from the National Technical Information Service include NUREG series reports and technical reports prepared by other federal agencies and reports prepared by the Atomic Energy Commission, forerunner agency to the Nuclear Regulatory Commission.

Documents available from public and special technical libraries include all open literature items, such as books, journal and periodical articles, and transactions. *Federal Register* notices, federal and state legislation, and congressional reports can usually be obtained from these libraries.

Documents such as theses, dissertations, foreign reports and translations, and non-NRC conference proceedings are available for purchase from the organization sponsoring the publication cited.

Single copies of NRC draft reports are available free, to the extent of supply, upon written request to the Division of Information Support Services, Distribution Section, U.S. Nuclear Regulatory Commission, Washington, DC 20555.

Copies of industry codes and standards used in a substantive manner in the NRC regulatory process are maintained at the NRC Library, 7920 Norfolk Avenue, Bethesda, Maryland, and are available there for reference use by the public. Codes and standards are usually copyrighted and may be purchased from the originating organization or, if they are American National Standards, from the American National Standards Institute, 1430 Broadway, New York, NY 10018.

NUREG/CR-5123
RA

Studies of the Pattern and Ages of Post-Metamorphic Faults in the Piedmont of Virginia and North Carolina

Manuscript Completed: December 1987
Date Published: April 1988

Prepared by
L. Glover, III, J.K. Costain, C. Coruh

Department of Geological Sciences
Virginia Polytechnic Institute and State University
Blacksburg, Virginia 24061

Prepared for
Division of Engineering
Office of Nuclear Regulatory Research
U.S. Nuclear Regulatory Commission
Washington, DC 20555
NRC FIN A9027
Under Contract No. NRC-04-75-237

TABLE OF CONTENTS

LIST OF ILLUSTRATIONS

SEISMICITY, SEISMIC REFLECTION, GRAVITY AND GEOLOGY OF THE CENTRAL VIRGINIA SEISMIC ZONE: PART 1, REFLECTION SEISMOLOGY

Cahit Çoruh, John K. Costain, Lynn Glover, III, Thomas L. Pratt, and Jeanne Brennan

Abstract

In central Virginia, 180 km of vibroseis reflection data have been obtained along the James River traverse from near Richmond on the east to the crest of the Blue Ridge anticlinorium on the west. Laboratory determinations of velocity and density were made to identify potential seismic marker horizons in the Valley and Ridge, Blue Ridge, and Piedmont provinces. One of the best impedance contrasts within Valley and Ridge relatively unmetamorphosed lower Paleozoic shelf strata is at the Rome-Shady boundary (Lower Cambrian) where a reflection coefficient of between 0.2 and 0.4 can be expected. Beneath the Blue Ridge, the master decollement is at a depth of approximately 3 km (1 s) at the western end of the traverse where Grenville basement is acoustically transparent. Eastward-dipping reflections between 1.0 - 3.0 s were recorded from beneath the Blue Ridge decollement, probably from relatively unmetamorphosed lower Paleozoic shelf strata (Rome). Large-amplitude reflections from a depth of about 9 km (2.8-3.0 s) beneath the Blue Ridge are from the Paleozoic shelf strata, probably from the Rome-Shady boundary in para-autochthonous strata at 9 km. East of the Blue Ridge, the best reflections originate from the Catoctin rift volcanics and the Chopawamsic island arc volcanics. Eastward-dipping reflections from the Catoctin Formation dominate the seismic section below the Triassic Scottsville basin. A window of poor reflection quality associated with the small impedance contrasts of the Triassic outlines the gross geometry of the Scottsville basin, and indicates a basin thickness of approximately 1.6 km (0.7 s). Measured sections of the Catoctin revealed thin beds of metamorphosed, epidotized basalts and sandstones interlayered with non-epidotized, foliated metabasalts and metasediments. Compressional velocities parallel and perpendicular to the dominant foliation in 12 Catoctin Formation samples and one Chilhowee Formation sample were determined. Velocities range from 5.13 km/s to 6.47 km/s for samples of the Catoctin epidozites, greenstones, phyllites, and volcanic breccia. Velocities parallel to foliation are higher than those perpendicular to foliation. The highly epidotized metamorphosed basalts and sandstones can result in reflection coefficients of magnitudes $\geq$ 0.1. Reflections from the Catoctin Formation of central Virginia are believed to be due in large part to constructive interference from acoustically thin, epidotized layers (epidosites and volcanic breccia) interlayered with non-epidotized layers (greenstones and phyllites), and to a lesser extent from non-epidotized greenstones and phyllites. We conclude that the successful definition of the regional geologic framework of the crystalline rocks of the Piedmont of Virginia depends to a large extent on the placement of reflection seismic traverses where thick sequences of metamorphosed basaltic/felsic volcanics, or metamorphosed basalts/sandstones occur in the subsurface.

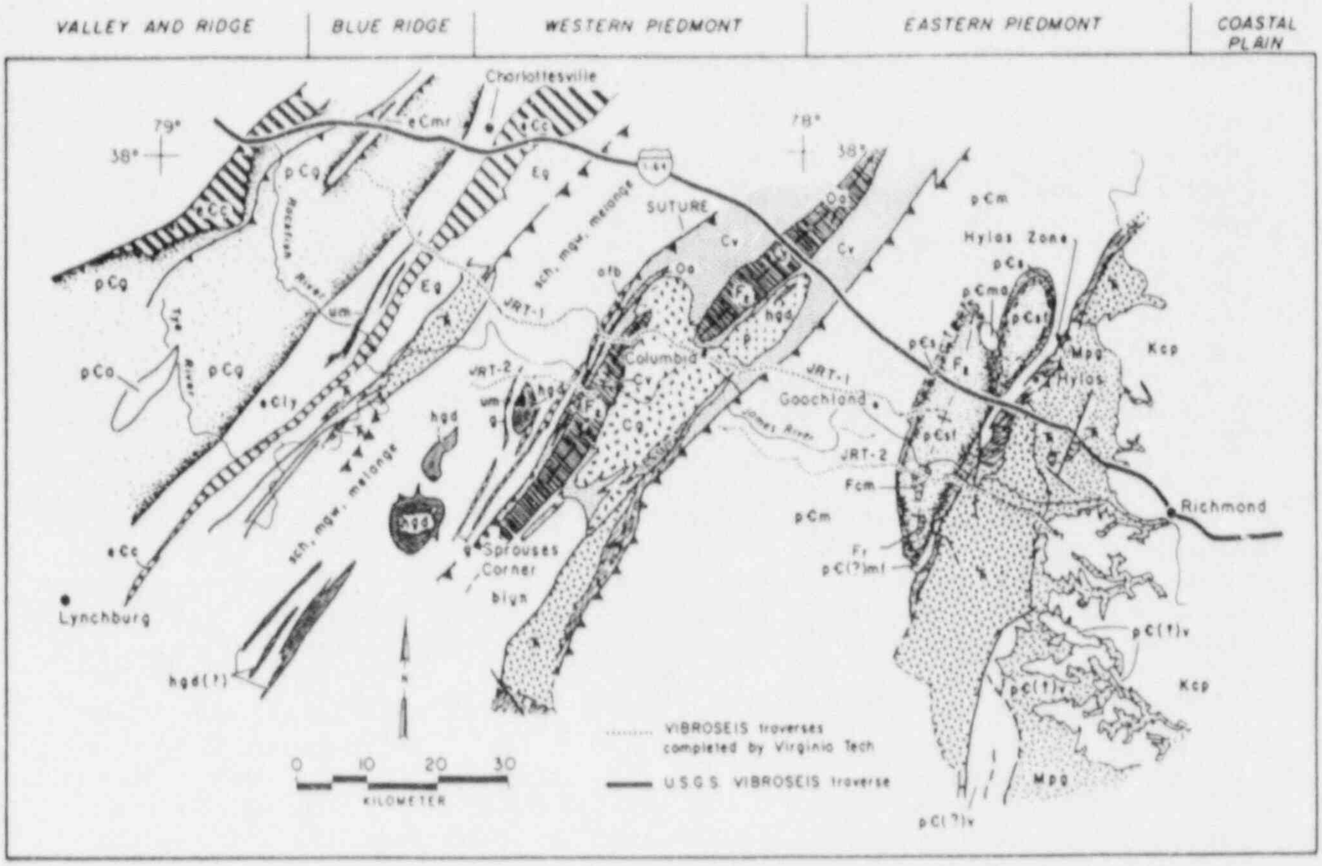
Introduction

In recent years seismic reflection profiling has been increasingly successful in helping to understand the dynamic evolution of the Earth's crust. Multifold reflection profiles over major tectonic features provide the data base for interpreting the geometry and internal structure of the crust. This is especially true in the crystalline Appalachians whose allochthonous architecture requires seismic reflection data to relate surface geology to deep crustal structure. A combined geophysical and geological study of the tectonic framework of the central Appalachians has been underway at the Regional Geophysics Laboratory (RGL) and the Orogenic Studies Laboratory (OSL) of the Department of Geological Sciences at Virginia Polytechnic Institute and State University since 1974. In Virginia, 180 km of vibroseis reflection data have been obtained by RGL along the James River (Figure 1).

Profile segments funded by different sponsors have been combined to give relatively continuous regional traverses. The first traverse (JRT-1) is located on the north side of the James River from near Richmond on the east to the crest of the Blue Ridge anticlinorium on the west (Figure 1). The total length of traverse JRT-1 is 108 km. A second traverse (JRT-2), 74 km long, is located south of the James River. In this paper, the key data segments for interpretation of the two seismic traverses are discussed along with general comments on seismic reflection profiling over crystalline terrane.

Previous work

Seismic reflection profiles discussed in this paper were obtained over an igneous and metamorphic terrane containing metamorphosed basalts, felsic volcanics and sandstone that was thrust over relatively unmetamorphosed shelf strata. Therefore, a discussion of the crustal reflectivity to be expected is helpful to understand features of the final processed record sections as well as limitations that must be considered while interpreting the data. Over the past dozen years, we have made determinations of velocity and density on samples of shelf strata and igneous and metamorphic rocks representative of lithologies in the Valley and Ridge, Blue Ridge, and Piedmont provinces of Virginia and North Carolina (Kolich, 1974; Edsall, 1974; Wells, 1975; Clark and others, 1978). Sample locations are given by Kolich (1974), Edsall (1974), and Wells (1975). Acoustic impedance (ρv) was determined from density determinations on representative fresh samples and from velocity determinations on the same samples in a pressure cell (Kolich, 1974). These were used to determine reflection coefficients that might be expected from the sedimentary strata and crystalline terranes in the southeastern U.S.. From these results, a range of impedance values for igneous, metamorphic



EXPLANATION

Kcp	Cretaceous Atlantic Coastal Plain	Cv	Cambrian Chopawamsic volcanics
	unconformity	e€c	Eocambrian Catoctin metabasalt
TR	Triassic/Jurassic(?) basin deposits	e€ly	Eocambrian Lynchburg formation
	unconformity	e€mr	Eocambrian Mechums River formation
Mpg	Petersburg granite	p€(?)mf	late Precambrian(?) Moseley felsic gneiss
Fr	Flat Rock granite	p€(?)v	late Precambrian(?) Eastern slate belt volcanics
Fcm	Fine Creek Mills granite		*unconformity*
Oa	Ordovician Arvonia formation	p€o	Grenville Roseland metanorthosite
	unconformity	p€ma	Grenville Montpelier metanorthosite
Cg	Ordovician Columbia granite	p€m	Grenville Maidens gneiss
sch, mgw, melange, ofb	Cambrian(?)-Ordovician(?) Includes g, greenstone; hgd, hornblende gabbro; um, ultramafic rocks; ofb, ocean floor basalt	p€s	Grenville Sabot amphibolite
		p€sf	Grenville State Farm gneiss
Eg	Cambrian(?) Evington group	p€g	Grenville Blue Ridge basement gneiss

Figure 1. Locations of seismic lines and general surface geology: Geology from Glover and others (second part of this report).

and sedimentary rocks similar to those that might be expected along the James River traverses was determined.

The signal-to-noise (S/N) ratio is the most important single parameter that affects the ultimate quality of seismic record sections. This depends to a certain extent on the magnitudes of the *reflection* coefficients as well as the two-way *transmission* coefficient (Waters, 1981, p. 26). Although a reflection coefficient greater than $\pm$ 0.3 is extremely high and reflects a significant portion of the energy, the *two-way* transmission coefficient across such a reflector is 0.91. Thus, good reflectors do not necessarily mean poor energy return from below the reflector. In addition, "tuning" can significantly increase the S/N ratio.

Normal incidence reflection coefficients, $(\rho_2 v_2 - \rho_1 v_1)/(\rho_2 v_2 + \rho_1 v_1)$ computed from the values of velocity and density given in Table 1 vary between about 0.03 and 0.4, the higher value representing the contrast between the Rome shale and Shady dolomite. Contrasts between tectonically juxtapositioned basalt-dolomite and basalt-gneiss result in poor reflectivity (reflection coefficient $\leq$ 0.03.).

It has been known for some time that unmetamorphosed Valley and Ridge shelf strata underlie at least part of the crystalline Blue Ridge (Clark and others, 1978; Cook and others, 1979; Harris and others, 1981; Harris and Bayer, 1979). Our field and laboratory determinations of velocity and density suggest that one of the best impedance contrasts in the Valley and Ridge unmetamorphosed lower Paleozoic section is at the Rome-Shady boundary (Lower Cambrian) where shales of the Rome Formation (acoustic impedance = 0.88×10^6 - 1.09×10^6 cgs units) overlie dolomites of the Shady Formation (acoustic impedance = 1.71×10^6 - 2.15×10^6 cgs units). A reflection coefficient of between 0.2 and 0.4 can be expected from this part of the sedimentary section; other good reflecting intervals have been illustrated by Gresko and Costain (1985). Thick sequences of predominantly shale can be expected in the Rome; where exposed in southwestern Virginia, the Shady is about 700 m in thickness (Perry and others, 1979). The Rome-Shady boundary is thus an important structural marker bed for seismic investigations where these strata have been overthrust by crystalline rocks of the Blue Ridge and Piedmont, as for example under the western end of JRT-1. In igneous and metamorphic terrane, we have found that the best reflections originate from volcanic lithofacies such as the Catoctin rift volcanics and the Chopawamsic island arc volcanics.

Along the James River traverse, reflection seismic profiles crossed thick units whose subsurface geometry and continuity, or lack of it, were expected to reveal the structure in the upper crust. Surface geologic mapping and interpretation of potential field data along the James River corridor have identified key structural elements and thick metavolcanic lithofacies that are believed to persist to depths of more than several kilometers in the crust (Reilly, 1980; Keller and others, 1985, 1986). Certain lithofacies such as the Chopawamsic and Catoctin metavolcanics were expected to have excellent seismic response.

Data Acquisition And Processing

All data were acquired by RGL personnel. Recording instrumentation consisted of 48-channel MDS-10 amplifiers, a field summing unit (SMM-1), and a field correlator (DC-2400) for quality control. All final correlation was done on the RGL VAX 11/780 computer. Receiver arrays were electronically weighted for a Chebychev response over an effective array equivalent to a length of two group intervals. The receiver group interval was 70 m for all lines except Line NSF-2 where the interval was 35 m. In spite of the relatively short source-receiver distance for NSF-2, which generally makes it more difficult to determine accurate stacking velocities, good reflection quality was obtained (Figure 2 and Figure 3). The principal reason for using a receiver group interval of 35 m for Line NSF-2 was to obtain increased resolution over the Scottsville Triassic basin.

A single Failing Y-1100 vibrator with a peak force of 27,000 pounds was used to form source arrays. The number of pad positions per source array was varied from 8 to 64; an array of 16 pad positions was found to be adequate. A downsweep varying in length from 16 to 24 s was used for all lines except Line NSF-1 which was obtained with an upsweep. Bandwidth of the sweep was normally 10-60 Hz, although in some cases 10-80 Hz was used (Line NSF-2). The source array length was 70 m, except 35 m for NSF-2, and the interval between source arrays was varied from 35 m to 140 m depending on the receiver group interval and the multiplicity desired. Most segments of the two traverses are 12-fold. Three segments on JRT-1 are 24-fold. Comparison of 12 and 24-fold data clearly favors the higher multiplicity for better data quality and interpretation.

Processing of the seismic data was done in RGL using Digicon's DISCO software plus special processing modules developed in RGL. The typical data processing sequence included demultiplexing, vibroseis whitening (Çoruh and Costain, 1983), trace editing, common-mid-point sorting, elevation statics corrections, velocity analysis (constant velocity panels and velocity spectra), deconvolution (if appropriate), automatic residual statics, stacking, time migration, and bandpass filtering. Most of these steps are described in detail by Waters (1981) and Robinson and Treitel (1980).

One of the most effective processing steps was the whitening (VSW) of vibroseis data. In VSW processing, the spectrum of the data is balanced before cross-correlation to attenuate coherent noise and to recover weak reflections that have a higher frequency content. This processing step was applied to all data. Static corrections were especially critical on line NSF-1 (Traverse JRT-1), and an unconventional method of calculating elevation statics was used (Bahorich and others, 1982). Multi-step application of automatic residual statics helped to improve data quality.

Migration of seismic data is found to be important in the crystalline terrane examined here because of generally high velocities and complex geology. Considerable effort, therefore, was devoted to testing processing parameters in order to obtain accurate stacking velocities and optimum processing parameters to which the migration process is sensitive.

NSF - LINE 2 STACK

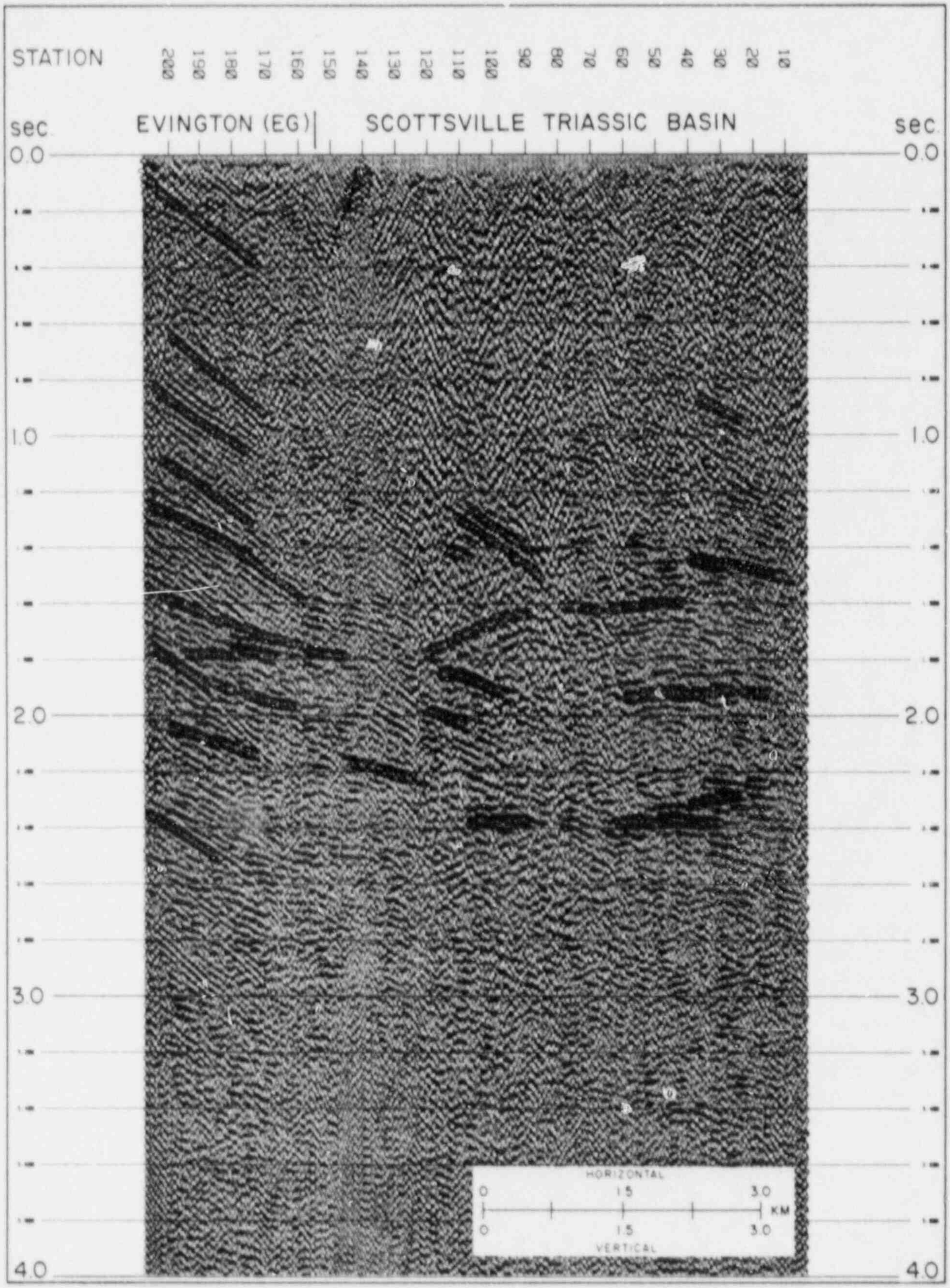


Figure 2. Line NSF-2, stacked: Stacked section; 24-fold and 35 m receiver interval.

NSF – LINE 2 MIGRATION

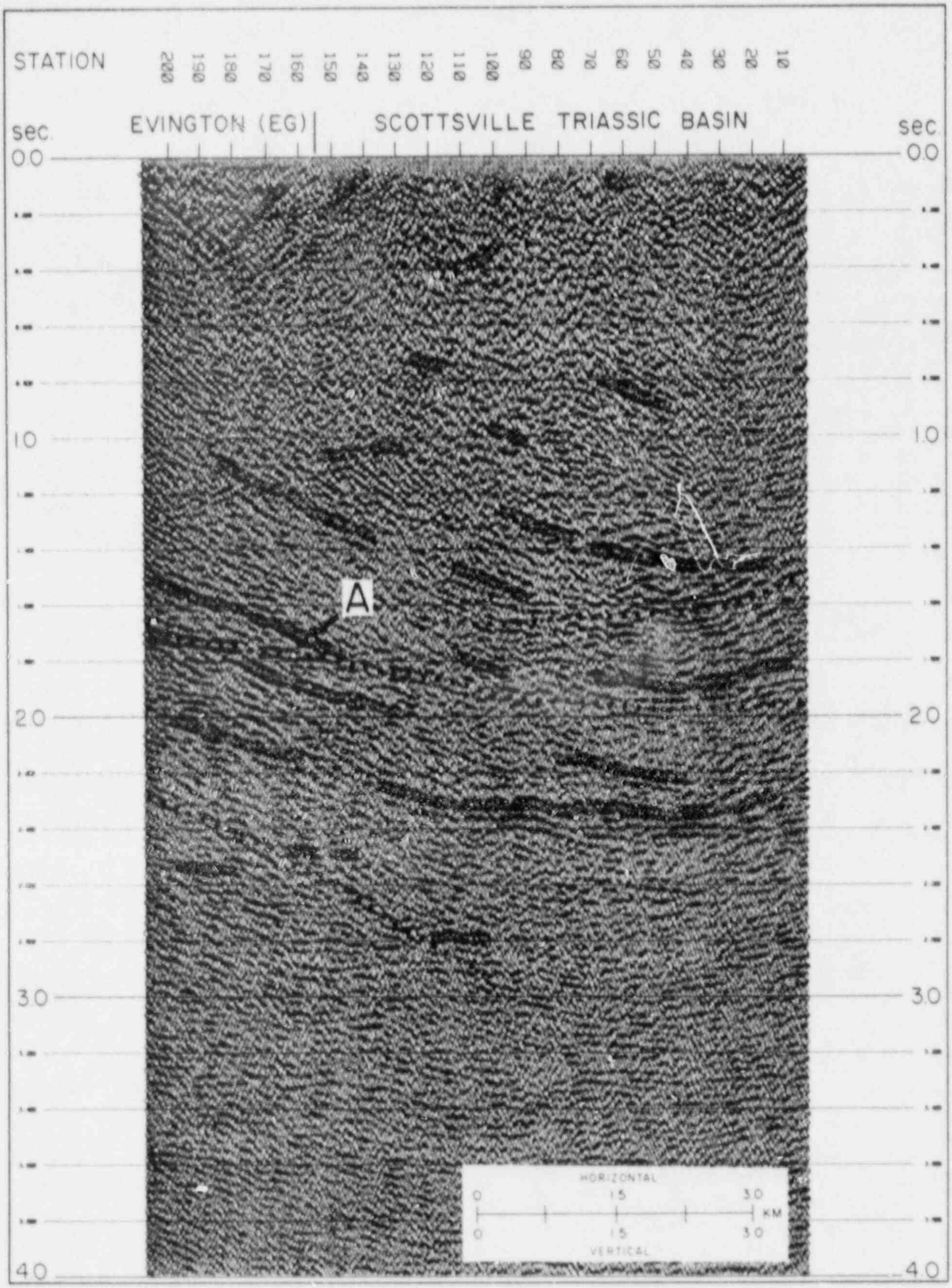


Figure 3. Line NSF-2, migrated: Time-migrated stacked section; 24-fold and 35 m receiver interval.

Seismic Signatures of Major Structural Elements and Tectonic Lithofacies in Central Virginia

Seismic traverse JRT-1 (Figure 1) is located north of the James River and is a composite of line segments NRC-5 (12-fold, 70 m group spacing), NSF-2 (24-fold, 35 m), NRC-7 (24-fold, 70 m), NSF-3 (12- and 24-fold, 70 m), and NSF-1 (24-fold, 70 m). A summary line drawing of the interpreted record section of traverse JRT-1 is shown in Figure 4 for correlation with illustrations of the actual seismic data shown below.

The seismic signatures of tectonic lithofacies are discussed on enlargements of the data taken from the seismic sections. In all cases, structural interpretations were carried out on time-migrated data which were compared with the unmigrated data. This paper (Part 2) deals primarily with the reflectivity and seismic signatures of igneous and metamorphic rocks in central Virginia. For this paper selected portions of traverse JRT-1 are discussed. In general, the excellent acoustic response of the rift-related Eocambrian Catoctin metabasalts and metasandstones, the Chopawamsic subduction-related metavolcanic rocks, and the relatively unmetamorphosed lower Paleozoic strata account for most of the energy return, and make an interpretation possible.

Blue Ridge Master Decollement. A portion of the stacked section of line NRC-5 (unmigrated) is shown in Figure 5. This is the westernmost extent of our traverse JRT-1. The Grenville basement of the Blue Ridge is exposed at the surface and accounts for the absence of shallow reflections to a time of about 1.0 s, the approximate location of the Blue Ridge master decollement. Good eastward-dipping reflections between 1.0 - 3.0 s appear from beneath the Grenville basement of the Blue Ridge. The quality and character of reflections in this window makes a source from within Grenville basement unlikely. The most plausible origin of these reflections is from unmetamorphosed lower Paleozoic strata (Rome). The Blue Ridge master decollement at the base of the Grenville is interpreted to be at a depth of approximately 3 km (1 s) at the west end of the traverse, in agreement with proprietary seismic data projected in from the northwest from the Valley and Ridge province, and at about the same depth as indicated on new seismic data from beneath the Blue Ridge about 400 km to the southwest in South Carolina (Çoruh and others, 1985).

Important evidence for the allochthonous nature of the Blue Ridge is shown in Figure 5 and Figure 6. The large-amplitude reflection between 2.8 and 3.0 s is characteristic of the deepest reflection from the unmetamorphosed rocks of the Valley and Ridge province, and is probably from near the Rome-Shady boundary near the sole fault. The eastward-dipping reflections at 2.5 s between stations 330 - 370 on Figure 5 and at 2.2 s between stations 150 and 170 on Figure 6 have a high-frequency signature, before filtering, similar to that seen on proprietary data in other areas of the Valley and Ridge below the Rome-Shady contact. Although Paleozoic strata are interpreted to occur in the time window from 2.2 s to 2.9 s, evidence of good reflections at times less than 1.0 s is not obvious on our data. Divergence of these reflections between stations 330 and 370 on Figure 5 and between stations 150 - 190 on Figure 6 is additional evidence for thrust faults.

