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Piedmont Seismic Reflection Study: A Program Integrated with Tectonics to Probe the Cause of Eastern Seismicity

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Abstract

A new tectonic model of the Appalachian orogen indicates that one, not two or more, terrane boundaries is present in the Piedmont and Blue Ridge of the central and southern Appalachians. This terrane boundary is the Taconic suture, it has been transported in the allochthonous Blue Ridge/Piedmont crystalline thrust nappe, and it is repeated at the surface by faulting and folding associated with later Paleozoic orogenies. The suture passes through the lower crust and lithosphere somewhere east of Richmond. It is spatially associated with seismicity in the central Virginia seismic zone, but is not conformable with earthquake focal planes and appears to have little causal relation to their localization.

A velocity and Q study in central Virginia implies that the gross mineralogy at depth in the upper crust is free of hydrous phases.

Subsurface structure in the central Virginia seismic zone differs in several ways from that along strike in the aseismic Roanoke River traverse. The metamorphic Blue Ridge/Piedmont plate probably overlies carbonates and clastics in both areas, but the metamorphic plate is 9 km thick in the central Virginia seismic zone but only 3 km thick in the Roanoke River traverse. As estimated by the amount of rollover (westward slumping during the Mesozoic), the central Virginia seismic zone may be more pervasively broken by distributed high angle normal faults than is the Roanoke River area. This implies greater access to deep upper crustal crystalline rocks by groundwater. Deeper penetration by groundwater may reduce the yield point of rock under stress and shorten the period of seismicity. This implies that the central Virginia seismic zone is localized by groundwater access. A corollary may be that the aseismic areas have very long period (>500 to 5000 ? years) seismicity and earthquakes of greater magnitude.

Focal mechanism planes of Munsey and Bollinger (1985) have attitudes of, 1) NW to NNW strike and steep NE or SW dips, or 2) ENE to NE strike and steep NW or SE dips. These planes are all at rather high angles to Paleozoic structure and would seem unrelated to it. The NNW set is somewhat concordant with the strike of Mesozoic dikes in the area but not with their dip.

Focal plane solutions in the Appalachians commonly give both northwesterly and northeasterly striking p-axes. Because it is unlikely that the same rock volume could transmit two distinct p-axes, one or both of them may be wrong.

Single seismic event p-axes are dependent only on the orientations of the focal planes which may be strongly influenced by crustal anisotropies (McKenzie, 19[illegible]). The focal planes and slip axes are the more likely to be real. Preliminary attempts to fit a single regional p-axis to all of the planes of Munsey and Bollinger (1985) gives an apparently good fit for a N55°E trending p-axis. This is approximately parallel with the dominant NE regional p-axis west of the Appalachians.

The best fit focal planes are oriented generally ENE, dip NW and SE steeply and are not concordant with any geologic structure in the area.

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Part A: A new tectonic model for the central and southern Appalachians.

By Lynn Glover, III

Preface

Current tectonic models lack agreement on the number and locations of continental sutures in the Appalachians. These first order structural features have been thought to exert some control on the seismicity of the region. Therefore it seems prudent to determine where these lithospheric-plate-bounding zones are when searching for the cause of localization of eastern seismicity.

The tectonic model for the Appalachians presented herein differs from existing models in four important respects; 1) there is a large uplift of 1Ga Grenville basement in the eastern Piedmont of VA. 2) Only one suture (Taconic) is recognized in the exposed Appalachians, and that separates the Carolina (Avalon) magmatic terrane from the Laurentian passive margin. 3) The Chopawamsic/ James Run volcanic belt is recognized as a part of Carolinia/Avalonia, and is not a different island arc. 4) The eastern margin of Laurentia (and its upper bounding surface, the Taconic suture) extends in the subsurface below the coastal plain at least 50 kilometers east of Richmond

Introduction

Bird and Dewey (1970) produced the first comprehensive modern tectonic model that included the central and southern Appalachians. It was essentially an extrapolation of northern Appalachian and Newfoundland data into the southeast. However, a model based primarily on northern Appalachian geology didn't seem to fit the central and southern Appalachians and, in 1972 Robert D. Hatcher, Jr., attempted the first comprehensive tectonic model for the southern Appalachians. His model proposed that the eastern Piedmont volcanics, (Charlotte, Carolina slate, Raleigh, and eastern slate belts, Figure 1) represented a late Precambrian to Early Ordovician island arc on the eastern edge of Laurentia. Westward subduction of oceanic crust was presumed to have generated an Andean-type orogeny during the Middle Ordovician-Silurian. Mid-Late Devonian to Permian collision with Africa resulted from continued westward subduction and produced the Acadian and Alleghanian orogenies.