A portion of the time-migrated stacked line NRC-5 is shown in Figure 6. This line crosses, from west to east, the Lynchburg Formation, Catoctin Formation, Evington Group, and the Scottsville Triassic basin. The eastern part of this segment (12-fold, 70 m) exactly overlaps with segment NSF-2 (24-fold, 35 m) shown in Figure 2 and Figure 3. Comparison of Figure 6 and Figure 3 clearly documents the advantages of higher multiplicity and spatial receiver density. Divergence of reflections below station 120 at 1.0 - 1.5 s (Figure 6) is interpreted to be characteristic of thrust

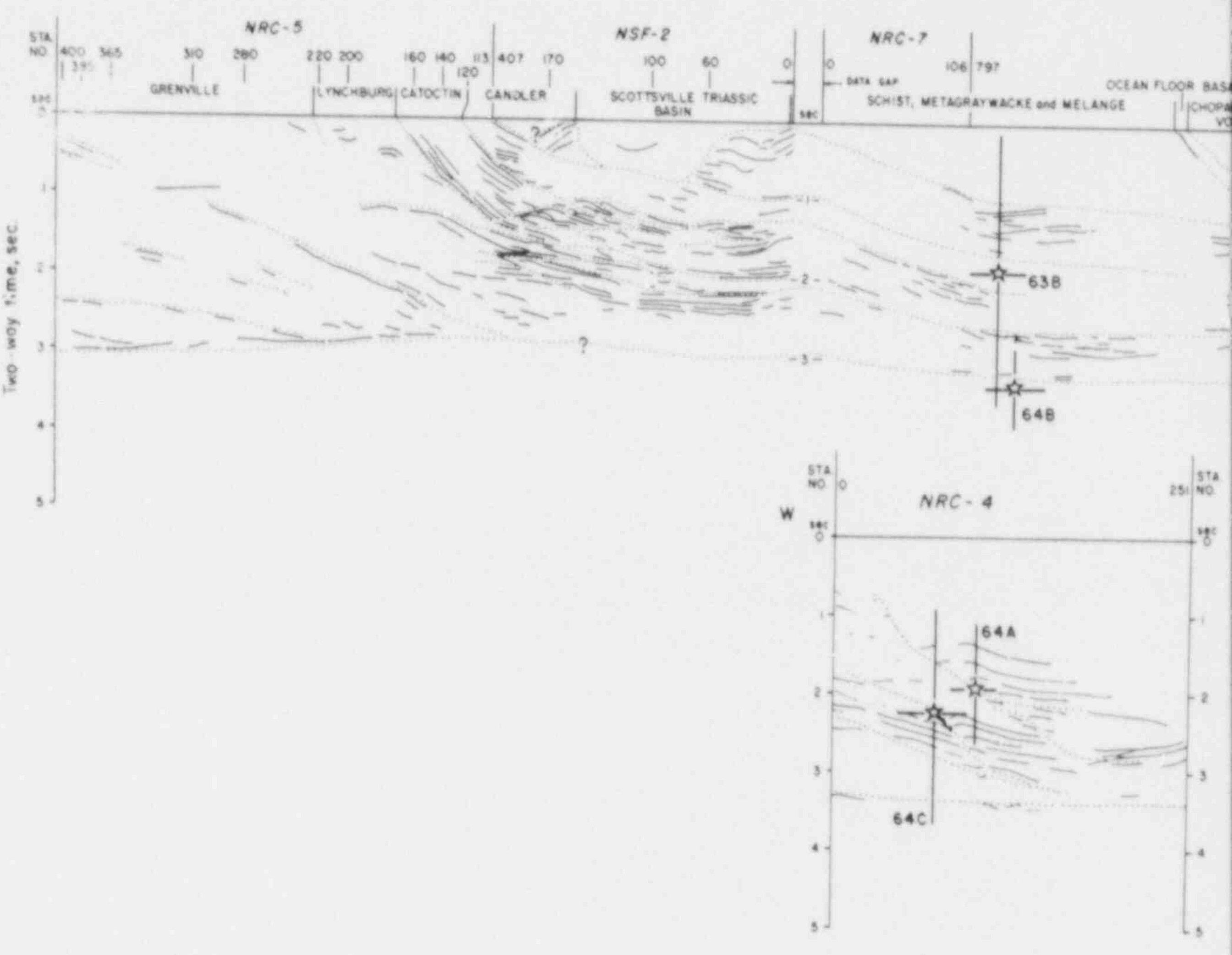
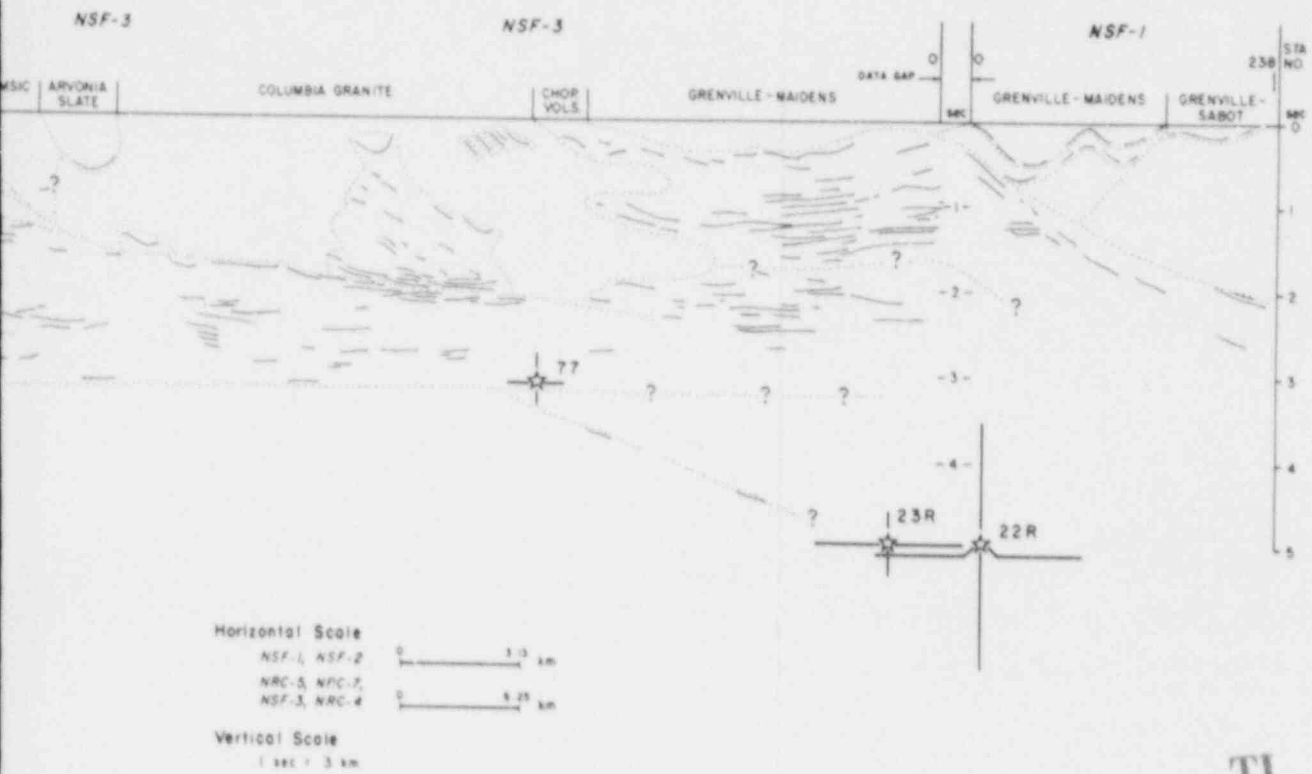


Figure 4. Line drawing of JRT-1 and NRC-4.

T-1



TI
APERTURE
CARD

Also Available On
Aperture Card

8805060296-01

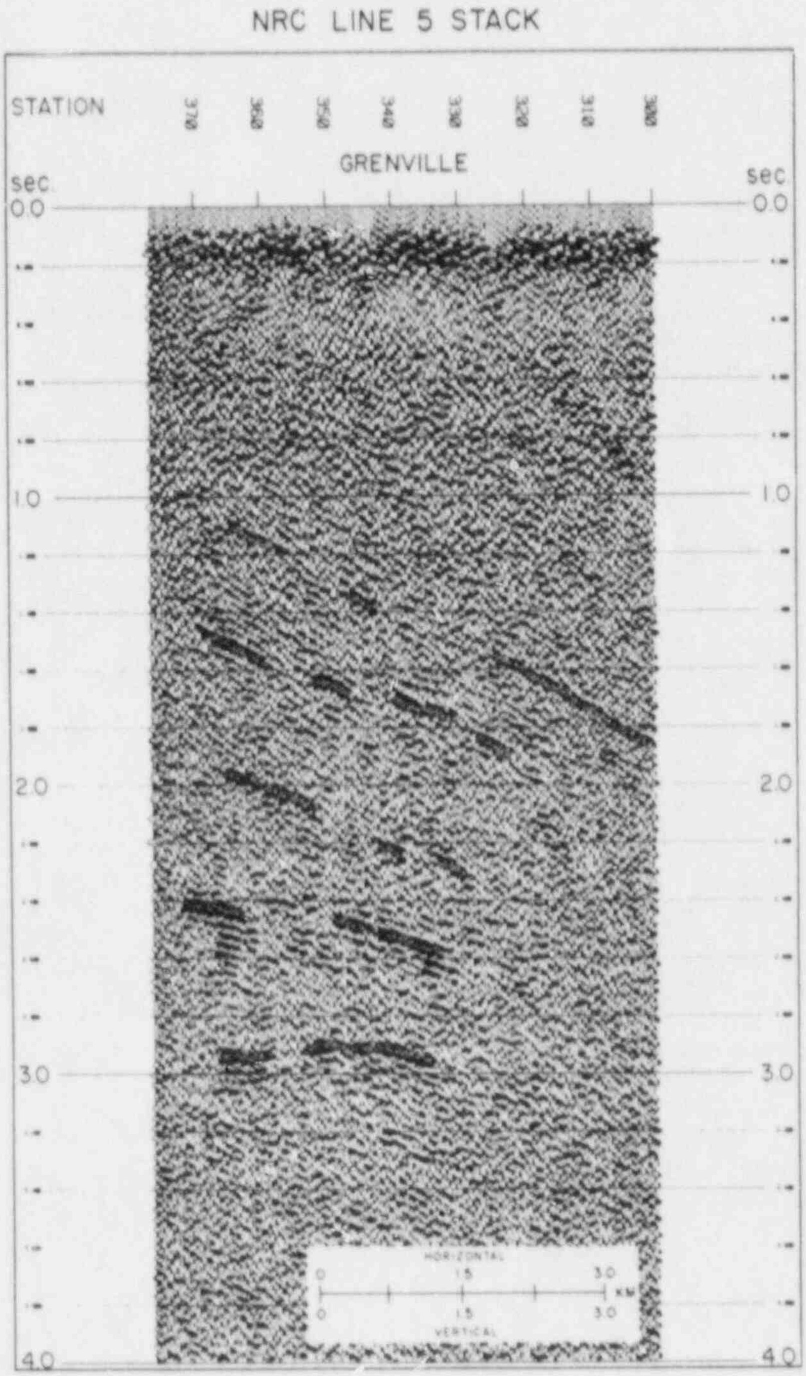


Figure 5. A portion of Line NRC-5, stacked: 12-fold; 70-m group interval.

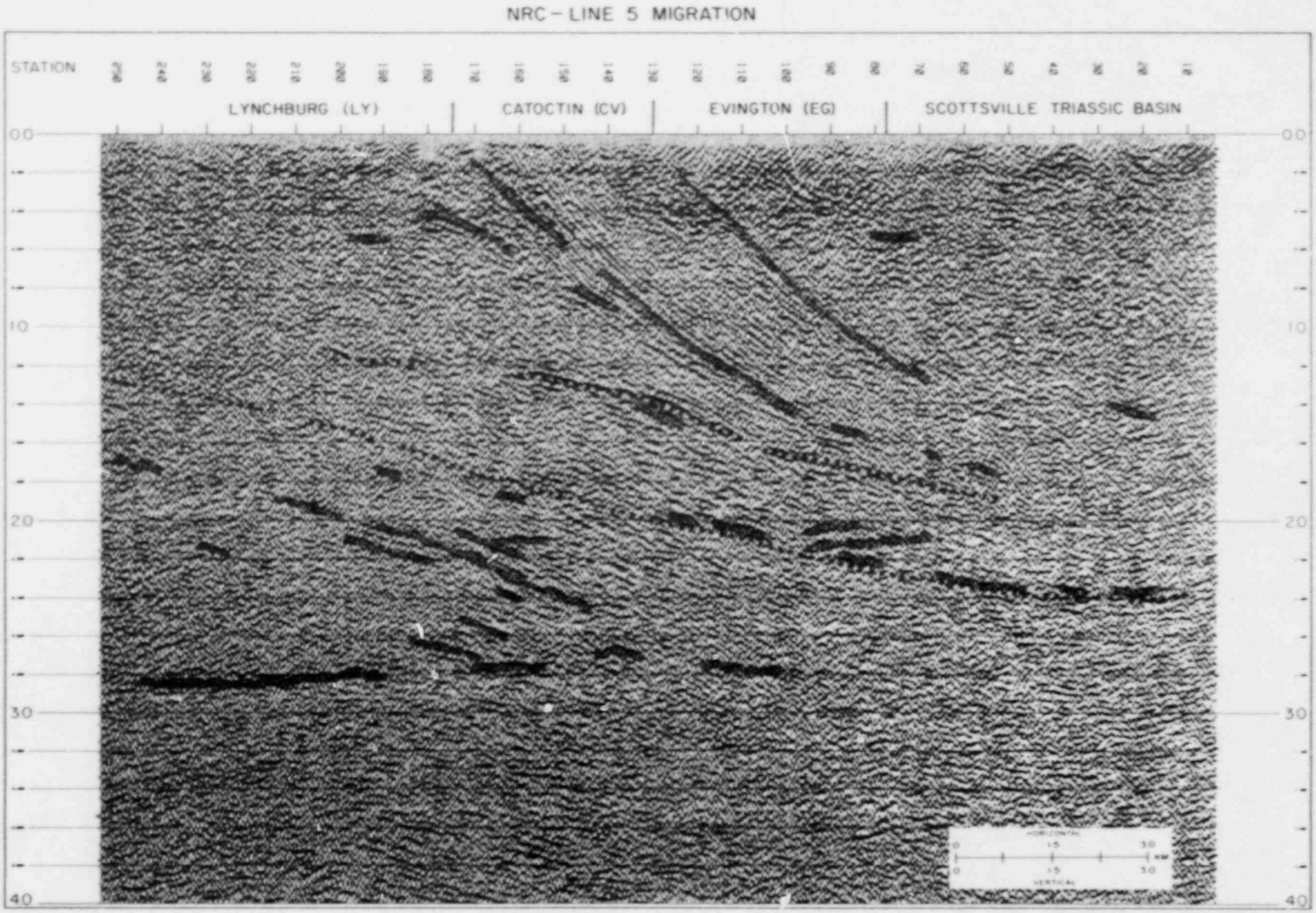


Figure 6. A portion of Line NRC-5, migrated: Time-migrated and stacked; 12-fold and 70 m receiver interval.

faults. Shallow westward-dipping reflections beneath stations 90 - 120 are consistent with dips of rocks of the Evington Group exposed at the surface.

Catoctin volcanic sequence. The seismic signature of the Catoctin volcanic sequence is well-defined in Figure 6; however, the signature is better with higher multiplicity and spatial density (Compare with Figure 2 and Figure 3. Divide station numbers by 2 because of a 35 m receiver interval). The reflections correlate well with mapped surface contacts of the Catoctin. The seismic signature of the Catoctin is characterized by approximately parallel reflections of large amplitude (Figure 6 and Figure 2). Relatively poor reflections are recorded from the top of the Catoctin, suggesting that the transition from the Catoctin into the Evington Group is not associated with as large an impedance contrast as that between the Catoctin and Lynchburg. The good reflection quality defined by large amplitudes and a well-formed Klauder wavelet at the Catoctin-Lynchburg transition is an aid to the interpretation of deeper reflections. The reflection at 1.4 s at station 130 (Figure 6) is one of the largest amplitude reflections on the section and, because of its truncation of reflectors below, it is interpreted as a thrust fault. The thrust probably represents the contact between the Basement and Paleozoic strata. Likewise, truncation of the horizontal reflectors near station 90 between 2.0 - 2.2 s is interpreted as an indication of a eastward dipping fault. Between stations 120 and 200 at approximately 2.8 s, eastward-dipping reflections are truncated by a horizontal reflection interpreted to be at or near a fault, and below 3.1 s the absence of prominent continuous reflections is consistent with the known reflection character of pre Rome strata and Grenville basement.

Scottsville Triassic Basin. The stacked section of Line NSF-2 (24-fold, 35 m) is shown in Figure 2. The line traverses the Scottsville Mesozoic basin (Costain and others, 1982). An excellent signal-to-noise ratio and well-formed Klauder wavelets are characteristic of this line. In general good eastward-dipping (from the Catoctin) and horizontal reflections dominate the section below the basin. A shallow westward-dipping event (station 170, time 0.2 s) in the Evington Group is consistent with surface mapping. An important feature suggested by this line is that the near-surface response of the Earth where the Evington Group is exposed is quite different than the response where the vibrator is over Triassic lithologies. On the left hand side of the section excellent reflections with well-formed Klauder wavelets indicate parallel reflectors. Although at first glance multiple reflections appear to be present in the data, velocity analyses indicate that the events are primary reflections. These excellent reflections correlate with the mapped surface geology, and clearly represent the acoustic response of the Catoctin Formation (Figure 1). Deeper events between 1.8 - 2.4 s at the left hand side of Figure 6 also show a similar Catoctin character. Divergence of reflected events suggests repetition by thrust faulting (see also Figure 3). Alternatively, these deeper reflections may be from a Paleozoic sedimentary section (see Glover and others, in the second part of this report).

The time-migrated section of Figure 2 is shown in Figure 3. Because of the large dips of reflections in Figure 2, migration has a significant effect on moving the reflections up to the northwest and, in some cases, out of the section (compare Figure 2 and Figure 3 at stations 170-210). A plausible geologic interpretation is possible only after migration. Dipping events are clearly truncated by other dipping or horizontal events. The reflections which truncate other reflections (A on Figure 3) are interpreted as being due to thrust faults. The deepest reflection arrives at approximately 2.8 s (8.7 km) and is probably from Grenville basement. If the reflections below 1.8 s are from the Catoctin, then an interpretation of back-sliding along a thrust fault might be made below station 170 on Figure 3 at about 1.8 s (5.4 km). This might have important implications for understanding the seismicity of the Central Virginia Seismic Zone.

Reflections from Triassic-Jurassic strata in onshore Mesozoic basins are often not of large amplitude (Costain and others, 1982). High amplitude reflections from below the Scottsville basin are evidence that energy from the vibrator is well-coupled at the surface over the Triassic rocks, and that energy transmission through the Triassic basins is excellent. The window of poor reflection quality associated with the low Triassic impedance contrasts is actually outlining the gross geometry of the Scottsville basin.

Some diffraction patterns associated with the basin were removed by migration. Special reprocessing was done to enhance the data from the Scottsville Triassic basin. A different mute, refinement of stacking velocities, and use of a lower filter passband resulted in the data shown in Figure 7, after time-migration, which shows the lower boundary of the basin. Truncation of eastward-dipping parallel reflections on the right hand side of the figure is taken as evidence of the basin boundary. Sub-horizontal reflections originate from within the Triassic basin. Depth to the bottom of the basin is approximately 1.6 km (0.7 s) with a better defined boundary on the east side. The continuity of the seismic data below the basin precludes downward projection of a steep fault on the west side of the basin without shallowing (see Figure 3 and Figure 7), which suggests that the western boundary is a listric fault. Even with split spread recording geometry with relatively short (35 m) receiver group spacing, wavelet stretching by removal of normal moveout degrades the resolution of desirable high frequency reflections, and thus reflections from the shallow Triassic basin do not have as high a frequency content as desired. Severe muting necessary to eliminate the effect of stretching resulted in low fold and a poor signal-to-noise ratio for the shallow data (< 0.4 s).

Chopawamsic volcanic sequence. The results obtained suggest that 24-fold data are desirable to record accurately the signature of Chopawamsic volcanic rocks (Figure 8 and Figure 9). Line segment NSF-3 as originally recorded, i.e., as 24-fold, 70 m data, is shown in Figure 8. The same line segment resorted by computer and stacked as 12-fold data, is shown in Figure 9. The difference in reflection quality and overall resolution of the seismic signature makes the higher 24-fold data an obvious choice; however, for economic reasons the remainder of the line was recorded as 12-fold data.

One of the best seismic signatures along the traverse is from the Chopawamsic volcanics, as shown in An excellent S/N ratio and well-formed Klauder wavelets with reflections which appear to truncate one another are characteristics of the seismic response. These characteristics may be a manifestation of the stratigraphy as well as isoclinal(?) folding and faulting. A repeated sequence of Chopawamsic volcanic rocks is interpreted to be present, as shown on the line drawing of Figure 4 (eastern part of NSF-3). Reflection quality from the expected contact between the Grenville-Maidens basement and Chopawamsic volcanics is poor. Figure 10 from Line NSF-3 (24-fold, 70 m data, time-migrated).

A portion of stacked section of line NSF-3 is shown in Figure 11. The signature of Chopawamsic volcanic rocks is apparent on the right hand side of the section. Evidence of faulting is present in this section where eastward-dipping reflections are truncated by a set of sub-horizontal high-amplitude reflections. Eastward-dipping reflections between stations 290 and 390 at 1.0 to 2.0 s are interpreted as originating from Chopawamsic metavolcanic rocks. They are not present to the east and west because of the Columbia Granite, leading to the interpretation shown (see Glover and others, in the second part of this report). Below the reflections interpreted as Chopawamsic volcanic rocks, segments of sub-parallel reflections suggest layering. These events might originate from a metamorphosed volcanic sequence other than Chopawamsic; the reflections are interpreted to be from Catoctin metabasalts and sandstones. Although the reflected energy return from Chopawamsic volcanic rocks is excellent we do not believe as mentioned earlier that a prohibitively large decrease in reflected energy from below the Chopawamsic volcanics would result because of the high reflectivity of the Chopawamsic. Below 3.0 s (9 km) the absence of additional reflections is consistent with energy return from Grenville basement in other areas.

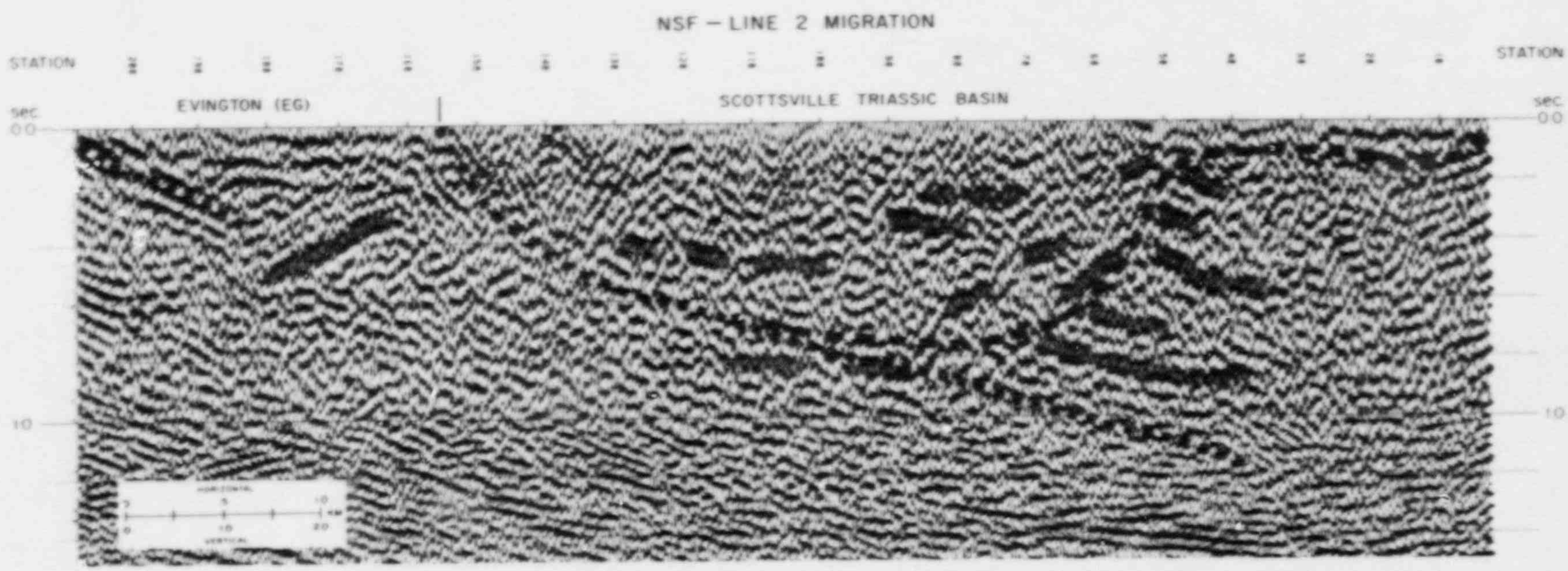


Figure 7. Line NSF-2, migrated: Time-migrated stacked section; reprocessed for shallow reflections from Scottsville Basin. 24-fold and 35 m receiver interval.

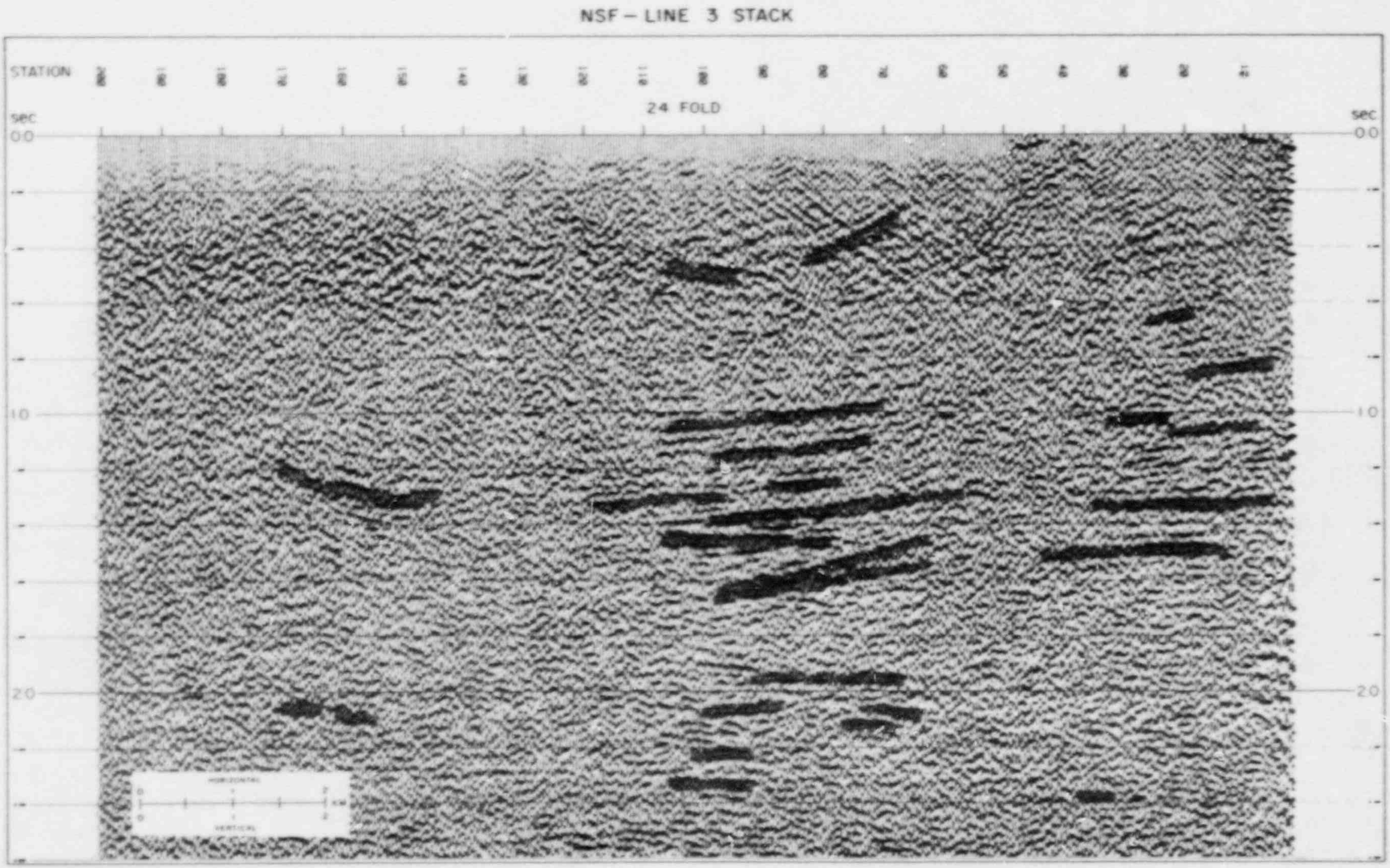


Figure 8. Line NSF-3, 24-fold, stacked: Stacked section; 24-fold and 70 m receiver interval.

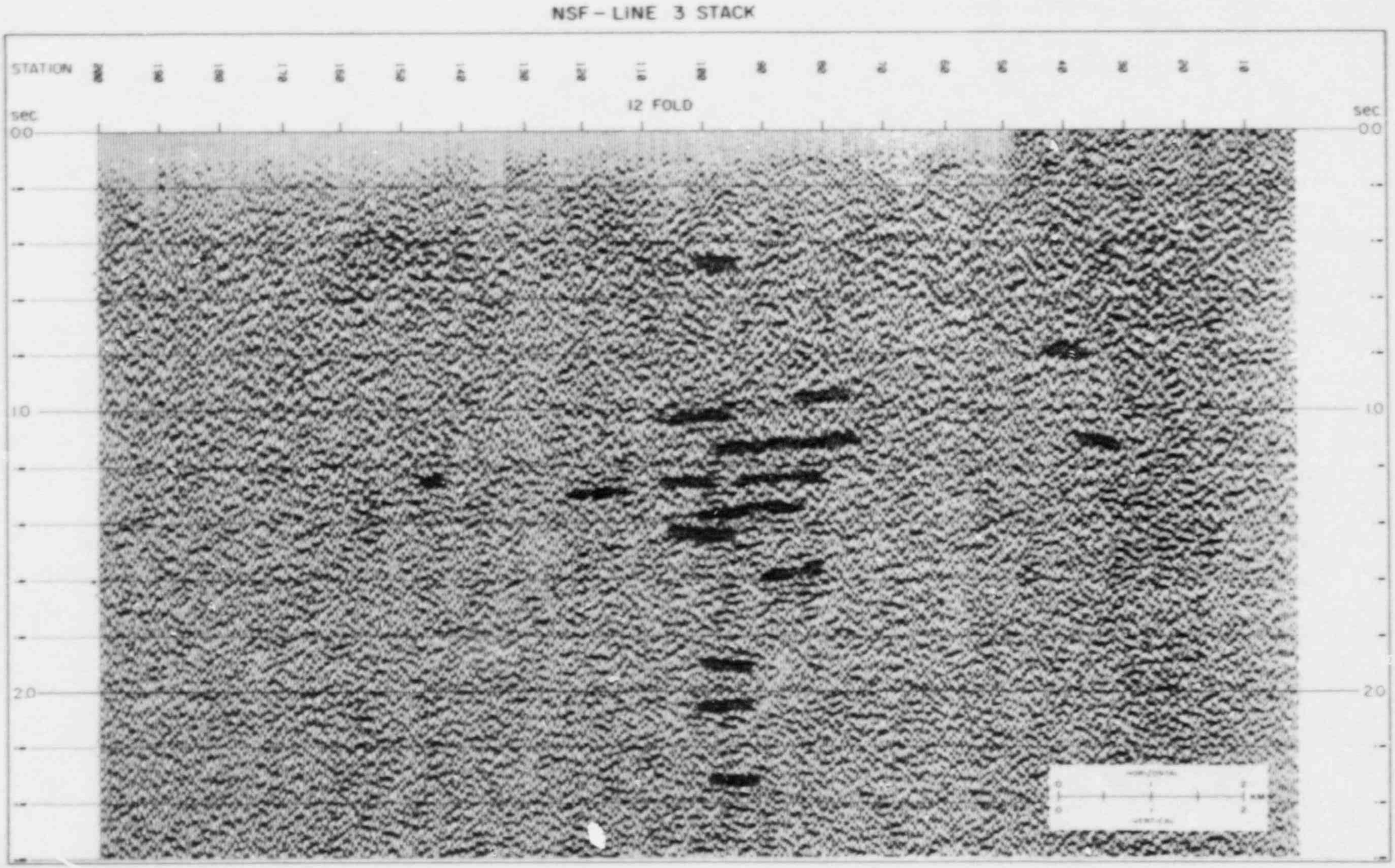


Figure 9. Line NSF-3, 12-fold, stacked: Stack section. Resorted by computer as 12-fold data and 70 m receiver interval.

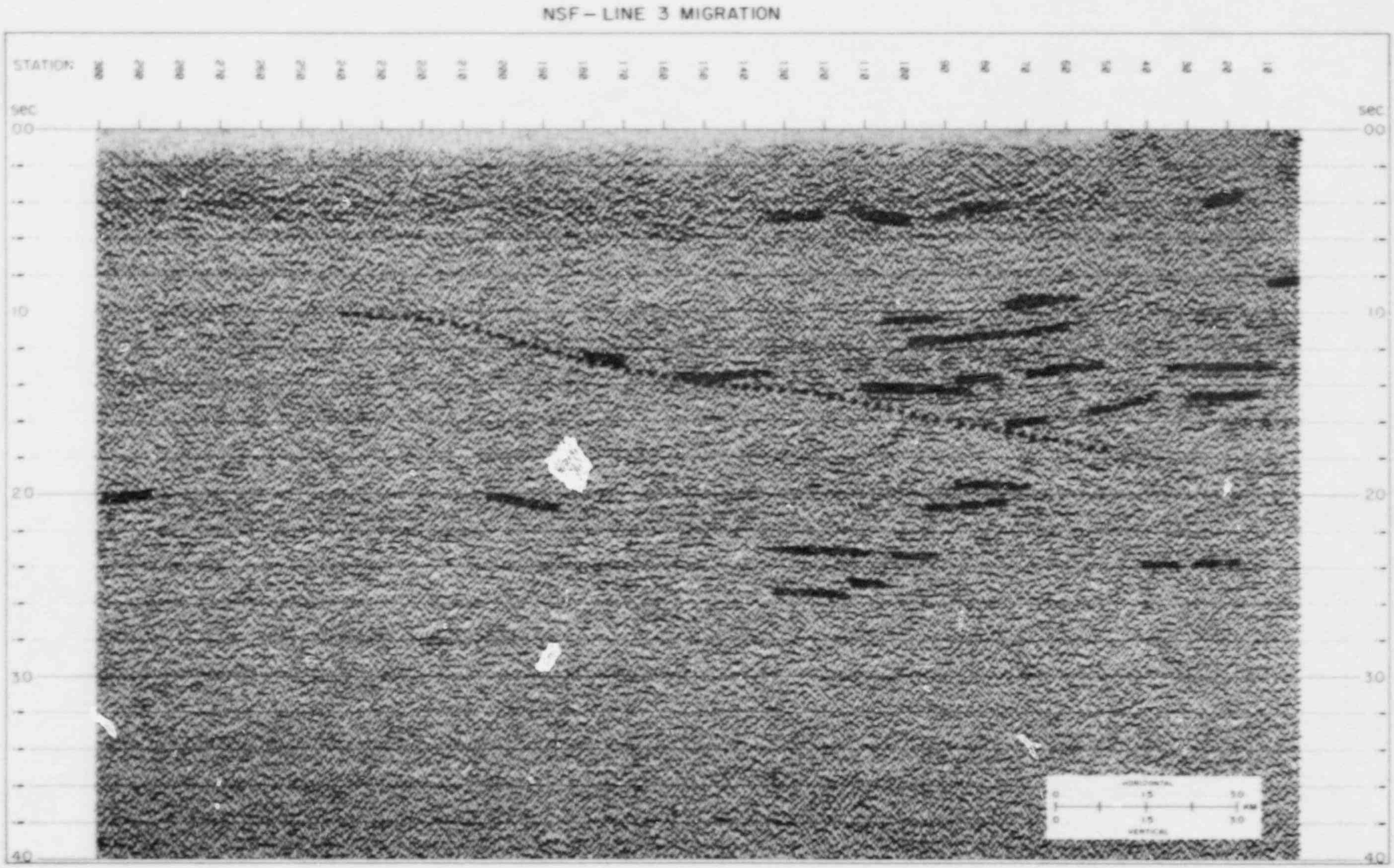


Figure 10. Line NSF-3, migrated: A portion of the time-migrated stacked section; 12-fold, 70 m receiver interval.

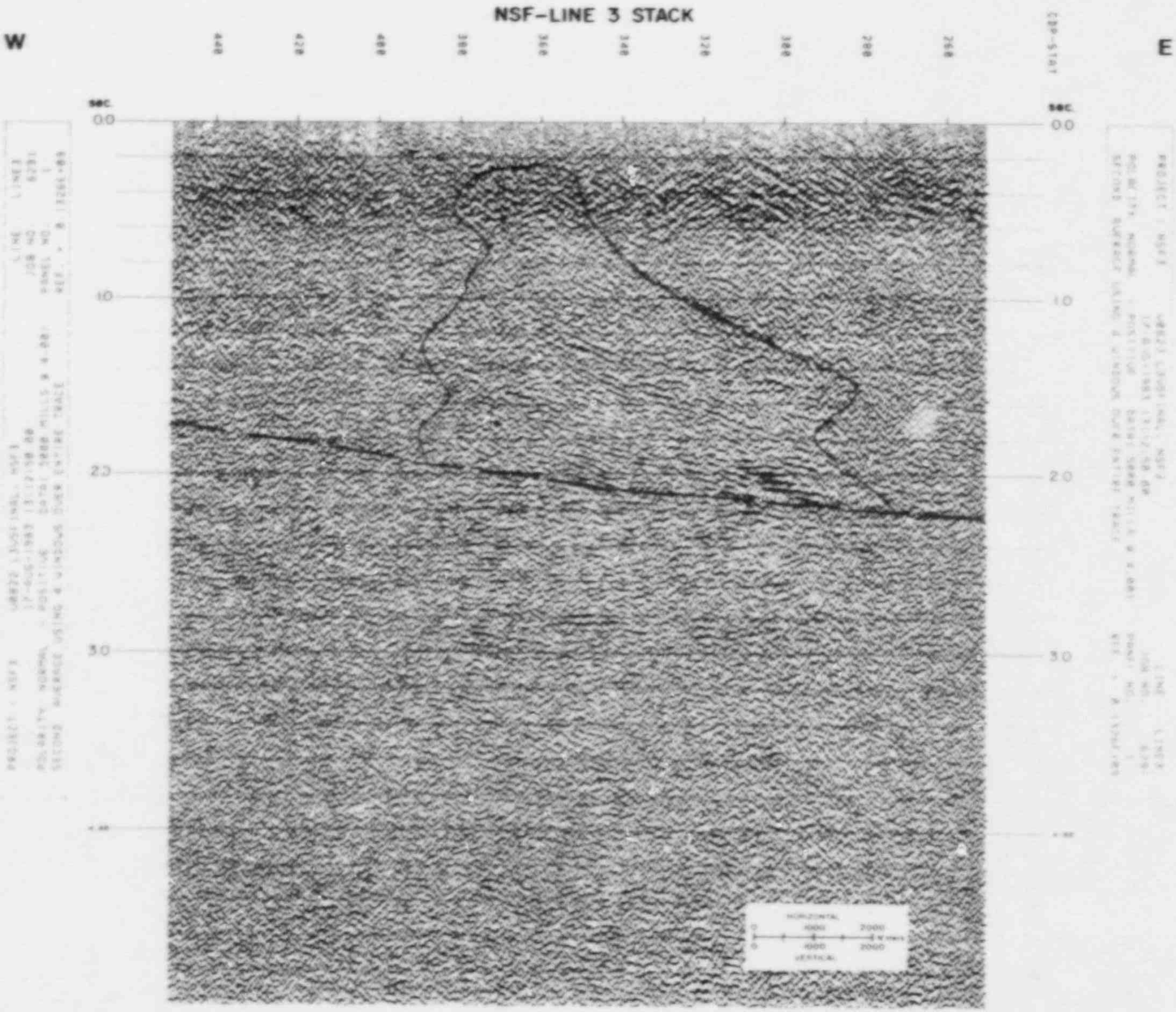


Figure 11. Line NSF-3, stacked: A portion of the stacked section; 12 fold-fold and 70 m receiver interval.

Figure 13 is a stacked section of line segment NRC-4 (12-fold, 70 m) on traverse JRT-2. The reflection quality on this line is excellent with reflections down to 3.5 s (10.5 km). Figure 12 is a time-migrated section of line segment NRC-4, and is a good example of why seismic data should be migrated before final geological interpretation is made. Duplex structures appear to be prominent (between stations 50-120 at 1.2-2.0 s). Low-angle thrust faults are interpreted to refract upward at steep angles when entering packages of more rigid metabasalts (A on Figure 12).

Line NRC-4 offers significant information about seismic profiling of the crust as well as the potential of using a single vibrator:

1. a) Although the multiplicity of this line is only 12-fold, reflections are excellent.

2. b) Reflections down to 3.5 s (10.5 km) show good resolution which allows correlation of events across faults.

3. c) There is no visible decay in reflection amplitude with depth.

One may therefore conclude that even with a single vibrator the return of reflected energy is excellent if a suitable subsurface reflector geometry and impedance contrasts are present. Because of the generally high Q of crystalline rocks the attenuation of energy is much less than in non-crystalline rocks.

Acoustic properties of the Catoctin volcanics

The Catoctin Formation of central Virginia is composed of metamorphosed basalts and sandstones that represent part of the Eocambrian rift volcanics of the eastern continental margin of North America (Wehr and Glover, 1985). On the west side of the Blue Ridge near line NSF-2, granulite facies rocks of 1 Ga Grenville Blue Ridge basement are overlain nonconformably by a thin unit of Eocambrian (?) Swift Run shallow-water clastics and nonmarine Catoctin volcanics metamorphosed to greenschist facies (Figure 1). Nonconformably overlying the Catoctin Formation in the adjacent Valley and Ridge are shallow marine and alluvial clastics of the Cambrian Chilhowee Group (Wehr and Glover, 1985).

Stratigraphic sections of the Catoctin Formation measured at two locations consist of thin beds of metamorphosed, epidotized basalts and sandstones interlayered with non-epidotized, foliated metabasalts and metasediments. The epidotized zones are flow breccias or highly altered zones of epidote and quartz. Gathright (1976) refers to these zones as epidote-amygdaloidal breccia. Basalt flows and rift sediments are also commonly a part of the Catoctin Formation.

Overlying the Lynchburg Formation east of the Blue Ridge are approximately 2 km of Catoctin greenschist-facies metamorphosed basalts interbedded with metasediments. Coarse epidote amygdules are present in outcrop (Wehr and Glover, 1985). The Evington deep-water sequence, a possible equivalent of the Cambro-Ordovician Valley and Ridge Chilhowee shelf sequence, stratigraphically overlies the Catoctin on the east side of the Blue Ridge (Wehr and Glover, 1985).

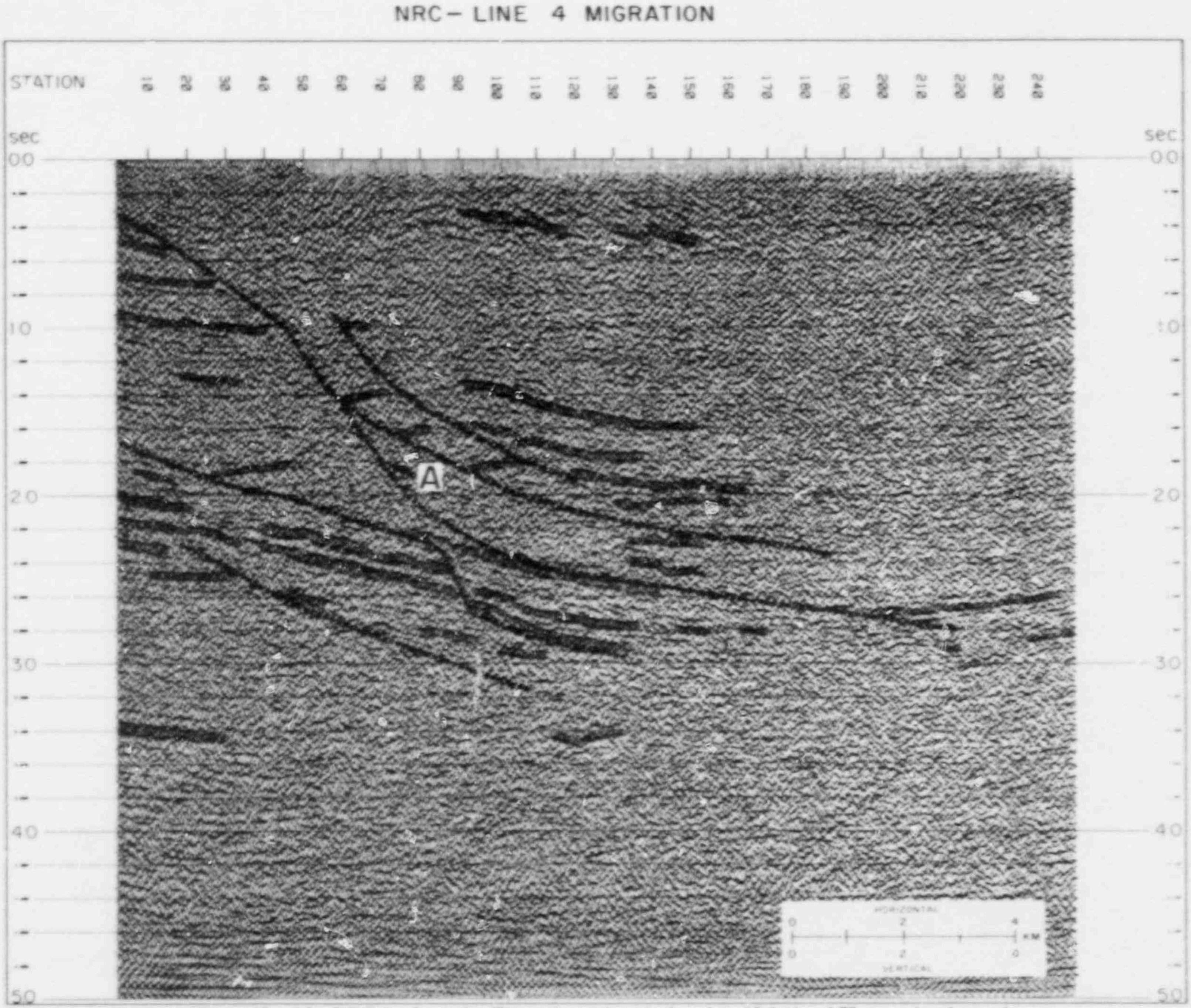


Figure 12. Line NRC-4, migrated: Time-migrated stacked section; 12-fold and 70 m receiver interval.

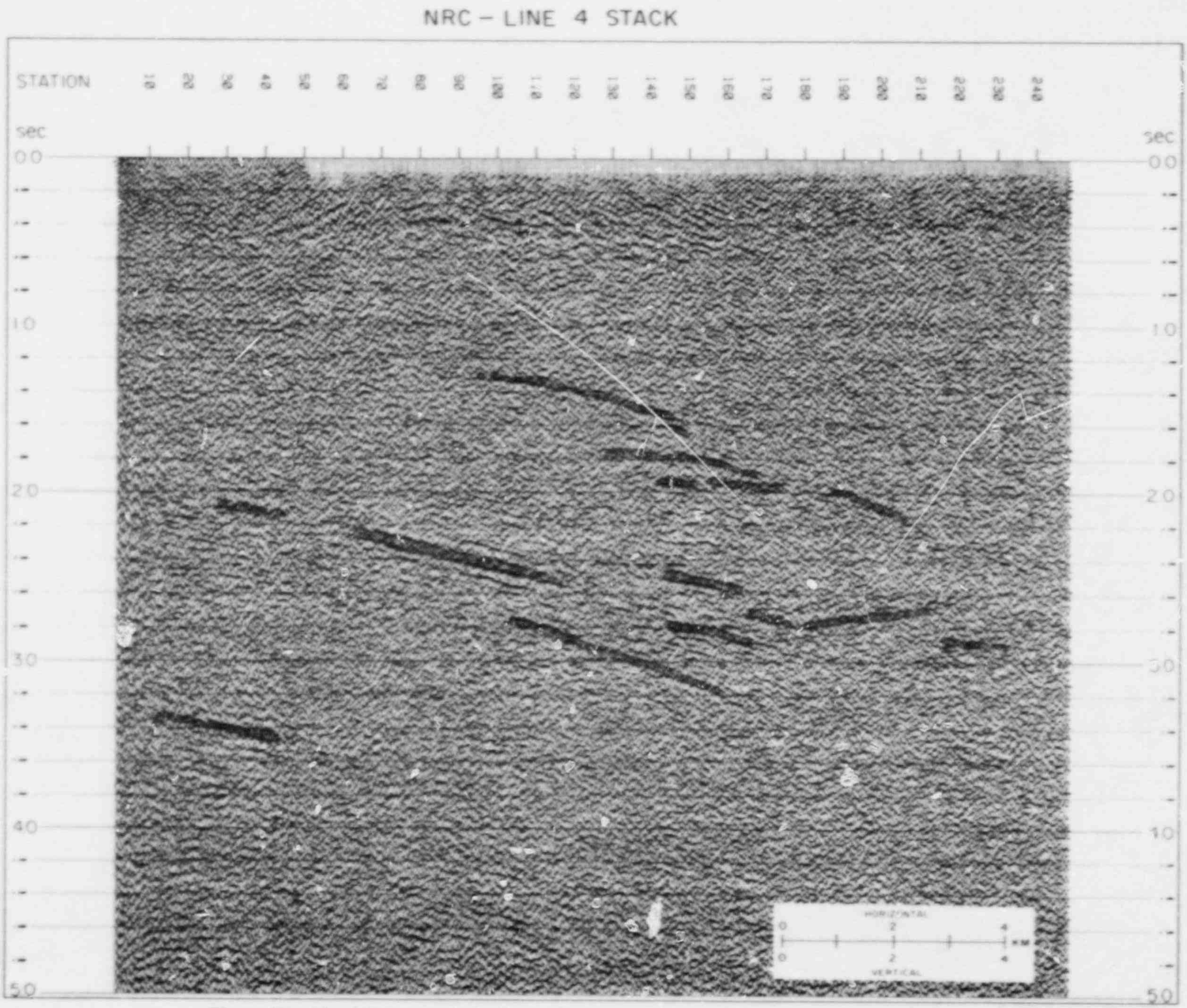


Figure 13. Line NRC-4, stacked: Stacked section; 12-fold and 70 m receiver interval.

The reflections from within the Catoctin Formation on line NSF-2 originate from within a rift sequence believed to have been deposited in a submarine environment.

The excellent reflectivity of the volcanics prompted additional field and laboratory studies of the Catoctin. Stratigraphic sections and sample collection locations are in Brennan (1985). The Catoctin Formation on the east side of the Blue Ridge where the seismic data were obtained is more foliated than that on the west side, and is more easily weathered than the western Catoctin. Samples of the Catoctin were not collected on the eastern side of the Blue Ridge because of the poorer sample quality of the more foliated section.

Compressional velocities parallel and perpendicular to the dominant foliation (when present) in 12 Catoctin Formation samples and one Chilhowee Formation sample were determined in the Regional Geophysics using the method described by Kolich (1974). Samples were collected at locations given in Brennan (1985) and at Luck Stone Quarry in central Virginia. Compressional velocities were determined at pressures of 400 and 600 atm, corresponding to depths of approximately 1.5, and 2.25 km respectively. Velocities determined at 600 atm range from 5.13 km/s to 6.47 km/s for samples of the Catoctin epidosites, greenstones, phyllites, and volcanic breccia (Table 2).

Velocities determined parallel to foliation are higher than those perpendicular to foliation for all of the foliated samples (greenstones). Velocities of the epidosites generally lie between a velocity parallel and perpendicular to the foliation of the greenstones. The difference in velocity from 400 to 600 atm is probably within the accuracy of the velocity determinations. Measurements of core lengths were repeatable to within 1%; traveltime measurements were repeatable to within 4% at 600 atm. Velocities are believed to be accurate to $\pm 5\%$ at 600 atm, and the values at 600 atm were used to calculate reflection coefficients. Velocities and densities of the Catoctin samples are given in Table 2. Velocity versus density for samples with no apparent foliation (epidosites; samples JB4-3C, JB5-5D, JB5-12A, JB5-11A, and JB4-4A) shows that velocity, v(km/s), is approximately a linear function of density, ρ(gm/cm^3), (e.g. Birch, 1961) with a least-squares fit of $v = (1.04 \pm 0.225)\rho + (2.83 \pm 0.696)$, $R = 0.9367$. A least-squares fit between velocities determined perpendicular to S_1 (S_1) and densities of Catoctin greenstones (samples JB5-10C, JB4-3D, JB4-4G, JB4-4D, and JB5-10D) showed an inverse relationship between velocity and density of the form $v_{S1} = -(1.53 \pm 0.513)\rho + (10.5 \pm 1.53)$, $R = -0.86553$. A similar but less well-defined inverse relationship between velocities parallel to S_1 and density (samples JB5-10C, JB4-4G, JB4-4D, JB4-3D, and JB5-10D) was of the form $v_{\|S1} = -(1.39 \pm 0.985)\rho + (10.5 \pm 3.0)$, $R = -0.6295$.

Normal incidence reflection coefficients were calculated for various juxtapositions of Catoctin Formation lithologies. The juxtaposition of Catoctin metabasalts (samples JB4-4A, JB5-11A, JB5-12A, JB5-3C, and JB5-5D) with epidotized metasandstones (JB4-3C) can result in reflection coefficients of magnitudes ≥ 0.1.

Pressure versus acoustic impedance perpendicular and parallel to foliation of the greenstones and epidosites indicate that maximum contrasts result from the juxtaposition of the Catoctin greenstones and the epidosites. A large contrast exists between acoustic impedance of sample JB4-3C and that parallel to the foliation of the greenstone samples. Large contrasts occur between the acoustic impedance of samples JB4-4A, JB5-11A, and JB5-12A and acoustic impedance perpendicular to foliation of the greenstone samples; the reflection coefficient between sample JB4-4A and sample JB5-10C S_1 is -0.08. Acoustic impedance contrasts within the epidosites between sample JB4-3C, a metasandstone, and samples JB4-4A, JB5-11A, and JB5-12A, the amygdaloidal metabasalts, are the largest observed; the reflection coefficient between samples JB4-3C and JB4-4A is 0.12.

Tuning

A thin bed is defined as one whose two-way traveltime thickness is less than the tuning thickness, T/2, of the source wavelet, where T is the dominant period of the wavelet (Sengbush and others, 1961). For a dominant frequency of 38 Hz and a compressional velocity of 6 km/s, the dominant wavelength, λ, is approximately 157 m. The Catoctin beds are commonly less than 30 m thick or less than $\lambda/4$, and are thus acoustically thin. Generally, for beds less than tuning thickness, reflections from the top and bottom are not resolved separately by conventional methods (Kallweit and Wood, 1982; Widess, 1973). At tuning thickness, the reflection begins to take on the shape of the first derivative of the source wavelet (Sengbush and others, 1961). The reflection from a thin bed of tuning thickness is approximately twice the amplitude of the same source wavelet reflected from a single interface. These thin-bed, first-derivative, reflections may themselves constructively interfere to form "tuned" first derivative reflections which are larger in amplitude than would be expected from a single interface, or from a single bed of tuning thickness. The result is a reflected waveform similar in shape to that leaving the source. As more thin beds are added, the reflections take on a reverberating appearance. Using reflection coefficients determined in this study, synthetic seismogram modeling of thin bed sequences was done to obtain seismograms that might simulate the reflection character observed on a segment of Line NSF-2. Typical results are given by Brennan (1985).

Models based on measured Catoctin sections illustrate that constructive interference of reflections from multiple thin beds can result in large-amplitude composite reflections that are characterized by first-derivative waveforms, Klauder-type waveforms, or ringing seismic signatures. Reflections from the synthetic seismogram model of geologic section B are of large amplitude with a slightly reverberating appearance (Brennan, 1985).

Conclusions

Metamorphosed basalts and sandstones (Catoctin) and metamorphosed felsic and mafic volcanic rocks (Chopawamsic) in the upper crust along the James River traverse have excellent reflectivity. Without these reflectors, reflection quality might normally be expected to vary from poor to good in metamorphic and igneous terrane. Our results indicate that, in general, the best crustal reflectivity in the Piedmont is associated with either metamorphosed basalts and felsic volcanics, or with metamorphosed basalts and sandstones. Reflections from the Catoctin Formation of central Virginia are believed to be due in large part to constructive interference from acoustically thin, epidotized layers (epidosites and volcanic breccia) interlayered with non-epidotized layers (greenstones and phyllites), and to a lesser extent exclusively from non-epidotized greenstones against phyllites. The successful definition of the regional geologic framework of seismicity in the crystalline rocks of the Piedmont depends to a large extent on the placement of reflection seismic traverses where thick sequences of such metamorphosed basalts/felsic volcanics, or metamorphosed basalts/sandstones are believed to occur in the subsurface.

Because of the relatively low attenuation coefficients due to high Q, the amplitude of reflected energy does not decay significantly in igneous and metamorphic rocks. Even a single vibrator with a two-octave sweep gives good results. From our data there is no evidence to suggest a depth limitation from a single vibrator provided enough energy is injected into the ground by using long sweeps. Comparison of our data (NRC-5) with multivibrator data obtained about 10 km to the northeast along strike of Catoctin outcrops on I-64 (Harris and others, 1982) indicates that the single vibrator data has a higher S/N ratio. A similar conclusion holds where we were able to make a comparison between proprietary multivibrator data and our data on the same road over the Blue Ridge at the western end of JRT-1.

In general, characteristic seismic signatures of certain metamorphic and igneous rocks can be recognized. The Grenville terrane is acoustically transparent. Metamorphosed volcanic lithofacies have excellent reflectivity. The Chopawamsic volcanics (Figure 11) and Catoctin (Figure 6) have different, recognizable, signatures. The former shows good quality reflections with well-formed Klauder wavelets that truncate one another. They suggest lenticular geometry and repetition by thrust faults. The Catoctin has a signature of large-amplitude, more parallel reflections. Boundaries between granite and basalts or amphibolite seem to result in poor quality reflections. Using the quality and geometry of the reflections in interpretation we obtained a good correlation with mapped surface geology. The geometry of the Scottsville Triassic basin along JRT-1 has been clarified (Figure 7). Good reflected events were recorded on Line NRC-5 beneath the Blue Ridge (Figure 5 and Figure 6) at about 3.0 s (9 km).

Signatures characteristic of reflections which are directly related to the depositional environment have the potential to identify questionable lithofacies below allochthonous plates where extrapolation of reflectors to the surface is not possible. Traverses JRT-1 and JRT-2 confirm that metamorphosed volcanic sequences, which are abundant in the architecture of the Appalachian orogen, and play a key role in evaluating tectonic models, can be easily detected by reflection seismology.

References

Bahorich, M. S., C. Çoruh, E. S. Robinson, and J. K. Costain, 1982, Static corrections on the southeastern Piedmont of the United States, Geophysics, 47, 540-1549.

Birch, F., 1961, The velocity of compressional waves in rocks to 10 kilobars, Part 2, Journal of Geophysical Research, v. 66, No. 7, 2199-2224.

Brennan, J. L., 1985, Interpretation of Vibroseis reflections from within the Catoctin Formation of central Virginia M. S. Thesis in Geophysics, Virginia Polytechnic Institute and State University, Blacksburg, 95 pp.

Clark, H., J. Costain, and L. Glover, III, 1978, Structure and seismic reflection studies on the Brevard ductile deformation zone near Rosman, North Carolina, American Journal of Science, 278, 419-441.

Cook, F., D. Albaugh, L. Brown, S. Kaufman, J. Oliver, and R. Hatcher, Jr., 1979, Thin-skinned tectonics in the crystalline southern Appalachians: COCORP seismic reflection profiling of the Blue Ridge and Piedmont, Geology, 7, 563-567.

Çoruh, C., and J. K. Costain, 1983, Noise attenuation by vibroseis whitening (VSW) processing, Geophysics, 48, 543-554.

Çoruh, C., J. K. Costain, T. L. Pratt, R. D. Hatcher, L. Glover, III, R. A. Phinney, J. B. Diebold, R. T. Williams, M. D. Zoback, and R. N. Anderson, 1985, Preliminary results from seismic-reflection regional lines, Appalachian Deep Drill Hole Project Site Investigation EOS, Transactions Amer. Geophys. Union, v. 66, No. 6, p. 1075.

Costain, J. K., T. L. Pratt, C. Çoruh, L. Glover, III, A. Froelich, and D. Ziegler, 1982, Reflection seismic characteristics of some onshore Triassic Basins in the Southeastern United States, Geological Soc. Am., Abstracts with Programs, 14, 467

Edsall, R. W., 1974, A seismic reflection study over the Bane anticline, M.S. thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 109 pp.

Gathright, T.M., II, 1976, Geology of the Shenandoah National Park, Virginia, Virginia Division of Mineral Resources Bulletin 86, 93p.

Gresko, M. J., and J. K. Costain, 1985, Compressional (P-wave) and Shear (SH-wave) reflection seismic case history over the Bane Dome, Giles County, Virginia, *55th Annual International Meeting Society of Exploration Geophysicists, Expanded Abstracts,* 339-402.

Harris, L.D., and Bayer, K.C., 1979, Sequential development of the Appalachian orogen above a master decollement - A hypothesis: Geology, v. 7, p. 568-572.

Harris, L. D., Harris, A. G., de Witt, Wallace, Jr., and Bayer, K. C., 1981, Evaluation of Southeastern overthrust belt beneath Blue Ridge-Piedmont thrust: American Association of Petroleum Geologists Bulletin, 65, 2497-2505.

Harris, L. D., de Witt, Wallace, Jr., and Bayer, K. C., 1982, Interpretive seismic profile along Interstate I-64 from the Valley and Ridge to the Coastal Plain in central Virginia, U. S. Geological Survey, Oil and Gas Investigations, Chart OC-123.

Kallweit, R. S. and L. C. Wood, 1982, The limits of resolution of zero-phase wavelets, Geophysics, v. 47, No. 7, p. 1035-1046.

Keller, M. R., E. S. Robinson, and L. Glover, III, 1985, Seismicity, seismic reflection, gravity, and geology of the central Virginia seismic zone: Part 3. Gravity, *Bull. Geol. Soc. America*, 1580-1584.

Keller, M.R., E.S. Robinson and L. Glover, III, 1986, Seismicity, seismic reflection, gravity, and geology of the central Virginia seismic zone: Part 3, Gravity. Reply. *Bull. Geol. Soc. America*, v. 97, 1285-1287.

Kolich, T. M., 1974, Seismic reflection and refraction studies in the Folded Valley and Ridge Province at Price Mountain, Montgomery County, Virginia: Virginia Polytechnic Institute and State University, M. S. thesis, 139 p.

Perry, W. J., A. G. Harris, and L. D. Harris, 1979, Conodont-based reinterpretation of Bane-Dome-Structural reevaluation of Allegheny frontal zone, Bull. Amer. Assoc. Pet. Geol., 63, 647-654.

Reilly, J. M., 1980, A geologic and potential field investigation of the central Virginia Piedmont, M. S. Thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 110 pp.

Robinson, E. A., and S. Treitel, 1980. Geophysical Signal Analysis, Prentice-Hall, Inc., Englewood Cliffs, N. J., 466 pp.

Sengbush, R. L., Lawrence, P. L., and McDonal, F. J., 1961, Interpretation of Synthetic Seismograms, Geophysics, v. 26, No. 2, p. 138-157.

Waters, K. H., 1981, Reflection Seismology, John Wiley and Sons, 2nd Edition, 453 pp.

Wehr, F.L. and Glover, L., III, 1985. Stratigraphy and tectonics of the Virginia-North Carolina Blue Ridge Evolution of a Late Proterozoic - Early Paleozoic Hinge Zone, Geological Society of America Bulletin, v. 96, p. 285-295.

Wells, P. D., 1975, A seismic reflection and refraction study near the Blue Ridge thrust in Virginia, M.S. thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 89 pp.

Widess, M. B., 1973, How thin is a Thin Bed?, Geophysics, v. 38, No. 6, 1176-1180.

Table 1

Compressional Velocity and Density of Valley and Ridge and Blue Ridge Rocks[1]

Age	Sample	Sample Number	Velocity (km/s) Pressure (Atm) 400	600	Average Density (g/cm^3)
Mississippian		1-19A	5.06	5.17	
		1-19B	5.03	5.09	
	Price	3-6A	5.48	5.58	2.67 ± 0.03
	sandstone	3-6B	5.53	5.53	
		3-6C	5.50	5.55	
	post-Cloyd	3-7A	4.62	4.70	2.72 ± 0.01
	claystone	3-7B	4.72	4.79	
	Cloyd	1-17A	5.50	5.53	2.58 ± 0.00
	conglomerate	1-17B	5.43	5.49	
	Parrott	1-16A	5.63	5.64	2.62 ± 0.00
	sandstone	1-15A	5.44	5.51	
Devonian		1-14A	4.53	4.64	
	Chemung	1-14B	4.58	4.70	2.61 ± 0.05
	sandstone	1-15A	4.63	4.79	
		1-15B	4.73	4.91	
	Millboro	1-13A	4.72	4.72	2.74 ± 0.01
	shale	1-13B	4.64	4.70	
Silurian	Keefer	1-21A	5.68	5.76	2.64 ± 0.01
	sandstone	1-21B	5.68	5.72	
Silurian		1-12A	5.30	5.31	
	Rose Hill	1-12B	5.32	5.33	3.07 ± 0.03
	sandstone	1-20A	5.29	5.32	
		1-20B	5.68	5.69	
		1-11A	5.79	5.80	
	Tuscarora	1-11B	5.76	5.76	2.64 ± 0.01
	sandstone	2-4A	5.55	5.63	
		2-4B	5.60	5.66	
Ordovician		1-10A	5.18	5.27	
Juniata	1-10B	5.26	5.31		
	sandstone	2-6A	4.51	4.58	2.62 ± 0.03

[1] Velocity and density determinations from Kolich (1974) except where noted.

		2-6B	4.31	4.54	4.66
		3-5A	5.19	5.34	5.42
	Martinsburg shale	3-1A	4.53	4.54	2.70 ± 0.00
		3-1B	4.66	4.83	
	Eggleston conglomerate	3-3A	5.31	5.35	2.65 ± 0.00
	Moccasin shale	1-9A	5.46	5.53	2.71 ± 0.03
		1-9B	5.41	5.48	
		1-9C	5.51	5.55	
		1-9D	5.24	5.30	
		3-4A	3.90	4.09	
		3-4B	3.89	4.09	
Ordovician	Bays sandstone	2-2A	5.59	5.59	2.68 ± 0.01
		2-2B	5.71	5.71	
		2-3A	4.96	4.99	
		2-3B	5.00	5.05	
	Witten limestone	1-6A	5.98	5.99	2.68 ± 0.01
		1-6B	6.06	6.07	
	Liberty Hall shale	2-8A	4.39	4.51	2.69 ± 0.02
		2-8B	4.40	4.41	
		2-10A	4.70	4.81	
		2-10B	4.59	4.61	
		2-10C	4.61	4.63	
		4-1A	5.46	5.49	
		4-1B	5.47	5.56	
		4-1C	5.81	5.82	
		4-2A	4.37	4.51	
		4-2B	4.44	4.60	
		4-2C	4.54	4.70	
		2-1A	6.17	6.22	
		2-1B	6.00	6.01	
	Lincolnshire limestone	1-7B	6.27	6.28	2.70 ± 0.01
		1-8A	6.21	6.22	
		1-8B	6.39	6.40	
		4-13A	6.27	6.32	
		4-13B	6.24	6.29	
Ordovician	Five Oaks limestone	1-5A	6.38	6.38	2.70 ± 0.00
		1-5B	6.37	6.38	
	New Market limestone	4-12A	6.45	6.46	2.72 ± 0.00
		4-12B	6.34	6.29	
	Elway limestone	1-3A	6.00	6.07	2.68 ± 0.01
		1-3B	6.29	6.30	
	Upper Knox	1-2A	6.32	6.40	2.82 ± 0.00

	dolomite	1-2B	6.26	6.31	
	Kingsport dolomite	4-11A	6.02	6.20	2.79 ± 0.00
		4-11B	6.44	6.49	
	Longview limestone	4-10A	6.61	6.62	2.71 ± 0.00
	Chepultepec limestone	4-5A	6.55	6.56	2.72 ± 0.00
		4-5B	6.44	6.45	
	Chepultepec dolomite	4-6A	6.44	6.57	2.83 ± 0.00
		4-6B	6.46	6.65	
Cambrian	Copper Ridge sandstone	4-7A	5.63	5.79	2.71 ± 0.00
		4-7B	5.89	6.01	
	Copper Ridge dolomite	4-8A	6.13	6.32	2.81 ± 0.01
		4-8B	6.21	6.33	
Cambrian	Elbrook dolomite	3-9A	6.23	6.24	2.80 ± 0.03
		3-9B	6.26	6.27	
		4-9A	6.53	6.58	
		4-9B	6.62	6.68	
	Honaker	1-1A	6.92	6.93	2.85 ± 0.01
	Honaker[2] dolomite	6A1	7.3	7.2	2.83 ± 0.01
		6A2	7.1	7.1	
		6B1	7.5	7.5	
		6B2	7.6	7.6	
	Rome[2] shale	5A1	3.7	4.1	2.67 ± 0.02
		5A2	3.7	3.8	
		5A3	3.7	3.9	
		5A4	3.6	3.7	
	Rome[2] dolomite	5B1	7.4	7.3	2.83 ± 0.00
		5B2	7.4	7.4	
	Shady[2] dolomite	4A1	6.7	6.9	2.84 ± 0.01
		4A2	6.7	6.9	
		4A3	6.7	6.9	
	Shady[2] limestone	4B1	7.0	6.9	2.73 ± 0.01
		4B2	6.8	6.8	
Cambrian	Erwin[2]	31A	6.0	6.0	2.59 ± 0.04
		3A2	6.1	6.1	

[2] Samples measured by Edsall (1974).

	sandstone	3B1	5.4	5.4	
		3B2	5.3	5.3	
	Hampton[2] shale	2(1)	5.4	5.4	2.71 ± 0.00
		2(2)	5.5	5.4	
	Unicoi[2] sandstone	1(1)	6.2	6.2	2.67 ± 0.01
		1(2)	5.9	6.0	
Precambrian	Augen[2] gneiss	9(1)	6.1	6.2	2.70 ± 0.00
		9(2)	5.8	6.0	
	Amphibolite[2]	7(1)	5.6	5.7	3.00 ± 0.00
		7(2)	5.6	5.7	
	Lynchburg[2] amphibolite	8C1	6.1	6.1	2.97 ± 0.03
		8C2	6.4	6.4	
	Lynchburg[2] gneiss	8A1	5.4	5.4	
		8A2	5.4	5.4	2.64 ± 0.08
		8B1	4.9	5.0	
		8B2	4.9	4.9	
Precambrian	1.1 billion year old gneiss, Grenville (?)	V105-200	6.05	6.06	
		V105-410	6.06	6.07	2.66 ± 0.04
		V105-550	5.96	6.00	
		V106-803	6.00	6.05	

Table 2

Compressional Velocity and Density of Catoctin Samples

Sample number	Description[3]	Velocity[4] (km/s)	Propagation[5] direction	Density (g/cm^3)
JB4-1	Chilhowee phyllite: 38%Chl, 38%Q + F, 15%Op, 9%Ep	5.29; 5.29	$\perp S_1$	2.79
JB4-3B	Phyllite: 38%Op, 26%Chl, 22%Q, 9%Ep, 4%Sph, 2%Cc	5.14; 5.13	$\perp S_1$	3.15
JB4-3C	Metasandstone: 45%Ep, 45%Q + F, 10%Op; (Epidosite)	5.73; 5.71	no S_1	2.80
JB4-3D	Schist: 47%Ep,28%Chl, 14%Q, 6%Cc, 5%Op (Greenstone)	6.25; 6.28	$\parallel S_1$	2.99
JB4-3D	Schist: 47%Ep, 28%Chl, 14%Q, 6%Cc, 5%Op (Greenstone)	5.98; 5.97	$\perp S_1$	2.95
JB4-4A	Metabasalt: 50%Ep, 17%Chl, 17%Q + F, 8%Cc, 8%Op (Epidosite)	6.23; 6.24	no S_1	3.19
JB4-4D	Schist: 38%Chl, 19%Op, 18%Ep, 5%Cc, 14%Q + F, 6%Sph (Greenstone)	6.38; 6.39	$\parallel S_1$	3.02
JB4-4D	Schist: 38%Chl, 19%Op, 18%Op, 5%Cc, 14%Q + F, 6%Sph (Greenstone)	5.80; 5.80	$\perp S_1$	3.01
JB4-4G	Schist: 45%Chl, 26%Sph, 21%Ep, 4%Q + F, 4%Act (Greenstone)	5.90; 5.92	$\perp S_1$	3.01
JB4-4G	Schist: 45%Chl, 26%Sph, 21%Ep, 4%Q + F, 4%Act (Greenstone)	6.48; 6.47	$\parallel S_1$	3.01

NRC – LINE 4 MIGRATION

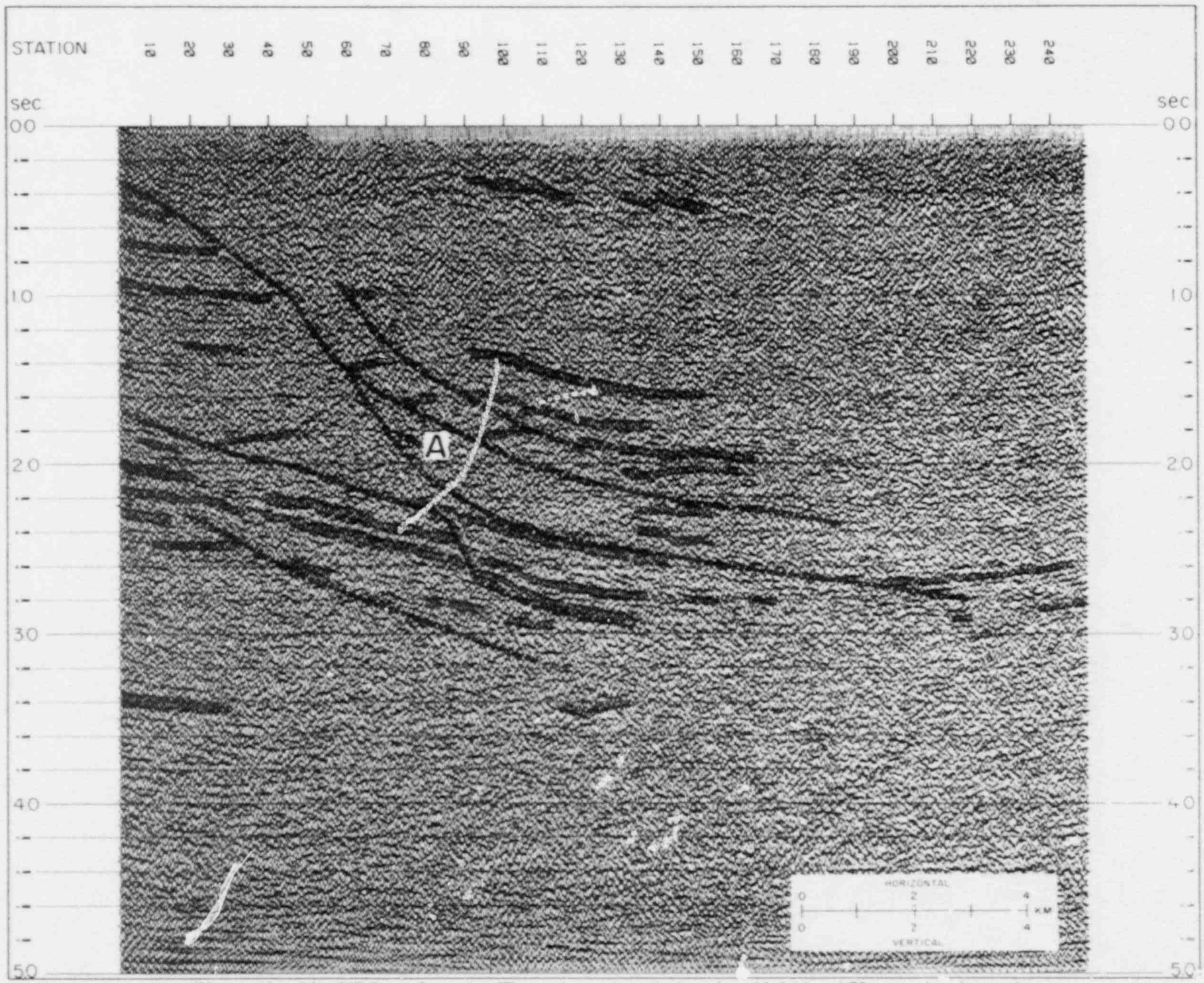


Figure 12. Line NRC-4, migrated: Time-migrated stacked section; 12-fold and 70 m receiver interval.

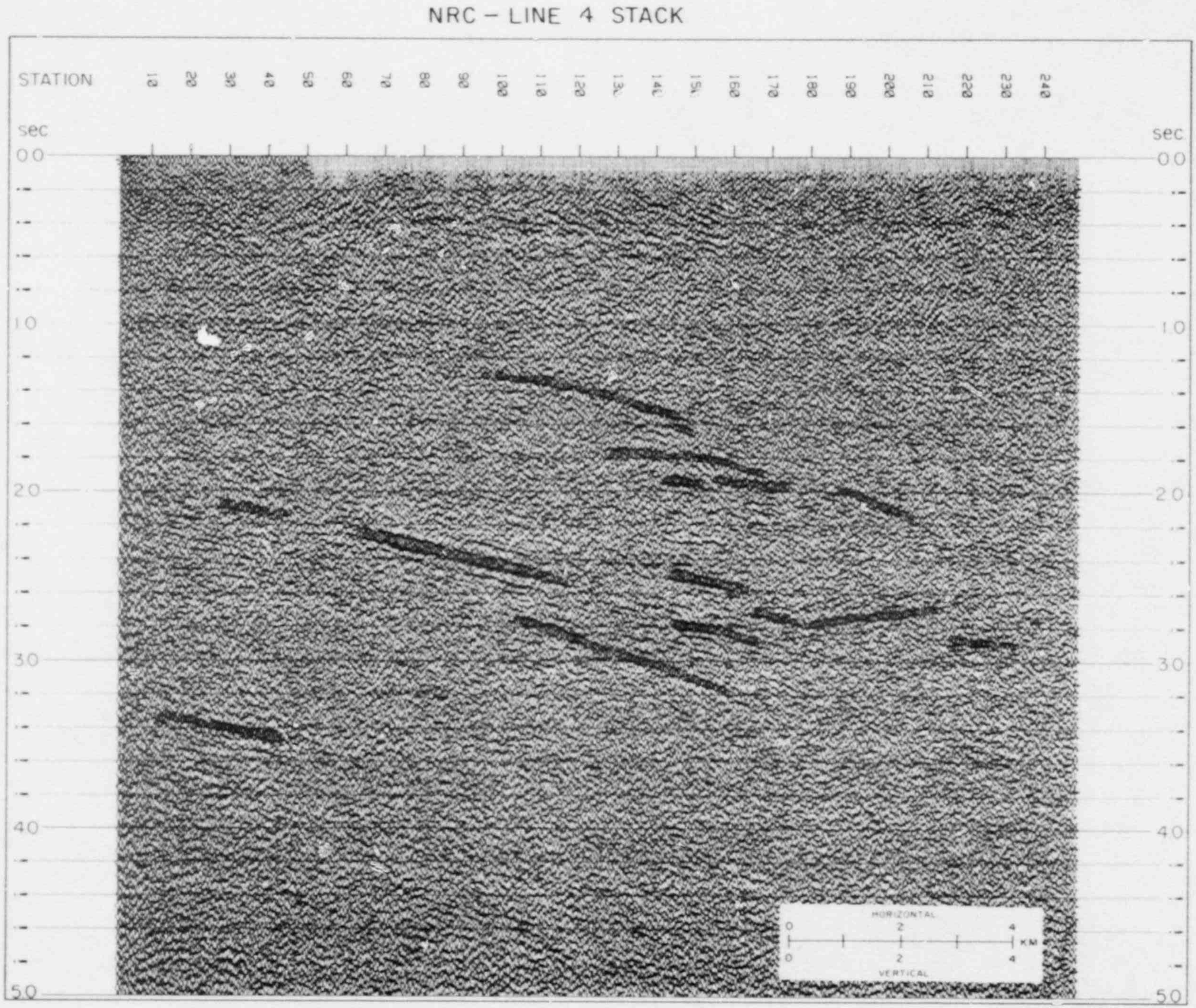


Figure 13. Line NRC-4, stacked: Stacked section; 12-fold and 70 m receiver interval.

The reflections from within the Catoctin Formation on line NSF-2 originate from within a rift sequence believed to have been deposited in a submarine environment.

The excellent reflectivity of the volcanics prompted additional field and laboratory studies of the Catoctin. Stratigraphic sections and sample collection locations are in Brennan (1985). The Catoctin Formation on the east side of the Blue Ridge where the seismic data were obtained is more foliated than that on the west side, and is more easily weathered than the western Catoctin. Samples of the Catoctin were not collected on the eastern side of the Blue Ridge because of the poorer sample quality of the more foliated section.

Compressional velocities parallel and perpendicular to the dominant foliation (when present) in 12 Catoctin Formation samples and one Chilhowee Formation sample were determined in the Regional Geophysics using the method described by Kolich (1974). Samples were collected at locations given in Brennan (1985) and at Luck Stone Quarry in central Virginia. Compressional velocities were determined at pressures of 400 and 600 atm, corresponding to depths of approximately 1.5, and 2.25 km respectively. Velocities determined at 600 atm range from 5.13 km/s to 6.47 km/s for samples of the Catoctin epidosites, greenstones, phyllites, and volcanic breccia (Table 2).

Velocities determined parallel to foliation are higher than those perpendicular to foliation for all of the foliated samples (greenstones). Velocities of the epidosites generally lie between a velocity parallel and perpendicular to the foliation of the greenstones. The difference in velocity from 400 to 600 atm is probably within the accuracy of the velocity determinations. Measurements of core lengths were repeatable to within 1%; traveltime measurements were repeatable to within 4% at 600 atm. Velocities are believed to be accurate to $\pm 5\%$ at 600 atm, and the values at 600 atm were used to calculate reflection coefficients. Velocities and densities of the Catoctin samples are given in Table 2. Velocity versus density for samples with no apparent foliation (epidosites; samples JB4-3C, JB5-5D, JB5-12A, JB5-11A, and JB4-4A) shows that velocity, v(km/s), is approximately a linear function of density, $\rho(\mathrm{gm/cm^3})$, (e.g. Birch, 1961) with a least-squares fit of $v = (1.04 \pm 0.225)\rho + (2.83 \pm 0.696)$, $R = 0.9367$. A least-squares fit between velocities determined perpendicular to S_1 (S_1) and densities of Catoctin greenstones (samples JB5-10C, JB4-3D, JB4-4G, JB4-4D, and JB5-10D) showed an inverse relationship between velocity and density of the form $v_{S1} = -(1.53 \pm 0.513)\rho + (10.5 \pm 1.53)$, $R = -0.86553$. A similar but less well-defined inverse relationship between velocities parallel to S_1 and density (samples JB5-10C, JB4-4G, JB4-4D, JB4-3D, and JB5-10D) was of the form $v_{\parallel S1} = -(1.39 \pm 0.985)\rho + (10.5 \pm 3.0)$, $R = -0.6295$.

Normal incidence reflection coefficients were calculated for various juxtapositions of Catoctin Formation lithologies. The juxtaposition of Catoctin metabasalts (samples JB4-4A, JB5-11A, JB5-12A, JB5-3C, and JB5-5D) with epidotized metasandstones (JB4-3C) can result in reflection coefficients of magnitudes ≥ 0.1.

Pressure versus acoustic impedance perpendicular and parallel to foliation of the greenstones and epidosites indicate that maximum contrasts result from the juxtaposition of the Catoctin greenstones and the epidosites. A large contrast exists between acoustic impedance of sample JB4-3C and that parallel to the foliation of the greenstone samples. Large contrasts occur between the acoustic impedance of samples JB4-4A, JB5-11A, and JB5-12A and acoustic impedance perpendicular to foliation of the greenstone samples; the reflection coefficient between sample JB4-4A and sample JB5-10C S_1 is -0.08. Acoustic impedance contrasts within the epidosites between sample JB4-3C, a metasandstone, and samples JB4-4A, JB5-11A, and JB5-12A, the amygdaloidal metabasalts, are the largest observed; the reflection coefficient between samples JB4-3C and JB4-4A is 0.12.

Tuning

A thin bed is defined as one whose two-way traveltime thickness is less than the tuning thickness, T/2, of the source wavelet, where T is the dominant period of the wavelet (Sengbush and others, 1961). For a dominant frequency of 38 Hz and a compressional velocity of 6 km/s, the dominant wavelength, λ, is approximately 157 m. The Catoctin beds are commonly less than 30 m thick or less than $\lambda/4$, and are thus acoustically thin. Generally, for beds less than tuning thickness, reflections from the top and bottom are not resolved separately by conventional methods (Kallweit and Wood, 1982; Widess, 1973). At tuning thickness, the reflection begins to take on the shape of the first derivative of the source wavelet (Sengbush and others, 1961). The reflection from a thin bed of tuning thickness is approximately twice the amplitude of the same source wavelet reflected from a single interface. These thin-bed, first-derivative, reflections may themselves constructively interfere to form "tuned" first derivative reflections which are larger in amplitude than would be expected from a single interface, or from a single bed of tuning thickness. The result is a reflected waveform similar in shape to that leaving the source. As more thin beds are added, the reflections take on a reverberating appearance. Using reflection coefficients determined in this study, synthetic seismogram modeling of thin bed sequences was done to obtain seismograms that might simulate the reflection character observed on a segment of Line NSF-2. Typical results are given by Brennan (1985).

Models based on measured Catoctin sections illustrate that constructive interference of reflections from multiple thin beds can result in large-amplitude composite reflections that are characterized by first-derivative waveforms, Klauder-type waveforms, or ringing seismic signatures. Reflections from the synthetic seismogram model of geologic section B are of large amplitude with a slightly reverberating appearance (Brennan, 1985).

Conclusions

Metamorphosed basalts and sandstones (Catoctin) and metamorphosed felsic and mafic volcanic rocks (Chopawamsic) in the upper crust along the James River traverse have excellent reflectivity. Without these reflectors, reflection quality might normally be expected to vary from poor to good in metamorphic and igneous terrane. Our results indicate that, in general, the best crustal reflectivity in the Piedmont is associated with either metamorphosed basalts and felsic volcanics, or with metamorphosed basalts and sandstones. Reflections from the Catoctin Formation of central Virginia are believed to be due in large part to constructive interference from acoustically thin, epidotized layers (epidosites and volcanic breccia) interlayered with non-epidotized layers (greenstones and phyllites), and to a lesser extent exclusively from non-epidotized greenstones against phyllites. The successful definition of the regional geologic framework of seismicity in the crystalline rocks of the Piedmont depends to a large extent on the placement of reflection seismic traverses where thick sequences of such metamorphosed basalts/felsic volcanics, or metamorphosed basalts/sandstones are believed to occur in the subsurface.

Because of the relatively low attenuation coefficients due to high Q, the amplitude of reflected energy does not decay significantly in igneous and metamorphic rocks. Even a single vibrator with a two-octave sweep gives good results. From our data there is no evidence to suggest a depth limitation from a single vibrator provided enough energy is injected into the ground by using long sweeps. Comparison of our data (NRC-5) with multivibrator data obtained about 10 km to the northeast along strike of Catoctin outcrops on I-64 (Harris and others, 1982) indicates that the single vibrator data has a higher S/N ratio. A similar conclusion holds where we were able to make a comparison between proprietary multivibrator data and our data on the same road over the Blue Ridge at the western end of JRT-1.

In general, characteristic seismic signatures of certain metamorphic and igneous rocks can be recognized. The Grenville terrane is acoustically transparent. Metamorphosed volcanic lithofacies have excellent reflectivity. The Chopawamsic volcanics (Figure 11) and Catoctin (Figure 6) have different, recognizable, signatures. The former shows good quality reflections with well-formed Klauder wavelets that truncate one another. They suggest lenticular geometry and repetition by thrust faults. The Catoctin has a signature of large-amplitude, more parallel reflections. Boundaries between granite and basalts or amphibolite seem to result in poor quality reflections. Using the quality and geometry of the reflections in interpretation we obtained a good correlation with mapped surface geology. The geometry of the Scottsville Triassic basin along JRT-1 has been clarified (Figure 7). Good reflected events were recorded on Line NRC-5 beneath the Blue Ridge (Figure 5 and Figure 6) at about 3.0 s (9 km).

Signatures characteristic of reflections which are directly related to the depositional environment have the potential to identify questionable lithofacies below allochthonous plates where extrapolation of reflectors to the surface is not possible. Traverses JRT-1 and JRT-2 confirm that metamorphosed volcanic sequences, which are abundant in the architecture of the Appalachian orogen, and play a key role in evaluating tectonic models, can be easily detected by reflection seismology.

References

Bahorich, M. S., C. Çoruh, E. S. Robinson, and J. K. Costain, 1982, Static corrections on the southeastern Piedmont of the United States, Geophysics, 47, 540-1549.

Birch, F., 1961, The velocity of compressional waves in rocks to 10 kilobars. Part 2, Journal of Geophysical Research, v. 66, No. 7, 2199-2224.

Brennan, J. L., 1985, Interpretation of Vibroseis reflections from within the Catoctin Formation of central Virginia, M. S. Thesis in Geophysics, Virginia Polytechnic Institute and State University, Blacksburg, 95 pp.

Clark, H., J. Costain, and L. Glover, III, 1978, Structure and seismic reflection studies on the Brevard ductile deformation zone near Rosman, North Carolina, American Journal of Science, 278, 419-441.

Cook, F., D. Albaugh, L. Brown, S. Kaufman, J. Oliver, and R. Hatcher, Jr., 1979, Thin-skinned tectonics in the crystalline southern Appalachians: COCORP seismic reflection profiling of the Blue Ridge and Piedmont, Geology, 7, 563-567.

Çoruh, C., and J. K. Costain, 1983, Noise attenuation by vibroseis whitening (VSW) processing, Geophysics, 48, 543-554.

Çoruh, C., J. K. Costain, T. L. Pratt, R. D. Hatcher, L. Glover, III, R. A. Phinney, J. B. Diebold, R. T. Williams, M. D. Zoback, and R. N. Anderson, 1985, Preliminary results from seismic-reflection regional lines, Appalachian Deep Drill Hole Project Site Investigation, EOS, Transactions Amer. Geophys. Union, v. 66, No. 6, p. 1075.

Costain, J. K., T. L. Pratt, C. Çoruh, L. Glover, III, A. Froelich, and D. Ziegler, 1982, Reflection seismic characteristics of some onshore Triassic Basins in the Southeastern United States, Geological Soc. Am., Abstracts with Programs, 14, 467

Edsall, R. W., 1974, A seismic reflection study over the Bane anticline, M.S. thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 109 pp.

Gathright, T.M., II, 1976, Geology of the Shenandoah National Park, Virginia, Virginia Division of Mineral Resources Bulletin 86, 93p.

Gresko, M. J., and J. K. Costain, 1985, Compressional (P-wave) and Shear (SH-wave) reflection seismic case history over the Bane Dome, Giles County, Virginia, *55th Annual International Meeting Society of Exploration Geophysicists, Expanded Abstracts*, 339-402.

Harris, L.D., and Bayer, K.C., 1979, Sequential development of the Appalachian orogen above a master decollement - A hypothesis: Geology, v. 7, p. 568-572.

Harris, L. D., Harris, A. G., de Witt, Wallace, Jr., and Bayer, K. C., 1981, Evaluation of Southeastern overthrust belt beneath Blue Ridge-Piedmont thrust: American Association of Petroleum Geologists Bulletin, 65, 2497-2505.

Harris, L. D., de Witt, Wallace, Jr., and Bayer, K. C., 1982, Interpretive seismic profile along Interstate I-64 from the Valley and Ridge to the Coastal Plain in central Virginia, U. S. Geological Survey, Oil and Gas Investigations, Chart OC-123.

Kallweit, R. S. and L. C. Wood, 1982, The limits of resolution of zero-phase wavelets, Geophysics, v. 47, No. 7, p. 1035-1046.

Keller, M. R., E. S. Robinson, and L. Glover, III, 1985, Seismicity, seismic reflection, gravity, and geology of the central Virginia seismic zone: Part 3. Gravity, *Bull. Geol. Soc. America*, 1580-1584.

Keller, M.R., E.S. Robinson and L. Glover, III, 1986, Seismicity, seismic reflection, gravity, and geology of the central Virginia seismic zone: Part 3, Gravity. Reply. *Bull. Geol. Soc. America*, v. 97, 1285-1287.

Kolich, T. M., 1974, Seismic reflection and refraction studies in the Folded Valley and Ridge Province at Price Mountain, Montgomery County, Virginia: Virginia Polytechnic Institute and State University, M. S. thesis, 139 p.

Perry, W. J., A. G. Harris, and L. D. Harris, 1979, Conodont-based reinterpretation of Bane-Dome-Structural reevaluation of Allegheny frontal zone, Bull. Amer. Assoc. Pet. Geol., 63, 647-654.

Reilly, J. M., 1980, A geologic and potential field investigation of the central Virginia Piedmont, M. S. Thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 110 pp.

Robinson, E. A., and S. Treitel, 1980, Geophysical Signal Analysis, Prentice-Hall, Inc., Englewood Cliffs, N. J., 466 pp.

Sengbush, R. L., Lawrence, P. L., and McDonal, F. J., 1961, Interpretation of Synthetic Seismograms, Geophysics, v. 26, No. 2, p. 138-157.

Waters, K. H., 1981, Reflection Seismology, John Wiley and Sons, 2nd Edition, 453 pp.

Wehr, F.L. and Glover, L., III, 1985. Stratigraphy and tectonics of the Virginia-North Carolina Blue Ridge Evolution of a Late Proterozoic - Early Paleozoic Hinge Zone, Geological Society of America Bulletin, v. 96, p. 285-295.

Wells, P. D., 1975, A seismic reflection and refraction study near the Blue Ridge thrust in Virginia, M.S. thesis, Department of Geological Sciences, Virginia Polytechnic Institute and State University, Blacksburg, VA, 89 pp.

Widess, M. B., 1973, How thin is a Thin Bed?, Geophysics, v. 38, No. 6, 1176-1180.

Table 1

Compressional Velocity and Density of Valley and Ridge and Blue Ridge Rocks[1]

Age	Sample	Sample Number	Velocity (km/s) Pressure (Atm) 400	600	Average Density (g/cm^3)
Mississippian		1-19A	5.06	5.17	
		1-19B	5.03	5.09	
	Price	3-6A	5.48	5.58	2.67 ± 0.03
	sandstone	3-6B	5.53	5.53	
		3-6C	5.50	5.55	
	post-Cloyd	3-7A	4.62	4.70	2.72 ± 0.01
	claystone	3-7B	4.72	4.79	
	Cloyd	1-17A	5.50	5.53	2.58 ± 0.00
	conglomerate	1-17B	5.43	5.49	
	Parrott	1-16A	5.63	5.64	2.62 ± 0.00
	sandstone	1-15A	5.44	5.51	
Devonian		1-14A	4.53	4.64	
	Chemung	1-14B	4.58	4.70	2.61 ± 0.05
	sandstone	1-15A	4.63	4.79	
		1-15B	4.73	4.91	
	Millboro	1-13A	4.72	4.72	2.74 ± 0.01
	shale	1-13B	4.64	4.70	
Silurian	Keefer	1-21A	5.68	5.76	2.64 ± 0.01
	sandstone	1-21B	5.68	5.72	
Silurian		1-12A	5.30	5.31	
	Rose Hill	1-12B	5.32	5.33	3.07 ± 0.03
	sandstone	1-20A	5.29	5.32	
		1-20B	5.68	5.69	
		1-11A	5.79	5.80	
	Tuscarora	1-11B	5.76	5.76	2.64 ± 0.01
	sandstone	2-4A	5.55	5.63	
		2-4B	5.60	5.66	
Ordovician		1-10A	5.18	5.27	
Juniata	1-10B	5.26	5.31		
	sandstone	2-6A	4.51	4.58	2.62 ± 0.03

[1] Velocity and density determinations from Kolich (1974) except where noted.

		2-6B	4.31	4.54	4.66
		3-5A	5.19	5.34	5.42
	Martinsburg shale	3-1A	4.53	4.54	2.70 ± 0.00
		3-1B	4.66	4.83	
	Eggleston conglomerate	3-3A	5.31	5.35	2.65 ± 0.00
		1-9A	5.46	5.53	
		1-9B	5.41	5.48	
	Moccasin shale	1-9C	5.51	5.55	2.71 ± 0.03
		1-9D	5.24	5.30	
		3-4A	3.90	4.09	
		3-4B	3.89	4.09	
Ordovician		2-2A	5.59	5.59	
	Bays sandstone	2-2B	5.71	5.71	2.68 ± 0.01
		2-3A	4.96	4.99	
		2-3B	5.00	5.05	
	Witten limestone	1-6A	5.98	5.99	2.68 ± 0.01
		1-6B	6.06	6.07	
		2-8A	4.39	4.51	
		2-8B	4.40	4.41	
		2-10A	4.70	4.81	
		2-10B	4.59	4.61	
		2-10C	4.61	4.63	
	Liberty Hall shale	4-1A	5.46	5.49	
		4-1B	5.47	5.56	2.69 ± 0.02
		4-1C	5.81	5.82	
		4-2A	4.37	4.51	
		4-2B	4.44	4.60	
		4-2C	4.54	4.70	
		2-1A	6.17	6.22	
		2-1B	6.00	6.01	
		1-7B	6.27	6.28	
		1-8A	6.21	6.22	
	Lincolnshire limestone	1-8B	6.39	6.40	2.70 ± 0.01
		4-13A	6.27	6.32	
		4-13B	6.24	6.29	
Ordovician	Five Oaks limestone	1-5A	6.38	6.38	2.70 ± 0.00
		1-5B	6.37	6.38	
	New Market limestone	4-12A	6.45	6.46	2.72 ± 0.00
		4-12B	6.34	6.29	
	Elway limestone	1-3A	6.00	6.07	2.68 ± 0.01
		1-3B	6.29	6.30	
	Upper Knox	1-2A	6.32	6.40	2.82 ± 0.00

	dolomite	1-2B	6.26	6.31	
	Kingsport	4-11A	6.02	6.20	2.79 ± 0.00
	dolomite	4-11B	6.44	6.49	
	Longview limestone	4-10A	6.61	6.62	2.71 ± 0.00
	Chepultepec	4-5A	6.55	6.56	2.72 ± 0.00
	limestone	4-5B	6.44	6.45	
	Chepultepec	4-6A	6.44	6.57	2.83 ± 0.00
	dolomite	4-6B	6.46	6.65	
Cambrian	Copper Ridge	4-7A	5.63	5.79	2.71 ± 0.00
	sandstone	4-7B	5.89	6.01	
	Copper Ridge	4-8A	6.13	6.32	2.81 ± 0.01
	dolomite	4-8B	6.21	6.33	
Cambrian		3-9A	6.23	6.24	
	Elbrook	3-9B	6.26	6.27	2.80 ± 0.03
	dolomite	4-9A	6.53	6.58	
		4-9B	6.62	6.68	
	Honaker	1-1A	6.92	6.93	2.85 ± 0.01
		6A1	7.3	7.2	
	Honaker[2]	6A2	7.1	7.1	2.83 ± 0.01
	dolomite	6B1	7.5	7.5	
		6B2	7.6	7.6	
		5A1	3.7	4.1	
	Rome[2]	5A2	3.7	3.8	2.67 ± 0.02
	shale	5A3	3.7	3.9	
		5A4	3.6	3.7	
	Rome[2]	5B1	7.4	7.3	2.83 ± 0.00
	dolomite	5B2	7.4	7.4	
		4A1	6.7	6.9	
	Shady[2]	4A2	6.7	6.9	2.84 ± 0.01
	dolomite	4A3	6.7	6.9	
	Shady[2]	4B1	7.0	6.9	2.73 ± 0.01
	limestone	4B2	6.8	6.8	
Cambrian		31A	6.0	6.0	
	Erwin[2]	3A2	6.1	6.1	2.59 ± 0.04

[2] Samples measured by Edsall (1974).

	sandstone	3B1	5.4	5.4	
		3B2	5.3	5.3	
	Hampton[2]	2(1)	5.4	5.4	2.71 ± 0.00
	shale	2(2)	5.5	5.4	
	Unicoi[2]	1(1)	6.2	6.2	2.67 ± 0.01
	sandstone	1(2)	5.9	6.0	
Precambrian	Augen[2]	9(1)	6.1	6.2	2.70 ± 0.00
	gneiss	9(2)	5.8	6.0	
	Amphibolite[2]	7(1)	5.6	5.7	3.00 ± 0.00
		7(2)	5.6	5.7	
	Lynchburg[2]	8C1	6.1	6.1	2.97 ± 0.03
	amphibolite	8C2	6.4	6.4	
		8A1	5.4	5.4	
	Lynchburg[2]	8A2	5.4	5.4	2.64 ± 0.08
	gneiss	8B1	4.9	5.0	
		8B2	4.9	4.9	
Precambrian	1.1 billion	V105-200	6.05	6.06	
	year old	V105-410	6.06	6.07	2.66 ± 0.04
	gneiss,	V105-550	5.96	6.00	
	Grenville (?)	V106-803	6.00	6.05	

Table 2

Compressional Velocity and Density of Catoctin Samples

Sample number	Description[3]	Velocity[4] (km/s)	Propagation[5] direction	Density (g/cm^3)
JB4-1	Chilhowee phyllite: 38%Chl, 38%Q + F, 15%Op, 9%Ep	5.29; 5.29	$\perp S_1$	2.79
JB4-3B	Phyllite: 38%Op, 26%Chl, 22%Q, 9%Ep, 4%Sph, 2%Cc	5.14; 5.13	$\perp S_1$	3.15
JB4-3C	Metasandstone: 45%Ep, 45%Q + F, 10%Op; (Epidosite)	5.73; 5.71	no S_1	2.80
JB4-3D	Schist: 47%Ep,28%Chl, 14%Q, 6%Cc, 5%Op (Greenstone)	6.25; 6.28	$\parallel S_1$	2.99
JB4-3D	Schist: 47%Ep, 28%Chl, 14%Q, 6%Cc, 5%Op (Greenstone)	5.98; 5.97	$\perp S_1$	2.95
JB4-4A	Metabasalt: 50%Ep, 17%Chl, 17%Q + F, 8%Cc, 8%Op (Epidosite)	6.23; 6.24	no S_1	3.19
JB4-4D	Schist: 38%Chl 19%Op, 18%Ep, 5%Cc, 14%Q + F, 6%Sph (Greenstone)	6.38; 6.39	$\parallel S_1$	3.02
JB4-4D	Schist: 38%Chl, 19%Op, 18%Op, 5%Cc, 14%Q + F, 6%Sph (Greenstone)	5.80; 5.80	$\perp S_1$	3.01
JB4-4G	Schist: 45%Chl, 26%Sph, 21%Ep, 4%Q + F, 4%Act (Greenstone)	5.90; 5.92	$\perp S_1$	3.01
JB4-4G	Schist: 45%Chl, 26%Sph, 21%Ep, 4%Q + F, 4%Act (Greenstone)	6.48; 6.47	$\parallel S_1$	3.01

JB5-5D	Metabasalt: 38%Ep, 29%Q + F, 23%Op, 10%Cc (Volcanic Breccia)	5.97; 5.98	no S_1	3.03
JB5-5F	Schist: 43%Chl, 17%Q + F, 15%Sph, 14%Op, 6%Ep, 5%Cc (Greenstone)	5.52; 5.68	$\perp S_1$	2.89
JB5-10C	Schist: 40%Chl, 22%Op, 21%Q, 13%Ep, 4%Cc (Greenstone)	6.41; 6.45	$\parallel S_1$	2.93
JB5-10C	Schist: 40%Chl, 22%Op, 21%Q, 13%Ep, 4%Cc (Greenstone)	5.99; 6.01	$\perp S_1$	2.91
JB5-10D	Phyllite: 45%Ep,34%Op, 15%Q + F, 6%Cc (Epidotised volcanic breccia)	5.65; 5.68	$\perp S_1$	3.11
JB5-10D	Phyllite: 45%Ep, 34%Op, 15%Q + F, 6%Cc (Epidotised volcanic breccia)	6.26; 6.21	$\parallel S_1$	3.07
JB5-11A	Amygdaloidal metabasalt: 42%Ep, 32%Op, 26%Q + F (Epidosite)	6.14; 6.16	no S_1	3.30
JB5-12A	Amygdaloidal metabasalt: 45%Op, 35%Q, 20%Ep (Epidosite)	6.16; 6.15	no S_1	3.15

[3] Q = quartz; F = albite; Ep = epidote; Op = opaques; Cc = carbonates; Chl = chlorite; Act = actinolite; Sph = sphene; Bio = biotite; Z = zircon.
Detailed modal analyses are in Brennan (1985, p. 13). Modal analyses were done on a 1000 point grid. Zircon was recognized in each thin section, but not in the count grid. Only one sample had more than 10% plagioclase (JB4-4D); several other samples had minor amounts. The small grain size and lack of twinning in the fine-grained samples made the distinction between quartz and feldspar difficult; for this reason, percentages of quartz and feldspar were combined (Q + F) in the sample descriptions.

[4] Compressional velocities measured under hydrostatic pressure at 400 and 600 atm, respectively. The difference is indicative of the accuracy of the velocity determinations.

[5] S_1 refers to primary foliation.
$\parallel S_1$ = velocity measured parallel to S_1.
$\perp S_1$ = velocity measured perpendicular to S_1.

SEISMICITY, SEISMIC REFLECTION, GRAVITY AND GEOLOGY OF THE CENTRAL VIRGINIA SEISMIC ZONE: PART II, GEOLOGIC FRAMEWORK

Lynn Glover, III, John K. Costain, Cahit Çoruh, Nicholas H. Evans,
G. A. Bollinger, Stewart S. Farrar, and Mathew S. Sibol

Abstract

A geologic corridor from the Blue Ridge to the eastern Piedmont near Richmond, Virginia is integrated into a preliminary tectonic model and extrapolated downward 10 to 15 km by means of seismic reflection and gravity studies. The geologic framework of current seismicity in the central Virginia seismic zone is then derived from the coincidence of subsurface fault structure with the location of well constrained hypocenters. The Blue Ridge appears to be a hinge zone that faced a rift-generated Iapetus Ocean. Eocambrian and Cambrian rift and drift-stage sediment accumulated in the North American margin. An eastern continent with an Eocambrian and Cambrian magmatic arc, collided with the North American continental margin in the Middle and Late Ordovician. The zone of suture locally still contains a subduction wedge complex of metagraywacke and ocean floor basalt. Subsequent Devono-Mississippian and Mississippian-Permian orogenesis from collisions(?) outboard of the early Paleozoic suture continued to drive thin thrust nappes onto North America. Early Mesozoic rift and pull apart(?) basins record the beginning of the Atlantic basin and during Middle Jurassic-present rift and drift, the margin of North America was covered by the Atlantic Coastal Plain sediments. Several constrained hypocenters of the central Virginia seismic zone, which occur in or adjacent to the plane of the reflection profile, show an apparent relation to structure. These records show a correspondence between seismically located faults and earthquake activity. We tentatively conclude that flat and ramp faults formed during Paleozoic nappe emplacement are currently being reactivated. The reactivation may be largely aseismic on the old thrust faults but from seismic first motion studies appears to be related to high angle transcurrent faults in the hanging wall where new rock breakage may be occurring.

Introduction

Geologic and geophysical studies by Virginia Tech along the James River traverse (Figure 1) were designed to provide a crustal section that would: 1) test current Appalachian orogenic models by comparison with a detailed transect across the Blue Ridge and Piedmont at this latitude, and 2) provide information about the geologic framework of current seismicity in the Central Virginia Seismic Zone. At the broadest level of inquiry we are attempting to clarify the accretionary history and growth of this part of the eastern North American continental margin during the last billion years. This paper is a discussion of some of our results to date and their relevance to current seismicity in central Virginia.

Two previous papers by Bollinger and Sibol (1985), and Keller and others, (1985,1985) have respectively discussed current seismicity, and gravity modelling along the James River traverse. Here we develop a tectonic model from surface geologic studies (based in part on detailed geologic mapping at 1:24,000 along the traverse) in the region and integrate that with the seismic reflection and gravity studies. The seismic reflection data are discussed by Çoruh and others in the first part of this report. Finally we relate current seismicity to the tectonic model.

Geologic Framework

A complex history of rift, drift, and collision is recorded in the development of the Blue Ridge and Piedmont rocks along the traverse. In the model presented the Taconic orogeny spans middle and Late Ordovician time. Acadian orogenesis is weakly documented here; however, it was relatively intense in the southern Appalachians (Glover and others, 1983). The Alleghanian was an intense event in the eastern Piedmont along this traverse. Brief descriptions and analyses of lithotectonic units along the traverse are given below.

Western Blue Ridge and Valley and Ridge

On the west side of the Blue Ridge, granulite facies granitoid rocks of the 1 Ga Grenville basement are overlain nonconformably by a thin unit of Eocambrian(?) Swift Run clastic rocks and nonmarine Catoctin basalt. These formations and the overlying Unicoi of Cambrian Chilhowee Group belong to the rift facies formed during the Eocambrian in eastern North America. Conformably overlying the Unicoi in the adjacent Valley and Ridge Province are shallow marine clastic and marine carbonate rocks (Gathright and others, 1977). These Cambrian and early Ordovician rocks comprise a dominantly passive margin drift sequence. Basalt erupted on the early Cambrian Unicoi shelf in southwestern Virginia (Rankin, 1975). This marks the end of the rift stage begun in latest Precambrian time. During late Early Cambrian deposition of the Rome Formation, the North American shelf was again briefly disrupted by rifting centered 50-100 km west of the Valley and Ridge in West Virginia, Maryland, and states to the southwest. This modified passive margin sequence is overlain in turn by the Middle and Late Ordovician (Taconic) orogenic clastic wedge, of which the Martinsburg Formation is the principal unit. Younger clastic wedges occurring to the west were formed during Middle Devonian through Early Mississippian orogeny ("Acadian"?) (Glover and others, 1983) and during the Middle Missippian to Permian (Alleghanian orogeny). All of these clastic wedges (interpretation slightly modified from Colton, 1970) were derived from eastern sources in the Blue Ridge and Piedmont Provinces.

Eastern Blue Ridge and western Piedmont

On the east flank of the Blue Ridge Grenville-age basement (pCg) gneisses and late Precambrian Crossnore intrusives are overlain nonconformably by east-dipping Eocambrian shallow to deep marine clastic rocks (Wehr, 1982, 1983) of the Lynchburg Group (eCly). The intrusives are bimodal, peralkaline and have been related to rifting by Rankin (1975). The Lynchburg represents an alluvial fan to deltaic system with transportation directly into a steep-sided basin from the west. The implied rapid subsidence and narrow shelf are consistent with the model of graben fill during Late Precambrian to Eocambrian rifting (Brown, 1973; Wehr, 1982).

A metamorphosed ultramafic sill within the Lynchburg Group near Schuyler, Virginia was probably emplaced during rifting. Origin as a sill is suggested by well-preserved thin beds of Lynchburg sedimentary rock that lie undeformed just below the basal contact of the ultramafic body and by the generally concordant outcrop pattern of the body. Thus, an ophiolite origin can safely be ruled out for the Schuyler.

Approximately 2 km of Catoctin metabasalt (eCc) overlie the Lynchburg Group along the traverse (Figure 1 on page 3). Directly underlying the Catoctin are lenses of conglomerate, quartzofeldspathic sandstone, and graphitic and non-graphitic mudstone with planar, massive and locally graded beds as well as lenses of mafic and rare felsic volcanic rocks (Wehr, 1983). These layers have the superficial appearance and stratigraphic position of the Swift Run Formation (Stose and Stose, 1946), which lies below the Catoctin on the northwest limb of the anticlinorium. Unlike the nonmarine Swift Run at the type locality, this southeastern sequence appears to be of marine origin (Wehr, 1983). The mixed lenses of conglomerate, sandstone, graphitic mudstone, mudstone and volcanics probably resulted from gravitational sliding, turbidites and volcanic disruption of the Lynchburg surface during initial eruption of Catoctin lavas.

The Catoctin contains interlayers of mafic pyroclastic rocks and arkosic sandstone, and coarse amygdules are common in the Charlottesville area. Pillow lava and hyaloclastite breccias occur near the top of the Catoctin along the Hardware River in Albermarle County. These features suggest that on the southeast side of the Blue Ridge the Catoctin is a shallow submarine unit, in contrast to its nonmarine nature on the northwest side (Reed, 1955).

A thin unit of quartzite occurs in depositional contact with the top of the Catoctin along the James River traverse. This forms the base of the Candler Formation of the Evington Group. Just above the base of the Candler, ferruginous basaltic breccias are interlayered with thin-bedded mudstone and graywacke (Evans, 1984). The upper part of the Candler along the JRT-1 profile is dominantly thin-bedded, metamorphosed mudstone and black claystone with minor carbonate rocks (Evans, 1984). Along strike to the southwest near Lynchburg, the Evington Group comprises the sequence, from base to top: deep water turbidites and carbonate (Candler and Archer Creek respectively), massive and planar bedded pelite (Pelier schist), shallow water quartz arenite and carbonate (Mount Athos), and Slippery Creek basalt (Brown, 1970; Bland, 1978). The Evington Group, because of its composition and stratigraphic position position above the Catoctin, has long been considered a deeper-water equivalent of the Cambrian and Ordovician(?) shallow shelf sequence of the Valley and Ridge (Brown, 1970 and references therein). Large scale crossbeds in the quartzite and locally large amygdules in the basalt also suggest a shallowing upward sequence; facies changes along strike in northern Virginia seem to support this inference. For example, the basal quartzite of the Candler on the east flank of the Blue Ridge near Charlottesville is supplanted by a much thicker sequence of quartz arenites, siltstones and mudstones in northern Virginia and Maryland. This sequence is shown as Chilhowee on the Geologic Map of Virginia (Calver, 1963).

Characterization of the Evington Group as an off-shelf passive continental margin-basin sequence is an important part of our tectonic model; however, because the stratigraphic sequence and structure of the Evington and the origin of intercalated basalt are controversial, we herein digress into a more regional discussion of these problems.

In the western Maryland Piedmont, the western Wissahickon Formation (Reinhardt, 1974; Fisher and others, 1979), which includes Marburg and Ijamsville schists and phyllites, Silver Run Limestone and Sams Creek Basalt, comprises a package of rocks along strike from, similar to, and probably correlative with the Evington Group (Bland, 1978 and references therein). Both the Sams Creek Basalt and the Slippery Creek Basalt are generally believed to be near the top of these eastern Piedmont Cambrian-Early Ordovician(?) sequences (Brown, 1970; Bland, 1978; Reinhardt, 1974). Redden (1963), however, proposed that the sequence south of Lynchburg near Altavista, Virginia was originally deposited in the reverse order of that mapped at Lynchburg by Brown. Although the problem of stratigraphic order near Altavista has yet to be resolved, the stratigraphic order with Candler at the base prevails over about 90 percent of the mapped extent. Additionally, along the Hardware River on the JRT-1 profile, the Candler seems to be in depostional contact (Evans, 1984) with the underlying Catoctin. For these reasons we tentatively accept the order of Brown (1970) which places the basalt lavas at the top of the sequences.

Espenshade (1954) noted the possibility of a correlation between manganoan carbonate in the Shady dolomite (Early Cambrian) of the Valley and Ridge and manganoan marble in the Mount Athos Formation of the Piedmont Evington Group. Barite deposits and sedimentary iron ore are associates of the manganoan strata. The Slippery Creek overlies the Mount Athos and the rift-related Rome Formation overlies the Shady. Bland (1978) gave trace element abundance data to support a rift origin for the Slippery Creek. We believe the association supports a model of Early Cambrian rifting, deposition of manganoan calcite, iron and barite from submarine hydrothermal springs and local eruption of basalt. These Early Cambrian rifting effects appear to be a separate episode later than the Late Precambrian to Earliest Cambrian rifting that generated the North American passive margin. They were superimposed on that margin as a result of rifting centered in states to the west of Virginia. Thus the Cambrian to Early Ordovician passive margin of eastern North America does not comprise a typical drift sequence but probably was modified by a rift event during the Early Cambrian.

The Cambro-Ordovician quartzite-carbonate shelf and siliciclastic slope sequences in the Blue Ridge-Frederick Valley area of Maryland prograded upward and eastward through time (Reinhardt, 1974). The Early Ordovician Grove Limestone forms the youngest unit involved in this progression. Thus the Cambro-Ordovician(?) Evington Group and western Wissahickon, like the underlying Late Precambrian or Eocambrian Lynchburg and Catoctin, give evidence of a major shelf to basin/slope transition across the axis of the Blue Ridge anticlinorium in northern Virginia and Maryland. This slope was formed during Late Precambrian rifting and dominated sedimentary facies development until the Middle Ordovician Taconic orogeny when it was destroyed by collision.

We conclude that in northern and central Virginia, the Blue Ridge appears to represent a structure of the continental margin analogous to the hinge zone (Watts, 1981) of modern continental passive margins. A hinge zone occurs at the break between shelf deposits, formed on a normal thickness of crust, and slope-basin or slope-rise deposits formed on transitional crust which has been thinned by ductile and brittle mechanisms during rifting. Thus a hinge zone is a tectonic element that displays structural, volcanic/intrusive, and sedimentological characteristics that can be recognized in ancient orogenic terranes (Glover and others, 1983, Wehr and Glover, 1985).

Hardware Metagraywacke

Along the James River traverse (Figure 1) the Evington Group is bounded on the southeast by the Buck Island fault contact with the Hardware Metagraywacke (Evans, 1984). The Mount Athos Formation and Slippery Creek greenstone of the Upper Evington Group are absent by facies change or possibly by faulting.

The westernmost unit of the Hardware metagraywacke consists of fine grained, thin beds or laminae of metagraywacke which are locally graded (Evans, 1984). An appropriate environment might be on the outer fan of a deep water turbidite sequence. Eastward these are abruptly transitional into mid fan(?), medium- to coarse-grained and thicker-bedded graywacke with local pebbly graywacke. Evans reports scattered, apparently allochthonous blocks of mafic igneous rocks in the metagraywacke. Detrital minerals in the eastern metagraywacke include quartz, plagioclase, tourmaline, epidote, magnetite, titanite, and muscovite after K-feldspar(?). Lithic fragments include dacitic tuff, gabbro, and granitoid fragments. The mineral fragments, lithic clasts and eastward coarsening all suggest a source from the east (Evans, 1984), perhaps the Shores, Chopawamsic and a Goochland-like terrane (Figure 1) which lie to the east in the James River traverse.

The Hardware graywacke sequence is similar to the Peters Creek Schist of Drake and Morgan (1981) in northern Virginia, as well as to the Wissahickon metagraywacke of the central and western Maryland Piedmont (Fisher and others, 1979). Stratigraphic and structural relations in these areas are unresolved but all authors found evidence for a source of sediment to the southeast. Evans (1984) preferred a model involving deposition of the metagraywacke directly upon the Evington Group, followed by multiple thrusting episodes at the contact. This seems to us to be consistent with the regional tectonics as presented here.

The age of the metagraywacke is younger than clasts of Chopawamsic volcanics (ca. 550 Ma Late Early Cambrian) included in it (Fisher and others, 1979) and younger than the Late Early Cambrian upper Evington Group that it was apparently deposited upon. Because the 500 Ma Occoquan Granite (Mose and Nagel, 1982) intrudes probable correlatives of the metagraywacke in northern Virginia (Drake and Morgan, 1981) the Hardware metagraywacke is probably of Middle or Late Cambrian age.

Shores Complex

The Shores Complex (Evans, 1984) was first recognized as a melange-like assemblage by Brown (1972). Major faults bound both sides of the complex (Figure 1). Evans described it as a polydeformed amalgam of epidote-chlorite gneiss, epidote-chlorite migmatitic gneiss, and hornblende-epidote-albite schist. Greenstone occurs in blocks as large as several meters in diameter in a matrix of epidote-chlorite gneiss or migmatitic gneiss. Some gneissic samples appear to have relict graywacke texture, with detrital grains of quartz and plagioclase. Migmatitic gneisses appear to have tonalitic lenses from a centimeter to several meters in thickness. Thus some of the gneissic textures imply metamorphic grades higher than the greenschist-facies mineral assemblages that now characterize the complex. Bland (1978) and Bland and Blackburn (1980) inferred an ocean floor basalt origin for Shores greenstones using trace element discrimination factors.

Taken as a whole the Shores Complex appears to represent slivers and blocks of oceanic crust scraped off and incorporated in metagraywacke during subduction (Brown, 1976, Glover et al., 1983, Evans, 1984). The melting relations described by Evans seem anomalous for the usual high P, low T conditions (Bird and others, 1975) of subduction zones; however, as Evans pointed out, melting could be reached in these compositions under conditions of P_{H_2O} approaching P_{total} at temperatures of around 630°C and pressures on the order of 6 to 8 kb (19 to 26 km).

Eastern Piedmont-Goochland and Volcanogenic Terranes

Two terranes comprise the eastern Piedmont along the James River traverse (Figure 1): a Late Precambrian(?) to Cambrian(?) volcanogenic terrane and the older (1 Ga) Goochland granulite terrane (Glover and others, 1978, 1983; Farrar and others, 1983). The Goochland terrane, which includes the State Farm Gneiss, (pCsf), Sabot Amphibolites (pCs) and Maidens Gneiss (pCm), has been seen only in fault contact with the younger terrane. Along its western margin it is thrust over the Cambrian Chopawamsic volcanic rocks of the younger terrane, and in the east, approximately 30 km southwest of Richmond, it is structurally overlain, in fault contact, by later Precambrian or Cambrian(?) Eastern slate belt volcanic rocks. South of the surface extension of the Goochland terrane into North Carolina, the volcanogenic sequence comprises the entire eastern Piedmont, although it is possible that the Goochland terrane occurs at depth. As discussed below, the Chopawamsic, Eastern slate belt, and Carolina slate belt are probably the same terrane.

Within the Goochland terrane the structurally lowest and oldest(?) unit is the State Farm Gneiss. It is a coarse- to medium-grained granodioritic to tonalitic granitoid with abundant titanite and ilmenite, indicating a high titanium content. It has a whole rock Rb-Sr isochron age of 1031 ± 94 Ma (2σ) (Glover and others, 1982). A tabular metabasalt, the Sabot Amphibolite, overlies the State Farm. It, in turn, is overlain by the Maidens Gneiss which is dominantly a biotite-quartz-plagioclase gneiss with minor intercalations of garnet-plagioclase-quartz-K-feldspar leucogneiss, pyroxene-hornblende-plagioclase amphibolite, K-feldspar + sillimanite-bearing pelitic gneiss partially to completely recrystallized to muscovite-quartz schist, and minor calc-silicate and marble layers. The Maidens is probably a feldspathic metagraywacke sequence. In addition to the above, the Montpelier metanorthosite (Clement and Bice, 1982; Bice and Clement, 1982) appears to be intrusive into the State Farm, Sabot, and Maidens at the northern end of the State Farm antiform (Farrar, 1984)(Figure 1).

Farrar (1984) showed that the Goochland terrane underwent granulite facies metamorphism at about 1 Ga and during Paleozoic events was retrograded to amphibolite- and upper greenschist-facies assemblages. This confirms the conclusion of Glover and others (1978) that the State Farm, Sabot and Maidens could all be of Grenville age.

Only the Sabot and Maidens lithologies of the Goochland terrane seem to differ from rock assemblages commonly seen in the Grenville basement of the Virginia Blue Ridge. The structurally underlying and possibly older State Farm Geniss is petrologically identical to Blue Ridge basement. Anorthosite occurs in both terranes and the age of granulite metamorphism is similar. Thus it seems probable that the Goochland and the Blue Ridge basement were once part of the same geologic terrane. Pratt and others (in press) have shown that the Goochland is a nappe of North American basement thrust over the Chopawamsic volcanics.

The younger, volcanic terrane in the eastern Piedmont is composed of 90 percent or more volcanogenic rocks. The easternmost volcanogenic rocks in Figure 1 consist of felsic metavolcanic rocks and phyllites of probable volcanic source (Bobyarchick, 1978). These rocks are similar to, along strike from, and probably correlative with the Eastern slate belt of North Carolina. Volcanics of the western part of the eastern Piedmont (Figure 1) are known as the Chopawamsic Formation and consist of predominantly rhyodacitic, dacitic and basaltic rocks that originated as lavas, and shallow intrusive and pyroclastic rocks. According to Bland (1978), the Chopawamsic rocks are calc-alkaline volcanics, a conclusion based primarily on trace element analysis. Bland (1978) concluded that eastward subduction of oceanic crust in a collapsing back-arc basin generated the calc-alkaline volcanics of the Chopawamsic Formation. Although we do not agree with a back arc interpretation of the Evington basin, melange (Shores Complex) found in association with ocean floor basalt seems consistent with the eastward polarity of subduction. This is also consistent with the present east-dipping structure of the Shores.

The age of the Chopawamsic volcanics is probably about 550 Ma or Early Cambrian (Pavlides, 1981, and references therein). Thus the Chopawamsic appears to correlate with the post Virgilina-deformation sequence (Glover, 1974; Briggs and others, 1978; Harris and Glover,1983) of the Carolina slate belt in the Albemarle area of central North Carolina. The Chopawamsic volcanics are laterally continuous with Charlotte belt volcanics (Higgins and others, 1971, Glover, 1974, and Conley, 1978). Thus, the Chopawamsic and Carolina slate belt (including the Charlotte belt) are probably all part of the same geologic terrane.

Synthesis of Late Precambrian to Late Ordovician Tectonic Events

The following points are offered to justify our tectonic model. Previous discussion shows that Crossnore intrusives in the Blue Ridge, and overlying strata indicate that rifting of the North American continent began about 700 Ma (Crossnore) and continued through the Eocambrian to about 570 Ma. Rifting produced the Evington basin along the North American continental margin. Rift facies are overlain by Cambrian to Early Ordovician drift facies which change across the hinge zone from a shelf sequence of shallow water quartz arenite and carbonate on the west to deeper water siliclastic rocks and minor carbonates in the Evington on the east. By Early Cambrian time (ca. 540 Ma) the basin may have prograded nearly to sea level in central Virginia if the correlation of Shady Dolomite (Valley and Ridge) and Mount Athos (Piedmont) is correct.

Several facies of the Early Cambrian sequences indicate that the passive margin and Evington basin were undergoing rifting and basaltic volcanism during this time. Ammerman and Keller (1979) have discussed the Early Cambrian development of the Rome nonmarine and shallow marine siliclastics and dolomite into the Valley and Ridge from Alabama to Pennsylvania. This event is commonly thought to be largely centered in states 50 to 100 km west of the Valley and Ridge; however, proprietary data from industry indicates that rifting also affected the Rome in the Valley and Ridge of Alabama. Somewhat earlier, basalt had erupted on the Chilhowee shelf in southwestern Virginia and probably signaled end of Late Precambrian - Early Cambrian rifting that generated Iapetus. As discussed previously, eruption of the Slippery Creek and Sams Creek basalts, both probably of Early Cambrian age, in the Evington basin are additional evidence of rifting along the eastern margin of North America at this time. How the Rome rifting comports with the model of a passive continental margin just prior to and perhaps during continental collision is an interesting problem for the future. At the moment the evidence is substantial that it was a tectonic event that operated to modify the Cambro-Ordovician drift sequence along the edge of the North American continental margin.

To return to Middle and Late(?) Cambrian events, while the carbonate bank (Elbrook, Knox) was accumulating west of the hinge zone on the platform, turbidites of the eastern Wissahickon were succeeded by diamictite (Sykesville) and metagraywacke. In Maryland most authors have concluded that the diamictite grades westward into metagraywacke (Fisher and others, 1979); however, in northern Virginia Drake and Morgan (1981) found that the metagraywacke and Piney Branch ophiolite were transported westward as hard, metamorphic slices over younger, unconsolidated Sykesville diamictite during their emplacement. Subsequently the tuffaceous Popes Head metasiltstone and phyllite (forearc basin?) were deposited unconformably upon the melange, and at about 500 Ma (Mose and Nagel, 1982) the Occoquan Granite intruded the whole sequence. Thus we are left with a fundamental problem that can only be solved by future field work; the coeval soft sediment deposition of the diamictite-graywacke suite in Maryland appears to be incompatible with the hard rock state of transport claimed for the metagraywacke and Piney Branch in northern Virginia. Additionally, emplacement of the Occoquan Granite (500 Ma), Ellisville granitoid (440 Ma; Pavlides and others,1982), Melrose Granite (ca. 470; Gates, 1981) and Leatherwood Granite (460; Odom and Russell, 1975) into the melange-metagraywacke sequences along strike in northern, central and southern Virginia collectively seem to show that most of these metamorphosed and unmetamorphosed strata were tectonically stacked before Early Ordovician time.

Evans (1984) and Fisher and others (1979) showed that the composition of the metagraywacke, Shores Complex and Sykesville diamictite required sources that included the eastern continent (Chopawamsic, Shores), and probably rift and drift facies of the North American continental margin apron. Except for the Late Cambrian Popes Head Formation (Drake and Morgan, 1981) which unconformably overlies thrust stacks of metagraywacke and melange, primary pyroclastic debris appears to be absent in the metagraywacke-melange-diamictite association. Because of the buoyancy of the subducting North American continent the dip of the subduction zone may have decreased after initial collision. Under similar conditions in the Himalayan orogene volcanism diminished in volume and moved toward the interior of the overriding continent. A similar response in the Appalachian orogene might explain the apparent Late Cambrian diminution of volcanism in the Chopawamsic terrane; however, the Popes Head volcanics and bentonites from the Middle Ordovician of the Valley and Ridge attest to sporadic and perhaps distant volcanism continuing from the eastern continent.

The Cambrian age of teh Shores and its melange of ocean floor basalt suggest accumulation or an accretionary wedge on the ocean floor prior to collision with the North American continent during the Middle and Late Ordovician Taconic orogeny.

Post-Taconic Paleozoic Events

The Taconic collision was accompanied by igneous intrusion, metamorphism, and deformation with uplift and formation of an erosional unconformity (Brown, 1969) over the region of the suture zone. The Late Ordovician to Early Silurian(?) Arvonia Formation was deposited above the

Taconic unconformity as subsidence produced a basin in the Piedmont. Younger Paleozoic stratigraphic units appear to be absent in the Piedmont.

Structural, metamorphic and igneous events that have been isotopically dated provide the rest of the Paleozoic history for the region of Figure 1. Synthesis of the ages of regional metamorphism and ductile deformation (Glover and others, 1983) indicates that both the Piedmont and Blue Ridge in the region of Figure 1 were metamorphosed and ductilely deformed during the Taconic. Acadian ductile deformation recrystallized rocks along the Taconic suture. At about 330 Ma the Petersburg and associated granites were emplaced in the eastern Piedmont (Wright and others, 1975); this may have been at about the beginning of the Alleghanian orogeny. Alleghanian metamorphism and ductile deformation produced amphibolite-facies assemblages in the eastern Piedmont with resultant retrogression of the granulite assemblages of the Goochland terrane. Alleghanian orogeny produced crustal shortening in the eastern Piedmont by reactivation of older faults but also by ductile shortening in the eastern Piedmont. The western Piedmont experienced reactivation of old fault zones as well as translation westward as an essentially rigid block during this time. As mentioned above, each of these orogenies produced clastic facies in the Valley and Ridge.

Triassic-Jurassic events

The Permian to Triassic transition from compressional to extensional tectonics has been discussed for the region of Figure 1 (Bobyarchick and Glover, 1979). During the Triassic-Jurassic, Paleozoic mylonitic thrust faults were reactivated in an extensional regime (Glover and others, 1980). Because thrust faults generally flatten with depth, the form of early Mesozoic reactivation faults bounding the initial Triassic basins can be predicted to be listric-normal. Vibroseis data discussed below tend to confirm this conclusion. Manspeizer (1981) summarized evidence north of Virginia for a southeast trending left lateral strike slip framework for the origin of these basins; in the south, subhorizontal slickensides occur in Late Triassic sediments along the western border fault of the Richmond basing in Virginia (Glover, field notes). It will be important in the future to determine the role of strike slip movement in the formation of these southern basins. Triassic border faults in Virginia follow old Paleozoic mylonite zones parallel to Appalachian structural trends (Glover and others, 1980) and these southern Appalachian Mesozoic basins do not contain Triassic basalt or diabase in the exposed Piedmont. Diabase and basalt appear during the Latest Triassic-Early Jurassic as volcanics in the basins north of Culpeper, Virginia and as dikes cutting all central and southern Appalachian basins at high angles (Manspeizer, 1981; Ragland and others, 1983). Therefore the timing of initial rifts, their control by older faults and lack of initial igneous activity is consistent with, but does not prove, simple extensional deformation in the upper crust without direct structural connection to the mantle source of basalt. As May (1971), Manspeizer (1981) and Ragland and others (1983) have summarized, diabase dikes in the Carolinas and Virginia appear during the Early Jurassic and are concentrated in the Piedmont (especially over the Piedmont gravity high) and Atlantic Coastal Plain basement (i.e., generally over thinner crust). The angular relation of these dikes to the basins can be generally explain in the southeastern U.S. by a left lateral shear model, but on a hemispheric pre-drift basis (May, 1971) may require mantle movements producing a radial dike pattern centering on the Blake Plateau, Bahama Platform and western Senegal Basin. The timing of deformation that tapped mantle basalt sources suggests to us that transcurrent movement postdates simple extensional opening of the Triassic basins. Alternatively both Triassic and Jurassic tectonics may be related to dextral (Triassic) and sinistral (Jurassic) transform shear that preceded opening of the Atlantic (Swanson, 1982).

The relative importance and timing of transcurrent versus dip slip faulting in the early Mesozoic is an important problem in understanding the tectonic history of the eastern U.S. Offshore data sheds some light on the subject. Grow (1981), summarizing unpublished data from D. Hutchinson, places the early Mesozoic to Recent hinge zone along the west flank of the Carolina and Baltimore Canyon troughs. This position corresponds to the New Jersey Coast in the north and to the offshore Brunswick anomaly in the south. Sediments in the troughs extend onto oceanic crust, which is believed to be as old as Early Jurassic. Thinned intermediate continental crust is believed to underlie all of the Carolina Trough and the western part of the Baltimore Canyon Trough. From the existing data base we believe that in both regions simple orthogonal or simple oblique (northern part of Baltimore Canyon Trough) separation formed oceanic crust. No evidence of northerly directed transform motion is reported in these Early Jurassic and younger records in contrast to the evidence on land summarized by Manspeizer (1981) and Swanson (1982). Thus, it seems possible that the transcurrent movement described by Manspeizer and Swanson is confined to the continent and is older than the opening of the Atlantic. The dikes and rifting along the Piedmont gravity high may represent a short lived and aborted rift that experienced some Latest Triassic and Early Jurassic lateral translation between Africa and North America as discussed by Swanson (1982).

Cretaceous to Recent Tectonics

Following the Triassic to Early Jurassic rift and wrench stage of development of the Atlantic margin, a drift stage of sedimentation began. Jurassic and younger sedimentary units progressively onlap the continental margin; the western edge of this seaward-thickening wedge appears in Figure 1 as the Atlantic Coastal Plain. Cooling and sedimentary loading of the lithospheric plate as it receded from the midocean spreading center explain much of the subsidence required to allow deposition of a dominantly shallow water sequence of drift stage sediments (Watts, 1981). The progressive onlap is apparently caused by seaward loading of a lithospheric plate that flexes farther toward the continental interior as its flexural strength increases by cooling and thickening through time.

An additional tectonic event affecting the drift sequence is high angle reverse faulting of small magnitude (Mixon and Newell, 1977). These faults appear to be compressional Paleozoic faults and mylonites (Bobyarchick and Glover, 1979; Glover and others, 1980) that were reactivated as listric normal and wrench faults in the early Mesozoic. They were again reactivated during the drift stage of passive margin development, but this time as high angle reverse faults cutting Cretaceous to Miocene and perhaps more recent strata (Mixon and Newell, 1977). Some of the faults described by Mixon and Newell strike into or are coextensive with the fault along the western edge of the Goochland terrane and the Hylas zone of Figure 1.

Interpretation of Vibroseis Profiles JRT-1 and JRT-2

In this section, Virginia Tech Vibroseis profiles across the Piedmont and Blue Ridge are presented and integrated with the preceding data and interpretation. JRT-1 begins on the Blue Ridge (Figure 1) and continues to the eastern Piedmont near Sabot, Virginia. The westernmost segment of JRT-2 is also discussed below. For estimates of depth a velocity of 6 km/sec can be assumed over most of the profile. Small data gaps exist in the eastern and western parts of the profile. Geologic units plotted along the top of the profile can be compared to the geologic map in

Figure 1. Faults and geologic contacts are extrapolated into the subsurface where seismic reflections provide control. At large distances from surface control, seismic signatures and inferences from the regional data and tectonic model presented above have been used in interpretation.

Blue Ridge

Grenville basement underlies the western end of profile JRT-1 (Figure 1 and Figure 14) and, except for an amphibolite intrusion just below the Lynchburg, and reflections interpreted as faults bounding the Mechums River Formation (pCmr, Figure 1), the Grenville is acoustically transparent. Below about 1.1 sec. there are a number of east-dipping reflectors; between 2.2 and 3 sec. there are sub-horizontal reflectors. The rocks between 1.1 and 3.2 sec. are interpreted to be Cambrian-Ordovician clastic and carbonate rocks of the Valley and Ridge sequence. There are two reasons for this interpretation: 1) below the outcrop of Catoctin and Lynchburg at about 1.1-1/2 sec. there is a strong reflector that truncates the Lynchburg. The magnitude of this reflection indicates good acoustic contrast as expected between crystalline and sedimentary rocks. The intersection of this reflector with that from the base of the Catoctin strongly indicates that the subhorizontal reflection is from a fault contact. To the west at about 1.1 sec. below sta. 280 a similar subhorizontal reflector suggest continuation of this fault. 2) Evidence that the Blue Ridge is a thin sheet of crystalline rock thrust over the unmetamorphosed Valley and Ridge sequence is available to the south along strike near Roanoke. There, several erosional embayments into the Blue Ridge reveal Rome Formation and other unmetamorphosed Valley Ridge strata below a thin, subhorizontal thrust plate of crystalline rocks on the west flank of the Blue Ridge (Calver, 1963). Therefore the model preferred here is that the Blue Ridge crystalline plate is only about 1 sec. (3 km) thick and that it overlies Cambrian to possibly Early Ordovician carbonate and clastic rocks of the shelf sequence. A section of Shady dolomite and Chilhowee quartz arenite is inferred (Figure 14) below the deepest reflector at 3 sec. (ca. 9 km) because of the likelihood that the reflections at 3 sec. originate in stratigraphically higher Rome Formation.

Discordant reflections just below the outcrop of the Lynchburg are correlated with mafic intrusives in the section. These are shown on the geologic map of Albermarle County (Nelson, 1962) and the reflection just west of the Lynchburg corresponds to a mapped mafic intrusive in the Grenville basement. In the interpretation shown in Figure 14, the Lynchburg wedges out by thrust truncation about 30 km east of its outcrop. As discussed below, Lynchburg may also be present in a wedge above basement under the eastern end of segment NSF-3 (Figure 14). Çoruh and others (previous section) characterized the seismic signature of the Catoctin and presented geophysical evidence for faulting associated with it. Segment NSF-2 of profile JRT-1 shows that the resolving power of 24-fold data versus the 12-fold segments that flank it (Çoruh and others, previous section). Numerous small ramp and fold (snake head) structures are particularly well recorded within and below the Catoctin (Figure 14). As Çoruh and others (previous section) have shown, a subhorizontal fault just below the Catoctin at about 1.8 sec. under the west edge of the Scottsville basin gives evidence of possible backsliding and truncation of the top of the Cambro-Ordovician section. The pattern of ramp faults within the Catoctin in NSF-2 is consistent with compressional thrusting directed westward. The subhorizontal thrusts may have developed in response to bending of the more rigid Catoctin lower units as they were flexed in a concave upward direction. Eastward extrapolation of the subsurface Catoctin is based on the oversimplified assumption that it retains approximately the same thickness and that it may be continued down dip concordantly with the reflectors as shown.

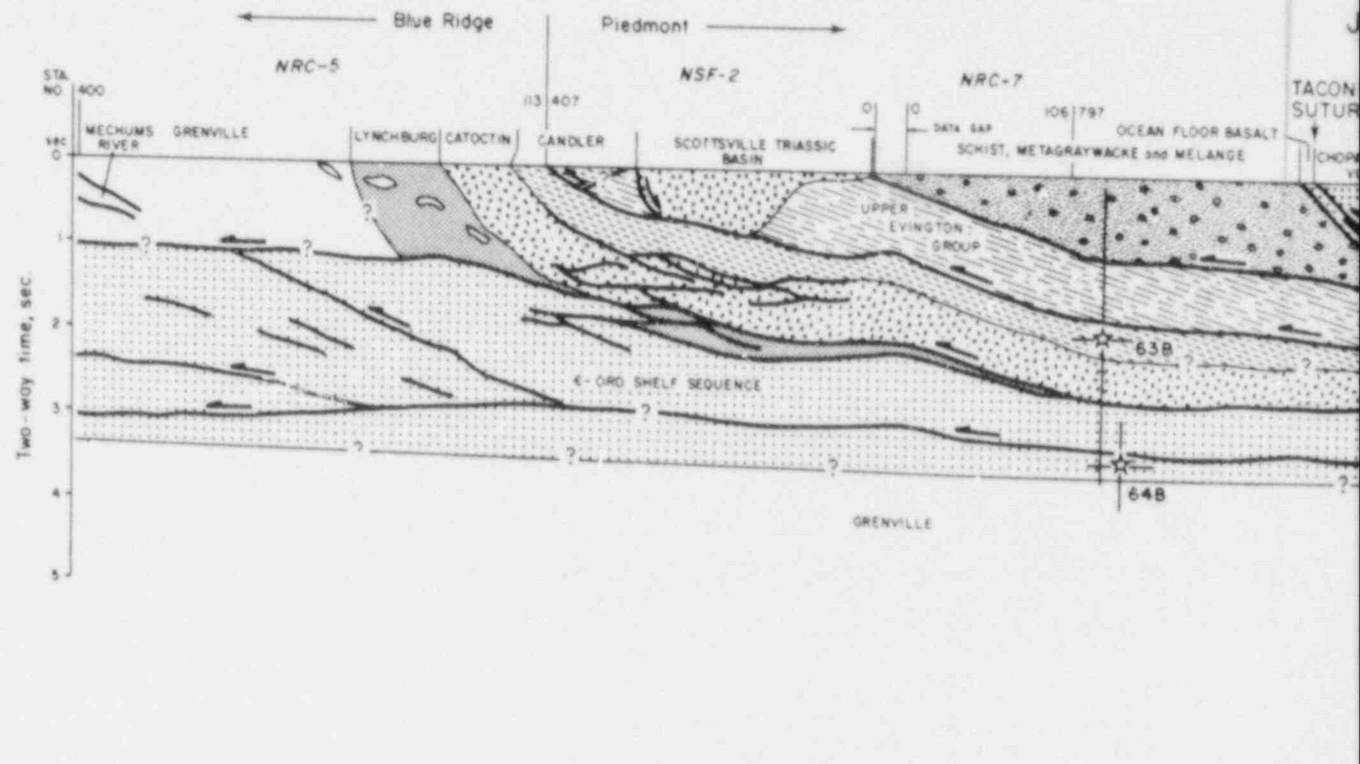
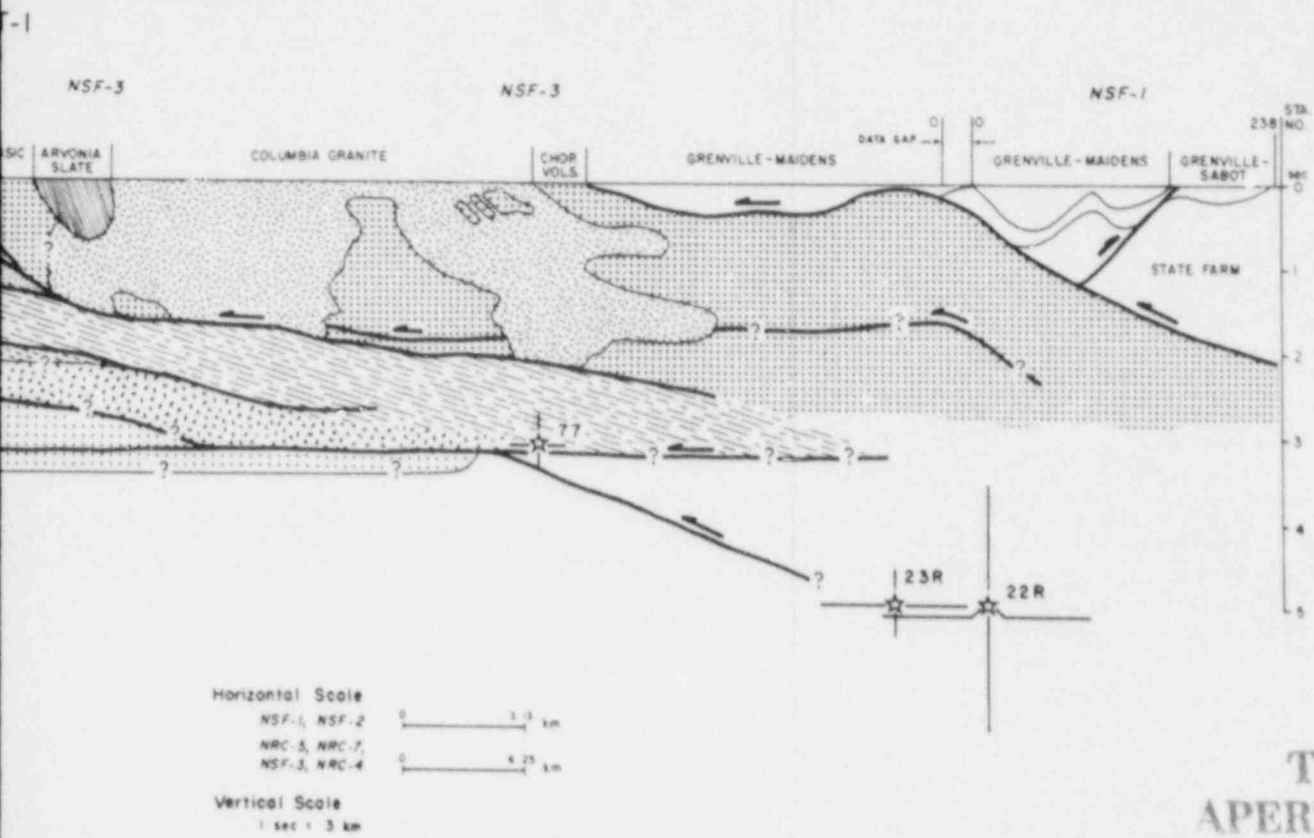


Figure 14. Geological interpretation of JRT-1.



8805060296-02

Piedmont

Above the Catoctin, the Candler Formation of the Evington Group is characterized in segments NSF-2 by a localized distribution of reflectors. Those near the west end at the outcrop of the group are probably from volcanic breccias. Discordance of some of these Evington reflectors at the west end of segment NSF-2 suggests thrusting along the Evington-Catoctin contact, but there is no evidence in the field or in the seismic record that this involves major translation.

The fault at the top of the Candler may have major displacement. Mafic lavas crop out along the southwestern edge of the Scottsville basin and, in NSF-2, appear to dip northwestward to where they are truncated by a fault. Southwest of the Scottsville basin the upper Evington group includes the Arch Marble, Pelier Schist, Mount Athos Quartzite, and Slippery Creek Greenstone. This layered sequence contributes to the abundance of reflectors seen in the upper Evington group above the Candler. The reflections from below the southeastern margin of the Scottsville basin form an anticlinal-synclinal flexure. Palinspastic closure of the Scottsville basin by moving the upper Evington group of the southeastern wall updip would result in restoration of a broken anticline. Çoruh and others (previous section) discuss reflections and seismic response of the Scottsville basin.

The geometry of the Scottsville triassic basin suggests that early Mesozoic extension produced backsliding on the low angle Paleozoic thrust fault that bounds the bottom of the basin. Perhaps because of the low overburden pressure at the west edge of the basins, a normal fault developed there and broke through the crest of a preexisting antiformal flexure rather than following the basal thrust westward to the surface. Erosion modified the resulting fault scarps as the basin filled. Surface outcrops do not show Paleozoic mylonite along the border of the Scottsville basin as is common in early Mesozoic basins elsewhere in Virginia which are more obvious cases of reactivation of Paleozoic faults. Nevertheless, the influence of older fabric on the localization of the Scottsville basin is well demonstrated by the seismic reflection data over this area.

The upper Evington group, which flanks the Scottsville basin, has been projected downdip on the basis of conformable reflectors judged to be interfaces between basalt and sedimentary rock. A thrust fault shown at the top of the sequence emerges at the north end of the Scottsville basin where it separates metagraywacke on the east from metamudstone of the Evington Group on the west. Apparently the anticlinal fold in the basalt/quartzite/marble/schist sequence plunges northward and these rocks do not crop out north of the southern end of the basin (Figure 1).

Subduction Complex

Rocks of the subduction complex of schist/metagraywacke/melange unit, which comprises the Evington Group(?) of Smith and others (1964), are acoustically transparent or dip too steeply to provide reflections in the profile. Mapping by Smith and others shows tight folding, overturning of beds and moderate to steep dip. The surface structure is therefore sharply discordant with the gently dipping and subplanar reflections in the underlying upper Evington Group. This discordance provides additional justification for interpreting a low angle thrust fault between them. The wedge shape of the subduction complex in cross section is constrained by projection of the gentle dip of the underlying upper Evington Group and the steeper dips of the Shores Complex and Chopawamsic volcanic rocks as measured at the surface. Obviously the constraints are not strong and more of the subduction wedge could be present in the subsurface to the east.

Ocean floor basalts (Bland, 1978) are present at the eastern edge of the subduction Shores complex (Figure 1), in contact with calc-alkaline volcanic arc rocks of the Chopawamsic Formation (Pavlides, 1981). Both groups of volcanic rocks show intense faulting. South of the James River, Brown (1969) mapped basalt and ultramafic rock in the eastern part of the Shores. The hornblende diorite and associated hornblende metadiorites (Figure 1) are also thought to be allochthonous mafic units (Brown, 1973).

Eastern Piedmont Volcanogenic Terrane

Çoruh and others (previous section) discuss the seismic signature of the Chopawamsic volcanics and specific structural features found in the seismic section in this area. The quality of the reflections in the 24-fold eastern part of segment NSF-3 is particularly good and these Chopawamsic rocks can be discriminated to some extent by the seismic signature from Catoctin reflections previously discussed.

The eastern half of JRT-1 (Figure 1) shows an allochthonous terrane of Cambrian Chopawamsic volcanic rocks, intruded by the Ordovician Columbia Granite. The Arvonia Slate is shown above a synclinally folded unconformity developed on the older Chopawamsic and Columbia. The base of the Arvonia is suggested by a single concave upward reflector. This is the Taconic unconformity of the Virginia Piedmont which resulted from the uplift and erosion that accompanied the Taconic deformation.

The Columbia Granite is acoustically transparent; its boundaries are drawn where reflections originating at Chopawamsic mafic-felsic interfaces start. The lower boundary of the Columbia is placed at the top of the gently dipping reflections from either the Chopawamsic or upper Evington Group. The abundance of subhorizontal reflections (Çoruh and others, previous section) below the Columbia suggests that the granite did not intrude through these layered rocks, and thus that the Chopawamsic and Columbia are both allochthonous with respect to the underlying rocks. The thrust boundary is projected in JRT-1 to the western edge of the outcropping Chopawamsic and is interpreted as the eastern edge of the Taconic suture.

East-dipping reflectors indicate a large mass of Chopawamsic volcanic rocks in the subsurface in the central part of the Columbia Granite. Discordance between the dip of the higher and lower reflectors in this enclave suggest the possibility of a preintrusive fault. Other east-dipping reflectors near the surface at the eastern margin of the Columbia are interpreted as thin, slab-shaped enclaves of mafic Chopawamsic. Similar slab-shaped enclaves occur at the surface along strike and just south of those shown in JRT-1 (Brown, 1969).

Gently dipping to subhorizontal reflectors display the subsurface continuation of the easternmost outcrops of Chopawamsic volcanic rocks. The abundance of reflectors probably records mafic and felsic interlayering in the Chopawamsic. A number of small thrust faults could be interpreted from the data, but only one is shown in the profile. At the eastern end of the profile, Chopawamsic reflectors and the boundary of the overlying Goochland terrane dip more steeply to the east.

The Goochland terrane, comprising State Farm granitoid gneiss, Sabot amphibolite and Maidens gneiss, structurally overlies the Chopawamsic at the eastern end of profile JRT-1 (Figure 1). A major contribution of Pratt and others (in press) was the interpretation of the Goochland Terrane as a nappe of North American basement that overrode the Chopawamsic metavolcanic rocks.

Along the western (segment NSF-3) outcrops of the Maidens, moderate eastward dips are seen at the surface. At the crest of the antiform in the underlying Chopawamsic, near the data gap, an antiformal structure has also been mapped at the surface. This structure, the relatively gentle regional dips in the Maidens, and the underlying gently dipping reflections in the Chopawamsic suggest a low angle thrust contact between the two. The Sabot Amphibolite produces reflectons, but the Maidens and State Farm are generally acoustically transparent or have dips that are too steep to yield reflections. An exception are some reflections in the State Farm that dip westward, concordant with the antiformal structure of this truncated dome. A possible small backthrust occurs near the surface outcrops of the Sabot. This can only be seen in the reflection data, but explains a poorly exposed but obviously thickened outcrop of Sabot shown on Figure 1.

Segment NRC-4 is the westernmost part of JRT-2 (Figure 1). Subduction wedge units (Shores Complex) crop out at the surface but identification of the deeper reflections is uncertain because of the short length the segment precludes tracing the reflectors to the surface. By comparison with the adjacent segment of JRT-1, however, it seems probable that the reflectors between 1 and 2.2 sec. are upper Evington Group. Those between 2.2 and above 3 sec. are Catoctin. Although NRC-4 is only 12-fold data, folding and ramp faulting is clearly shown below station 91 at about 2.4 sec. Çoruh and others (previous section) show evidence for other faults in this segment. The position of the sole fault is not well constrained but is essentially consistent with its position in JRT-1.

USGS I-64 Vibroseis profile

An 8 sec reflection profile (Harris and others, 1982) along highway I-64 passes subparallel to JRT-1 at distances of 5 to 30 km to the north (Figure 1). The published profile was unmigrated and difficult to interpret in its original state. The interpretation of Harris and others shows many similarities with that of our profile, but there are also important differences. Correlation of the 1 Ga Sabot amphibolite with the ca. 550 Ma Chopawamsic volcanic rocks is an error that significantly differs from the geologic interpretation shown by us in JRT-1. This profile was reprocessed to extend its depth to include Moho depths at Virginia Tech (Pratt and others, 1987) and is the basis for our conclusion that the Grenville Goochland terrane is an upthrust plate of North American basement.

Structural Synthesis of Profile JRT-1 and JRT-2

As in previous seismic reflection studies of the central and southern Appalachian (Clark and others, 1978; Cook and others, 1979; Harris and others, 1981 and Harris and others, 1982) the JRT-1 and JRT-2 profiles demonstrate thin-skinned tectonics involving the Valley and Ridge, Blue Ridge and Piedmont sequences. A sole fault, probably stratigraphically controlled within the Rome Formation, dips gently eastward to a point in the eastern Piedmont were it appears to turn more sharply downward, cutting into the Greenville Basement. This steeper, eastwardly dipping zone is analogous to a similar feature in the COCORP traverse beneath the Elberton Granite in Georgia (Cook and others, 1979). Iverson and Smithson (1982) have interpreted the steeply dipping zone on the COCORP traverse as a root zone and in a general way we concur with this conclusion.

We suggest that in the JRT-1 profile, the point on the sole fault where the dip increases sharply is not far west of the pre-collisional position of the western part of the Blue Ridge hinge zone. That is, the western subsurface edge of Greenville Basement in the Blue Ridge can be restored palinspastically to a point not far east of the JRT-1 section line and juest east of Richmond based on our interpretation of the U.S.G.S. I-64 profile. In this restoration the Blue Ridge hinge zone marks the edge of Cambro-Ordovician shallow shelf sedimentation and a contour along the ancient continental margin, west of which the crust was of normal thickness. East of this contour the continental crust was attenuated during Late Precambrian rifting. This attenuated crust was the locus of deposition of Late Precambrian-Early Ordovician dominantly deep water clastic facies. With additional seismic profiling toward the southeast it may be possible to determine the extent of the remaining Late Precambrian continental edge beneath the Atlantic Coastal Plain. Palinspastic reconstruction of the limited profile in JRT-1 places the pre-Taconic continental edge well east of Richmond, Virginia.

It is noteworthy that the M-discontinuity (James and others, 1968; Pratt and others, 1987) rises from nearly 40 km below the western end of traverse JRT-1 to about 34 km at the eastern end. Yet identification of the pre-Taconic hinge zone near the eastern end of the profile and the demonstrated shallow shelf sedimentation west of the hinge seem to suggest a nearly constant depth to M over the profile prior to the Taconic. Therefore the present shape of the M-discontinuity may be a relatively young feature. We suggest that the M-discontinuity attained its present shape during the early Mesozoic rifting of the North America continent. The early Mesozoic dikes occur within the attenuated part of the present North America margin; they, along with the Triassic-Jurassic basins, are possible manifestations of the crustal attenuation.

Above the sole fault under the Blue Ridge, as much as 6 km of unmetamorphosed(?) Cambrian to Early Ordovician(?) shelf sequences are stacked in a duplex structure. The Blue Ridge itself is a relatively thin (3 km) sheet of metamorphic rock. Above the Blue Ridge sequence to the southeast occur North American shelf and slope sequence rocks that have been overthrust by a subduction wedge and the volcanogenic and microcontinent that collided with North America during the Ordovician.

Subsequently the Acadian (Middle Devonian through Early Mississippian in this region) orogeny deformed these rocks (Glover and others, 1983) and the Alleghanian Orogeny produced an amphibolite-grade ductile deformed metamorphic belt in the eastern part of the region (Glover and others, 1983). Each of these deformational events involved many kilometers of westward transport and probably reactivated older thrust faults.

At first glance, a wide discrepancy seems to exist between the structures seen in the field that result from multiple, intense deformations and the extensive, gently dipping strata and thrust faults revealed by the seismic work. An analysis of the relation of cleavage and folding to the development of crustal structure as revealed in profile JRT-1 is in progress. Mesoscopic isoclinal and tight folds are known from surface studies (Bobyarchick and Glover, 1979), but such structures do not have many counterparts on the scale of JRT-1. Regional, blanket-shaped units became highly contorted and multiply deformed on a small scale, yet regionally have acted as competent units that deform by thrust faulting in a manner similar to the unmetamorphosed Valley and Ridge sequence. This is a major new conclusion only revealed by the recent application of seismic reflection techniques to crustal structure in the crystalline Appalachians.

A few clues to the rheology during thrusting are available. Glover and others (1983) noted that the western and central Georgia Piedmont, which was at high temperature during the Alleghanian, contain little-deformed granites that were emplaced in the eastern Piedmont. At this time the central and western Piedmont was passing through a declining temperature gradient from 500°-300°C, yet it behaved as an essentially rigid body, being translated westward and deforming the Valley and Ridge sequence by buckle-folding and thrusting. Although the central and western Georgia Piedmont was quite hot, it had declined from its thermal peak and could no longer easily recrystallize. The rigidity of this terrane may have resulted from the effects of uplift, erosion and cooling that attended initiation of thrusting. The effect would be to reduce water pressure and temperature in the uplifting terrane and thus inhibit the ability to recrystallize in adjusting to the new conditions. Because ductile behavior requires recrystallization, a hot terrane just receding from the threshold of prograde metamorphism might become relatively brittle.

Mylonite zones in the metamorphic Appalachians are commonly best developed just after the thermal peak of prograde metamorphism (Bobyarchick and Glover, 1979; Sinha and Glover, 1978). Although they display ductile behavior and ease of recrystallization within the zone, they can be viewed on a regional scale as semibrittle, isolated planar deformation zones (usually thrust faults) within a terrane that had just ceased ductile development of foliation. The ductile behavior (recrystallization) within the zone probably results from the energy added to the system by localized compression and shearing. In summary: once crustal thickening and thrusting are initiated, the upper levels can behave as rather rigid plates even though they remain relatively hot. This may, in part, explain the regionally planar structural features revealed by the vibroseis studies.

Localization of seismicity through structural heredity

In a companion paper Bollinger and Sibol (1985) have analyzed the characteristics of the Central Virginia seismic zone. Characteristics pertinent to this discussion include: 1) a horizontally and vertically diffuse pattern that implies recurrent movement on faults scattered over a wide area. 2)In the central and western Piedmont most of the activity is concentrated in the upper 11 km of the crust; in the eastern Piedmont activity occurs to depths of about 15 km. 3) The epicentral pattern defines an arcuate, almost circular area bound by a convex curve on the southeast and by a nearly straight NE trending line on the northwest.

Seven hypocenters that have vertical errors of less than $\pm$ 5 km and lie within or near reflectors on profiles JRT-1 and JRT-2. Horizontal errors are averages of the two ellipse axes given by Bollinger and Sibol. Hypocenters 23R and 22R lie within 7 km of JRT-1; the others lie on or within about 1 km of JRT-1 or JRT-2 (Bollinger and Sibol 1985, Fig. 9). Hypocenters that lie off the plane of the profiles were projected into the planes of the profile along structural axes determined from surface geology, optimizing the chance that they fall on the profile in the same fault structure in which they originated. Accuracy of the hypocentral location is given by Bollinger and Sibol. These are among the most accurately determined locations within the Central Virginia seismic zone and are representative of the general increase in maximum depth from about 10 km on the west to about 15 km at the east end of the Central Virginia seismic zone. Because the zone is still active, much greater accuracy of hypocentral location will be attained in the future by relocating the recording stations directly over the currently active structures. Thus future work has the possibility of greatly improving the correlation between structure and seismicity, by changing the geometry of the recording network and by improving structural resolution with higher fold seismic reflection data over an already identified active structure. Focal mechanism solutions could then provide an additional confirmation of the probably faults involved.

Hypocenter 63B in profile JRT-1 has a vertical error of $\pm$ 5 km and falls between two faults whose existence seem well founded but whose actual positions are poorly constrained (see previous discussion). Hypocenter 64B, however, has a vertical error of only $\pm$ 2 km and is closely associated with the sole fault of the allochthon. The western segment (NRC-4) of profile JRT-2 contains two hypocenters, both of which lie in the plane of the profile. The vertical error of 64C is $\pm$ 4 km and that on 64A is $\pm$ 2 km (Bollinger and Sibol, Table 3, 1985). Although the errors are relatively large, the hypocentral locations (statistically the most probable locations) fall directly on two well defined faults in the vibroseis profile. If 64B (on JRT-1) is plotted on NRC-4 profile JRT-2, it also falls on the sole fault at 3.3 sec. Therefore, both the sole and ramp faults in JRT-1 and JRT-2 appear to be currently active, and the activity appears to be confined to the allochthonous plate.

To the east, Hypocenter 77 has its epicentral location in the plane of segment NSF-3. This hypocenter has a remarkably small vertical error of 1 km and a correspondingly small horizontal error. It plots in JRT-1 along the continuation of the sole fault just east of the point of inflection where the sole fault dips more steeply east. This tends to confirm the estimated depth of the sole fault at this location as well as the previous conclusion that the sole is currently active. Near the west end of the profile, hypocenters 23R and 22R have been projected northward about seven km along the surface structural axis into the plane of JRT-1. They lie at a depth of abut 15 km and along the dip of the sole fault shown in the profile. Although the dip of the sole fault is not strongly supported by the seismic reflection data, its location and dip seem supported by the gravity modelling of Keller and others (in review). Therefore, the increase in depth of earthquake activity in the eastern Piedmont seems consistent with the structural evidence for eastward rooting of the allochthonous plate.

From the above we tentatively conclude that much of the current seismicity in the Central Virginia seismic zone is localized as a result of movement along old Paleozoic sole and ramp faults within the allochthonous Appalachian plate. Seismicity may result from new rock breakage in the hanging wall and aseismic slip on old thrust faults. This model is suggested by the high angle nodal planes in first motion studies by Bollinger's group (Munsey, 1984). Such a model is being tested in our current program.

The region was doubtless seismically active during the early Mesozoic when extension and transcurrent faulting reactivated some of the old thrust faults and developed the early Mesozoic basins. During the Tertiary these faults as well as others that may have been annealed since the Paleozoic were again reactivated in a compressional regime when high angle reverse faulting occurred. The diffuse nature of the Central Virginia seismic zone, and the apparent association with ancient thrust and ramp faults, many of which seem unrelated to early Mesozoic basins, suggest that any fault properly oriented with respect to the current stress field is a candidate for localization of earthquake activity.

The above conclusion leads to a more important question about the nature of seismicity in the eastern U.S. If the structure shown in profile JRT-1 is similar to that elsewhere in the central and southern Appalachian Piedmont, as it is known to be, why is the current activity in the Piedmont (see Bollinger, 1973) concentrated in the Central Virginia and South Carolina-Georgia seismic zones?

A number of recent papers, including Barosh (1981), Engelder (1982), Hamilton (1981), McKeown (1978), Seeber and Armbruster (1981), and Sykes (1978), summarized hypotheses to account for eastern seismicity. A complete discussion of the various hypotheses is beyond the scope of this paper; for brevity we present hypotheses that best seem to fit the central Virginia seismic zone data.

Zoback and Zoback (1980) summarized the contemporary stress fields in the United States. In the eastern U.S. two provinces are recognized: 1) the mid-continent province with greatest horizontal compression oriented northeasterly and 2) the Atlantic Coast province with compression oriented northwesterly. The boundary between the two provinces was thought to be in an ill-defined zone corresponding to the western margin of the Appalachian orogenic system. Subsequent unpublished work has cast doubt on the Atlantic Coast province orientations which now seem to be aligned parallel with the mid-continent province. According to the Zobacks, drag resistance to absolute plate velocity is compatible with the mid-continent stress field.

Seeber and Armbruster (1981), in a discussion of the 1886 Charleston, S.C. earthquake suggested slip on the Appalachian detachment by backsliding toward the coast. Isostatically, the Appalachian mountains on the west must continue to rise while lithospheric plate cooling and sediment loading depress the eastern margin. Thus the long term motions continue to increase the seaward slope on the detachment and may promote episodic slip in various parts of the plate. Currently the aseismic Piedmont in North Carolina may be locked or undergoing aseismic creep. Conversely the Central Virginia seismic zone and the South Carolina-Georgia zone are probably undergoing aseismic or seismic slip along or above the decollement and in ramp faults within the allochthonous block. It is noteworthy that the epicentral zone outline of the Central Virginia seismic zone (Bollinger and Sibol, 1985) has the general shape of a localized slump induced in an otherwise locked plate. This suggests a partial test of the backsliding model proposed herein. That is, if the localized slump model is correct, contraction faulting in the seismic area might be matched by low level seismicity of an extensional nature in the upslope direction. This might be tested by seismic monitoring leading to focal mechanism solutions.

Conclusions

1. The Blue Ridge in Virginia was a hinge zone of the North American continent during Eocambrian through Early Ordovician time. east of the hinge zone sedimentation occurred on attenuated continental crust that incorporated the rift-generated Evington basin. Rift stage sedimentation was dominantly Eocambrian; however, some rifting occurred into the Early Cambrian. Passive margin drift stage sedimentation spans the Early Cambrian through Early Ordovician. During this time an ocean basin existed between the eastern Piedmont volcanic terrane and the western Piedmont and Blue Ridge rift and drift sequences.

2. During the Late Ordovician (Taconic orogeny) closure of the ocean basin, collisional tectonics decapitated the continental hinge zone (ancestral Blue Ridge), and during successive Paleozoic orogenies transported it more than 100 km westward to its present position.

3. The Goochland Terrane is a westward thrust nappe of underlying North America basement that overrode the Chopawamsic metavolcanic rocks.

4. A sole fault at about 9-10 km depth probably rides in the Early Cambrian Rome Formation above an autochthonous Cambrian shelf sequence and Late Precambrian Grenville Basement. In the eastern Piedmont the sole fault is rooted in the Grenville Basement at the original site of the Blue Ridge continental hinge zone sequence.

5. Numerous sole and ramp faults occur within the allochthonous plate. Some were reactivated in extensional and strike slip regimes just prior to and during early Mesozoic opening of the Atlantic. The present westward dip of the M-discontinuity may have taken its shape at that time. If so, the Piedmont gravity high was formed in Early Mesozoic time.

6. During the Cretaceous to Recent, reactivation of selected faults continued as reverse fault motions in some areas of the Piedmont and Coastal Plain.

7. Fair to excellent correspondence of Central Virginia seismic zone hypocenters with faults identified in the subsurface from combined geologic and vibroseis data strongly imply that the Paleozoic sole and ramp faults in the allochthonous plate are moving although the high angle first motion planes suggest that the seismic breakage is in new faults that form in the hanging wall of the old thrust plates.

8. Localized backsliding in the larger Appalachian allochthon may explain the Central Virginia seismic zone.

References

Ammerman, M.L. and Keller, G.R., 1979. Delineation of Rome trough in eastern Kentucky by gravity and deep drilling data: American Association of Petroleum Geologists Bulletin, v. 53, p. 341-353.

Ando, C.J., Cook, F.A., Oliver, J.E., Brown, L.D., and Kaufman, S., 1983, Crustal geometry of the Appalachian orogen from seismic reflection studies: *in* Hatcher, R.D., Jr., Williams, Harold and Zietz, Isidore, eds., Contributions to Tectonics and Geophysics of Mountain Chains: Geological Society of America Memoir no. 158, p. 83-101.

Barosh, P.J., 1981. Cause of seismicity in the eastern United States: a preliminary appraisal; in Beavers, J.E. (ed.), Earthquakes and Earthquake Engineering; the Eastern United States: Knoxville, Tennesee, Sept. 14-16, 1981, v. 1, p. 397-417.

Bice, K.L. and Clement, S.C., 1982. A study of the feldspars of the Montpelier andesine anorthosite, Hanover County, Virginia: Geological Society of America Abstracts with Programs, v. 14, p. 5.

Bird, P. 1978. Initiation of intracontinental subduction in the Himalaya: Journal of Geophysical Research, v. 83, p. 4975-4987.

Bland, A.E., 1978. Trace element geochemistry of volcanic sequences of Maryland, Virginia and North Carolina and its bearing on the tectonic evolution of the central Appalachians: Ph.D. Dissertation, University of Kentuckey, 320 p.

Bland, A.E. and Blackburn, W.H., 1980. Geochemical studies on the greenstones of the Atlantic Seaboard Volcanic Province, central and southern Appalachians: in Wones and others (eds.), The Caledonides in the U.S.A.: Virginia Polytechnic Institute and State University Memoir 2, p. 263-270.

Bland, A.E. and Brown, W.R., 1977. Evington Group(?) of the Arvonia District, Virginia: rear arc sedimentation and volcanism: Geological Society of America Abstracts with Programs, v. 9, p. 120-121.

Bobyarchick, A.R., 1978. Reconnaissance geologic setting of the Petersburg Granite and regional geologic framework for the Piedmont in southeastern Virginia; in Costain, J.K., Glover, L., III, and Sinha, A.K., Principal Investigators, Evaluating and Targeting of Geothermal Energy Resources in the Southeastern United States: VPI&SU-5648-4 Report prepared for the U.S. Dept. of Energy under Contract No. ET-78-C-05-5648, p. A2-A37.

Bobyarchick, A.R. and Glover, L., III, 1979. Deformation and metamorphism in the Hylas Zone and adjacent parts of eastern Piedmont in Virginia: Geological Society of America Bulletin, v. 90, p,. 739-752.

Bollinger, G.A., 1973. Seismicity and crustal uplift in the southeastern United States: American Journal of Science, v. 273-A., p. 396-408.

Bollinger, G. A. and Sibol, M.S., 1985 Seismicity, seismic reflection and geology of the central Virginia seismic zone: Part 1, Seismicity: Geological Society of America Bulletin v. 96, p. 49-57.

Bonini, W.E., and Woolard, G.P., 1960, Subsurface geology of North Carolina - South Carolina Coastal Plain from seismic data: American Association of Petroleum Geologists Bulletin, v. 44, p. 248-315.

Bourland, C.W. and Rigby, J.K., 1982. Sponge spicules from early Paleozoic rocks of the Carolina slate belt: in Bearce and others (eds.), Tectonic studies in the Talladega and Carolina slate belts, southern Appalachian Orogen: Geological Society of America Special Paper 191, p. 125-132.

Brennan, J.L., 1985, Interpretation of Vibroseis reflections from within the Catoctin Formation of central Virginia: M.S.Thesis, Virginia Polytechnic Institute and State University, Blacksburg, VA, 95 p.

Briggs, D.F., Gilbert, M.C., and Glover, L., III, 1978. Petrology and regional significance of the Roxboro Metagranite, North Carolina: Geological Society of America Bulletin, v. 89, p. 511-521.

Brown, W. R., 1969. Geology of the Dillwyn quadrangle, Virginia: Virginia Division of Mineral Resources Report of Investigations 10, 77 p.

Brown, W.R., 1970. Investigations of the sedimentary record in the Piedmont and Blue Ridge of Virginia; in Fisher, G.W. and others (eds.), Studies of Appalachian Geology: Central and Southern, Interscience, p. 335-349.

Brown, W.R., 1973. Evidence in Virginia for Precambrian separation of North America from Africa: Geological Society of America Abstracts with Programs, v. 5, p. 381-382.

Brown, W.R., 1976. Tectonic melange(?) in the Arvonia slate district of Virginia: Geological Society of America Abstracts with Programs, v. 8, p. 142.

Calver, J.L., 1963. Geologic map of Virginia, Department of Conservation and Economic Development, Division of Mineral Resources, Charlottesville, VA.

Clark, H.B., Costain, J.K. and Glover, L., III, 1978. Structural and seismic reflection studies of the Brevard ductile deformation zone near Rosman, North Carolina: American Journal of Science, v. 278, p. 419-441.

Clement, S.C., and Bice, K.L., 1982. Andesine anorthosite in the eastern Piedmont of Virginia: Geological Society of America Abstracts with Programs, v. 14, p. 10.

Cloud, P., Wright, J.E. and Glover, L., III, 1976. Traces of animal life from 620 million-year old rocks in North Carolina: American Scientist, v. 64, p. 396-406.

Colton, G.W., 1970. The Appalachian basin - its depositional sequences and their geologic relationships; in Fisher, G.W. and others (eds.), Studies of Appalachian Geology: Central and Southern, Interscience, p. 5-47.

Conley, J.F., 1978. Geology of the Piedmont of Virginia - interpretations and problems; in Virginia Division of Mineral Resources, Contributions to Virginia geology - III: Virginia Division of Mineral Resources Publication 7, p. 115-149.

Cook, F.A., Albaugh, D.S., Brown, L.D., Kaufman, S., Oliver, J.E. and Hatcher, R.D., Jr. 1979. Thin-skinned tectonics in the crystalline southern Appalachian; COCORP seismic-reflection profiling of the Blue Ridge and Piedmont: Geology, v. 7, p. 563-567.

Çoruh, C., Bollinger, G.A., and Costain, J.K., in press, Seismogenic Structures in the Central Virginia Seismic Zone, *Geology*.

Drake, A.A., Jr. and Morgan, B.A., 1981. The Piney ranch Complex - a metamorphosed fragment of the central Appalachian ophiolite in northern Virginia: American Journal of Science, v. 281, p. 484-508.

Engelder, T., 1982. Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America?: Tectonics, v. 1, p. 161-177.

Espenshade, G.H., 1954. Geology and mineral deposits of the James River-Roanoke River manganese district, Virginia: United States Geological Survey Bulletin 1008, 155 p.

Espenshade, G.H., 1970, Geology of the northern part of the Blue Ridge anticlinorium: *in* Fisher, G.W., Pettijohn, F.J., Reed, J.C., and Weaver, K.N., eds., Studies of Appalachian geology: Central and southern, New York, Wiley-Interscience, p. 199-211.

Espenshade, G.H., and Clarke, J.W., 1976, Geology of the Blue Ridge anticlinorium in northern Virginia: Geological Society of America Northeast-Southeast Sections Joint Meeting Field Trip Guidebook No.5, 26 p.

Evans, N.H., 1984. Late Precambrian to Ordovician orogenesis in the Blue Ridge and Western Piedmont Virginia Appalachians: Ph.D. Dissertation, Virginia Polytechnic Institute and State University.

Farrar, S.S., 1982. The Goochland granulite terrane, eastern Piedmont Virginia: Petrographic evidence of Grenville granulite facies and Alleghanian amphibolite facies metamorphism: Geological Society of America Abstracts with Programs, v. 14, p. 17.

Farrar, S.S., 1984. The Goochland granulite terrane: Remobilized Grenville basement in the eastern Virginia Piedmont: Geological Society of America Special Paper 194, p. 215-227.

Farrar, S.S., Glover, L., III, and Costain, J.K., 1983. An alternate tectonic model for the eastern Appalachian Piedmont suggested by a decollement between the Goochland terrane and the Carolina-Eastern slate belts: Geological Society of America Abstracts with Programs, v. 15, p. 570.

Fisher, G. W., Higgins, M.W. and Zietz, I., 1979. Geological interpretations of aeromagnetic maps of the crystalline rocks in the Appalachians, Northern Virginia to New Jersey: Report of Investigations No. 32, Dept. of Natural Resources, Maryland Geological Survey, p. 1-43.

Fitcher, L.S., and Diecchio, R.J., 1986, Stratigraphic model for timing the opening of the proto-Atlantic Ocean in northern Virginia: Geology, v. 14, p. 307-309.

Gates, A.E., 1981. Geology of the western boundary of the Charlotte belt at Brookneal, Virginia: Masters Thesis, Virginia Polytechnic Institute and State University, Blacksburg, Virginia, 88 p.

Gates, A.E., Simpson, C., and Glover, L., III, 1986a, Appalachian carboniferous dextral strike-slip faults: an example from Brookneal, Virginia: Tectonics, v. 5, p. 119-133.

Gathright, T.M., III, Henika, W.S., and Sullivan, J.L., III, 1977. Geology of the Waynesboro East and Waynesboro West quadrangles, Virginia: Virginia Division of Mineral Resources Publication 3, 53 p.

Glover, L., III and Sinha, A.K., 1973. The Virgilina deformation, a late Precambrian to early Cambrian(?) orogenic event in the Central Piedmont of Virginia and North Carolina: American Journal of Science, v. 273A, p. 234-251.

Glover, L., III, 1974. Speculations on the relation between eastern and western Piedmont volcanism: Geological Society of America Abstracts with Programs, v. 6, p. 757.

Glover, L., III, Mose, D.G., Poland, F.B., Bobyarchick, A.R., and Bourland, W.C., 1978. Grenville basement in the eastern Piedmont of Virginia: implications for orogenic models: Geological Society of America Abstracts with Programs, v. 10, p. 169.

Glover, L., III, Poland, F.B., Tucker, R.D., and Bourland, W.C., 1980. Diachronous Paleozoic mylonites and structural heredity of Triassic-Jurassic basins in Virginia: Geological Society of America Abstracts with Programs, v. 12, no. 4, p. 178.

Glover, L., III, Mose, D.G., Costain, J.K., Poland, F.B., and Reilly, J.M., 1982. Grenville basement in the eastern Piedmont of Virginia: a progress report: Geological Society of America Abstracts with Programs, v. 14, p. 20.

Glover, L., III, Costain, J.K., Çoruh, C., and Farrar, S.S., 1983. The Eocambrian-Early Ordovician margin of North America and the Taconic suture in Virginia: Geological Society of America Abstracts with Programs v. 15, p. 582.

Glover, L., III, Speer, J.A., Russell, G.S., and Farrar, S.S., 1983. Ages of regional metamorphism and ductile deformation in the central and southern Appalachian: Lithos, v. 16, p. 223-245.

Grow, J.A., 1981. The Atlantic margin of the United States: in Bally, A.W. and others (eds.), Geology of Passive continental Margins: History, Structure and Sedimentologic Record (with special emphasis on the Atlantic margin): American Association of Petroleum Geologists, Eastern Meeting Education Course Note Series 19.

Hamilton, R.M., 1981. Geologic origin of eastern United States seismicity; in Beavers, J.E. (ed.), Earthquakes and earthquake engineering: the eastern United States: Knoxville, Tennessee, Sept. 14-16, 1981, v. 1, p. 3-23.

Harris, C.W., and Glover, L., III, 1983. An unconformity in the Carolina slate belt of central North Carolina: New evidence for the areal extent of the ca 600 Ma Virgilina Deformation: Geological Society of America Abstracts with Programs, v. 15, p. 111.

Harris, L.D., and Bayer, K.C., 1979, Sequential development of the Appalachian orogen above a master decollement - A hypothesis: Geology, v. 7, p. 568-572.

Harris, L.D., Harris, A.G., DeWitt, W., Jr., and Bayer, K.C., 1981. Evaluation of southeastern overthrust belt beneath Blue Ridge-Piedmont thrust: American Association of Petroleum Geologists Bulletin, v. 65, p. 2497-2505.

Harris, L.D., DeWitt, W., Jr., and Bayer, K.C., 1982. Interpretative seismic profile along Interstate I-64 from the Valley and Ridge to the Coastal Plain in central Virginia: United States Geological Survey Oil and Gas Investigations Chart OC-123.

Henika, W.S., 1977, Geology of the Blairs, Mount Herman, Danville, and Ringgold quadrangles, Virginia: Virginia Division of Mineral Resources Publication 2, 45 p.

Higgins, M.W., Sinha, A.K., Tilton, G.R., and Kirk, W.S., 1971. Correlation of metavolcanic rocks in the Maryland, Delaware and Virginia Piedmont: Geological Society of America Abstracts with Programs, v. 3, p. 320.

Iverson, W.P. and Smithson, S.B., 1982. Master decollement root zone beneath the southern Appalachian and crustal balance: Geology, v. 10, p. 241-245.

James, D.E., Smith, T.J. and Steinhart, J.S., 1968. Crustal structure of the middle Atlantic States: Journal of Geophysical Research, v. 73, p. 1983-2007.

Johnson, S.S., 1975, Bouguer gravity in southeastern Virginia: Virginia Division of Mineral Resources Report of Investigations no. 39, 41 p.

Keller, M.R., Robinson, E.S., and Glover, L., III, 1985, Seismicity, seismic reflection and geology of the central Virginia seismic zone: Part 3: Gravity: Geological Society of America Bulletin, v. 96, p. 49-57.

Keller, M.R., Robinson, E.S., and Glover, L., III, 1986, Seismicity, seismic reflection and geology of the central Virginia seismic zone: Part 3: Gravity. Reply: Geological Society of America Bulletin v. 97, p. 1285-1287.

Manspeizer, W., 1981. Early Mesozoic basins of the central Atlantic passive margin: in Bally and others (eds.), Geology of passive continental margins: history, structure and sedimentologic record (with special emphasis on the Atlantic margin): American Association of Petroleum Geologists, Eastern Meeting education Course Note series 19.

May, P.O., 1971. Pattern of Triassic-Jurassic diabase dikes around the North Atlantic in the context of predrift position of the continents: Geological Society of America Bulletin. v. 82, p. 1285-1292.

McKeown, F.A., 1978. Hypothesis: many earthquakes in the central and southeastern United States are causally related to mafic intrusive bodies: Journal of Research, United States Geological Survey, v. 6, p. 41-50.

Mixon, R.B., and Newell, W.L., 1977. Staffort fault system: structures documenting Cretaceous and Tertiary deformation along the Fall Line in northeastern Virginia : Geology, v. 5, p. 437-440.

Mose, D.G. and Nagel, M.S., 1982. Plutonic events in the Piedmont of Virginia: Sedimentary Geology, v. 23, p. 25-39.

Munsey, J.W., 1984, Focal mechanisms for recent (1978-1984) Virginia Earthquakes: unpublished Master of Science Thesis, Virginia Polytechnic Institute and State University, Blacksburg, Virginia, 214 p.

Nelson, W.A., 1962. Geology and mineral resources of Albemarle County: Virginia Division of Mineral Resources Bulletin 77, 92 p.

Odom, A.L., and Russell, G.S., 1975. The timing of regional metamorphism in the Inner Piedmont, North Carolina and Smith River Allochthon: inferences from whole-rock ages: Geological Society of America Abstracts with Programs, v. 7, p. 52.

Pavlides, L., 1976. Guidebook for Field Trips 1 and 4, Piedmont Geology of the Fredericksburg, Virginia Area and Vicinity: Geological Society of America Meeting, Arlington, Virginia.

Pavlides, L., 1981. The central Virginia volcanic-plutonic belt: an island arc of Cambrian(?) age: United States Geological Survey Professional Paper 1231-A, 34 p.

Pavlides, L., Arth, J.C., Daniels, D.L., and Stern, T.W., 1982. Island-arc, back-arc, and melange terranes of northern Virginia; tectonic temporal and regional relationships: Geological Society of America Abstracts with Programs, v. 14, p. 584.

Pratt, T.L., Çoruh, C., and Costain, J.K., 1987, Lower crustal reflections in central Virginia, *in* Deep Seismic Reflection Profiling of the Continental Lithosphere, edited by D. Maththews and C. Smith, Special Issue of the Geophysical Journal of the Royal Astronomical Society, v. 89, 163-169.

Pratt, T.L., Çoruh, C., Costain, J.K., and Glover, III, L., in press, A Geophysical Study of the Earth's Crust in Central Virginia: Implications for Appalachian Crustal Structure, *J. Geophysical Research*.

Ragland, P.C., Hatcher, R. D., Jr., and Wittington, D., 1983. Juxtaposed Mesozoic diabase dike sets from the Carolinas: a preliminary assessment: Geology, v. 11, p. 394-399.

Rankin, D., 1970. Stratigraphy and structure of Precambrian rocks in northwestern North Carolina: in Fisher and others (eds.), Studies of Appalachian Geology: Central and Southern: John Wiley Interscience, p. 227-245.

Rankin, D.W., 1975. The continental margin of eastern North America in the southern Appalachians: the opening and closing of the proto-Atlantic Ocean: American Journal of Science, v. 275-A, p. 298-336.

Redden, J.A., 1963. Stratigraphy and metamorphism of the Altavista area; in Weinberg, E. L. and others (eds.), Geological Excursions in Southwestern Virginia: Virginia Polytechnic Institute Engineering Extension Service, Geological Guidebook 2, p. 77-99.

Reed, J. C., Jr. 1955. Catoctin Formation near Luray, Virginia: Geological Society of America Bulletin, v. 64, p. 871-896.

Reinhardt, J., 1974. Stratigraphy, sedimentology and Cambro-Ordovician paleogeography of the Frederick Valley, Maryland: Department of Natural Resources Report of Investigations No. 23, Maryland Geological Survey, p. 1-74.

Rogers, J.J.W., 1982. Criteria for recognizing environments of formation of volcanic suites; application of these criteria to volcanic suites in the Carolina slate belt: Geological Society of America Special Paper 191, p. 99-107.

St. Jean, J., Jr., 1973. A new Cambrian trilobite from the Piedmont of North Carolina: American Journal of Science, v. 273A, p. 196-216.

Samson, S. L., Secor, D.T., Jr., Snoke, A.W., and Palmer, A.R., 1982. Geologic implications of recently discovered middle Cambrian trilobites in the Carolina slate belt: Geological Society of America Abstracts with Programs, v. 14, p. 607.

Seeber, L. and Armbruster, J.G., 1981. The 1886 Charleston, South Carolina earthquake and the Appalachian detachme it: &jgs., v. 86, p. 7874-7894.

Sinha, A.K. and Glover, L., II. 1978. U-Pb systematics of zircons during dynamic metamorphism - a study from the Brevar.' fault zone: Contributions to Mineralogy and Petrology, v. 66, p. 305-310.

Sinha, A.K., Hanan, B.B., Sans, J.R. and Hall, S.T., 1980. Igneous rocks of the Maryland Piedmont: indicators of crustal evolution: in Wones and others (eds.), The Caledonides in the U.S.A., Virginia Polytechnic Institute and State University Memoir 2, p. 131-135.

Smith, J.W., Milici, R.C. and Greenberg, S.S., 1964. Geology and mineral resources of Fluvanna County: Virginia Department of Conservation and Economic Development, Division of Mineral Resources Bulletin 79.

Stose, A.J. and Stose, G.W., 1946. Geology of Carroll and Frederick Counties: Maryland Department of Geology, Mines and Water Resources, Carroll and Frederick Counties Report, p. 11-131.

Swanson, M.T., 1982. Preliminary model for an early transform history in central Atlantic rifting: Geology, v. 10, p. 317-320.

Sykes, L.R., 1978. Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation: Reviews of Geophysics and Space Physics, v. 16, p. 621-688.

Tobisch, O.T. and Glover L., III, 1971. Nappe formation in part of the southern Appalachian Piedmont: Geological Society of America Bulletin, v. 82, p. 2209-2230.

Watts, A.P., 1981. The U.S. Atlantic continental margin: subsidence history, crustal structure and thermal evolution, in Bally, A.W. (ed.), Geology of Passive Continental Margins, American Association of Petroleum Geologists Education Course Notes series 19, p. 2/i-2/75.

Wehr, F.L., 1982. Marine-nonmarine transition in the lower Lynchburg Group, Late Precambrian, Virginia: Geological Society of America Abstracts with Programs, v. 14, p. 644.

Wehr, F. L. 1983. Geology of the Lynchburg group, Virginia: Unpublished Ph.D. Dissertation, Virginia Polytechnic Institute and State University, Blacksburg, Virginia, 244 p.

Wehr, F. and Glover, L., III, 1985. Stratigraphy and tectonics of the Virginia-North Carolina Blue Ridge: Evolution of a Late Proterozoic - Early Paleozoic Hinge Zone, Geological Society of America Bulletin, v. 96, p. 285-295.

Williams, Harold, (compiler), 1978, Tectonic Lithofacies Map of The Appalachian orogen: Map No. 1a, Memorial University of Newfoundland, St. John's, Newfoundland, Canada,

Wright, J.E., Sinha, A.K., and Glover, L., III, 1975. Age of zircons from the Petersburg granite, Virginia; with comments on belts of of plutons in the Piedmont: American Journal of Science, v. 275, p. 848-856.

Zoback, M.L. and Zoback, M., 1980. State of stress in the conterminous United States: Journal of Geophysical Research, v. 85, p. 6113-6156.

ACKNOWLEDGEMENTS

This work was supported primarily under Nuclear Regulatory Commission Contract No. NRC-04-75-237 to Lynn Glover, III, John K. Costain, and Cahit Çoruh. Additional support was provided from National Science Foundation Grant No. EAR-8009549-02 to Drs. Lynn Glover, III, John K. Costain, and Cahit Çoruh,and ARCO and CONOCO Fellowships through the Department of Geological Sciences to J. Brennan. John Wonderley provided technical support for the design and calibration of the pressure cell used to determine velocities. Russell Guy did the petrographic modal analysis of the Catoctin Formation. Professor W. D. Lowry guided JKC to field locations where representative samples of Blue Ridge and Valley and Ridge rocks were collected for laboratory determinations of velocity and density.

The authors gratefully acknowledge critical reviews by S. M. Becker, D. G. Gee, R. Wheeler, and A.E. Gates.

NRC FORM 335 (2-84) NRCM 1102, 3201, 3202

U.S. NUCLEAR REGULATORY COMMISSION

BIBLIOGRAPHIC DATA SHEET

SEE INSTRUCTIONS ON THE REVERSE

1. REPORT NUMBER (Assigned by TIDC, add Vol. No., if any)

NUREG/CR-5123

2. TITLE AND SUBTITLE

Studies of the Pattern and Ages of Post-Metamorphic Faults in the Piedmont of Virginia and North Carolina

3. LEAVE BLANK

4. DATE REPORT COMPLETED

MONTH	YEAR
December	1987

5. AUTHOR(S)

L. Glover, III, J. K. Costain, and C. Coruh

6. DATE REPORT ISSUED

MONTH	YEAR
April	1988

7. PERFORMING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code)

Department of Geological Sciences
Virginia Polytechnic Institute and State University
Blacksburg, VA 24061

8. PROJECT/TASK/WORK UNIT NUMBER

9. FIN OR GRANT NUMBER

A9027

10. SPONSORING ORGANIZATION NAME AND MAILING ADDRESS (Include Zip Code)

Division of Engineering
Office of Nuclear Regulatory Research
U.S. Nuclear Regulatory Commission
Washington, DC 20555

11a. TYPE OF REPORT

Technical

b. PERIOD COVERED (Inclusive dates)

12. SUPPLEMENTARY NOTES

13. ABSTRACT (200 words or less)

A geologic corridor from the Blue Ridge to the eastern Piedmont of Virginia is integrated into a tectonic model and extrapolated downward 10 to 15 km by means of seismic reflection and gravity studies. The Blue Ridge appears to be a hinge zone that faced a rift-generated Iapetus Ocean. An eastern continent with an Eocambrian and Cambrian magmatic arc and sediments of the same age, collided with the North American continental margin in the Middle and Late Ordovician. Subsequent Devono-Mississippian and Mississippian-Permian orogenesis continued to drive thin thrust nappes onto North America. Early Mesozoic rift basins record the beginning of the Atlantic basin and, from Middle Jurassic to Present, the margin of North America was covered by Coastal Plain sediments. Several constrained hypocenters of the central Virginia seismic zone, adjacent to a reflection profile, show an apparent relation to structure. We tentatively conclude that flat and ramp faults formed during Paleozoic nappe emplacement are currently being reactivated. The reactivation may be largely aseismic on the old thrust faults, but seismicity appears to be related to high angle transcurrent faults where new rock breakage may be occuring.

14. DOCUMENT ANALYSIS – a. KEYWORDS/DESCRIPTORS

Tectonics
Faults
Reflection Seismology
Central Virginia

b. IDENTIFIERS/OPEN-ENDED TERMS

15. AVAILABILITY STATEMENT

Unlimited

16. SECURITY CLASSIFICATION

(This page)
Unclassified

(This report)
Unclassified

17. NUMBER OF PAGES

18. PRICE

☆U.S. GOVERNMENT PRINTING OFFICE:1988-202-292:80043

UNITED STATES
NUCLEAR REGULATORY COMMISSION
WASHINGTON, D.C. 20555

OFFICIAL BUSINESS
PENALTY FOR PRIVATE USE, $300

SPECIAL FOURTH-CLASS RATE
POSTAGE & FEES PAID
USNRC
PERMIT No. G-67

120555078877 1 1AN1RA
US NRC-OARM-ADM
DIV OF PUB SVCS
POLICY & PUB MGT BR-PDR NUREG
W-537
WASHINGTON DC 20555

NUREG/CR-5123 STUDIES OF THE PATTERN AND AGES OF POST-METAMORPHIC FAULTS IN THE PIEDMONT OF VIRGINIA AND NORTH CAROLINA APRIL 1988