Odom and Fullagar (1973) and Rankin (1975) suggested models in which the Brevard zone along the eastern Blue Ridge was a suture

Rodgers (1972) suggested that the Carolina slate belt rocks were part of Avalonia and probably developed as an island arc on oceanic crust far from Laurentia. During the Taconic they were thought to have collided with Laurentia.

Glover and Sinha (1973) noted that the Carolina slate belt had affinities with magmatic arcs on continental crust. However, they also noted that because volcanic detritus is absent in the early Paleozoic shelf rocks of Laurentia, it is unlikely that the partly coeval Carolina slate belt volcanics were deposited on or adjacent to the Laurentian continent. From this they concluded that the western edge of the Kings Mountain/ Char-

Figure 1. Geologic belts of the central and southern Appalachians. From Glover, 1989.

lotte/Carolina slate belt magmatic arc at its juncture with the Inner Piedmont was probably the locus of the suture which resulted from eastward subduction and collision (Figures 1, 2). This collision may have closed a back arc basin or a major ocean basin in the Middle and Late Ordovician.

Hatcher (1978) revised his earlier model, increasing the number of sutures from one to three. The basements of the Inner Piedmont and Charlotte/slate belts (Figure 1) were presumed to be continental fragments rifted from Laurentia between 800 and 700 Ma. Westward subduction draped the outer fragment with Charlotte/slate belt volcanics from about 700 to ca 450 Ma. Simultaneously the oceanic basins between the Laurentian continent/Inner Piedmont fragment and between the Inner Piedmont and Charlotte belt/slate belt fragments were closing, culminating in the Taconic orogeny during Middle/Late Ordovician. Continued westward subduction of oceanic crust beneath the Charlotte/slate belt closed the Iapetan Ocean until continental collision with Africa took place in the Acadian/Alleghanian orogenies during the Late Paleozoic.

Hatcher and Odom (1980) modified the 1978 model to include: Taconic collision between the Piedmont fragment and the North American craton; Acadian collision between Avalonia and the Piedmont-North American block; and Alleghanian collision between Avalonia and Africa.

The suspect terrane concept, formalized in the western North American Cordillera (Coney and others, 1980), sparked the beginning of a new tangent in the development of Appalachian tectonic models. In 1982 Williams and Hatcher published a paper on the accretionary history of the Appalachians. This paper essentially cast the Hatcher 1978 model, for the central and southern Appalachians, in the new terminology, but added several new terranes thought to possibly be bounded by suture zones. Currently, at least three papers (Rankin and others, 1989; Horton and others, 1989; Keppie and Dallmeyer, 1989) divide the central and southern Appalachians into 10's of terranes, each considered by their authors to be bounded by possible sutures!

Glover and others (1983), reporting on the ages of ductile deformation and metamorphism in the central and southern Appalachians, concluded that the only suture in the Piedmont and Blue Ridge of the central and southern Appalachians is the Taconic suture. This suture is found along the western boundary of the Kings Mountain belt in North Carolina and extends into Virginia along the western boundary of the Charlotte belt and Chopawamsic volcanics (Figures 1, 2). Other sutures, of Acadian and/or Alleghanian ages must lie under the Atlantic Coastal Plain or offshore in basement rocks (Figure 2).

Hatcher (1987) further revised the Hatcher and Odom (1980) model to include the Penobscottian orogeny (Early Cambrian to Early Ordovician) as an early stage in the collision of the "Piedmont arc" with the North American craton.

Largely because of the mafic-ultramafic association in the eastern Blue Ridge, there has been an overwhelming inclination to view at least part of the post Grenville sequence as collisional ophiolitic melange, thus creating a terrane boundary (suture) of Precambrian to Ordovician age (Hatcher and others, 1984; Abbott and Raymond, 1984; Conley, 1985; Hatcher, 1989; Rankin *in* Rankin and others, 1989; Horton and others, 1989, Stanley and Ratcliffe, 1985).

Glover (1989) presents a new model, from which this paper is extracted, showing that the Taconic suture is the only suture in the exposed central and southern Appala-

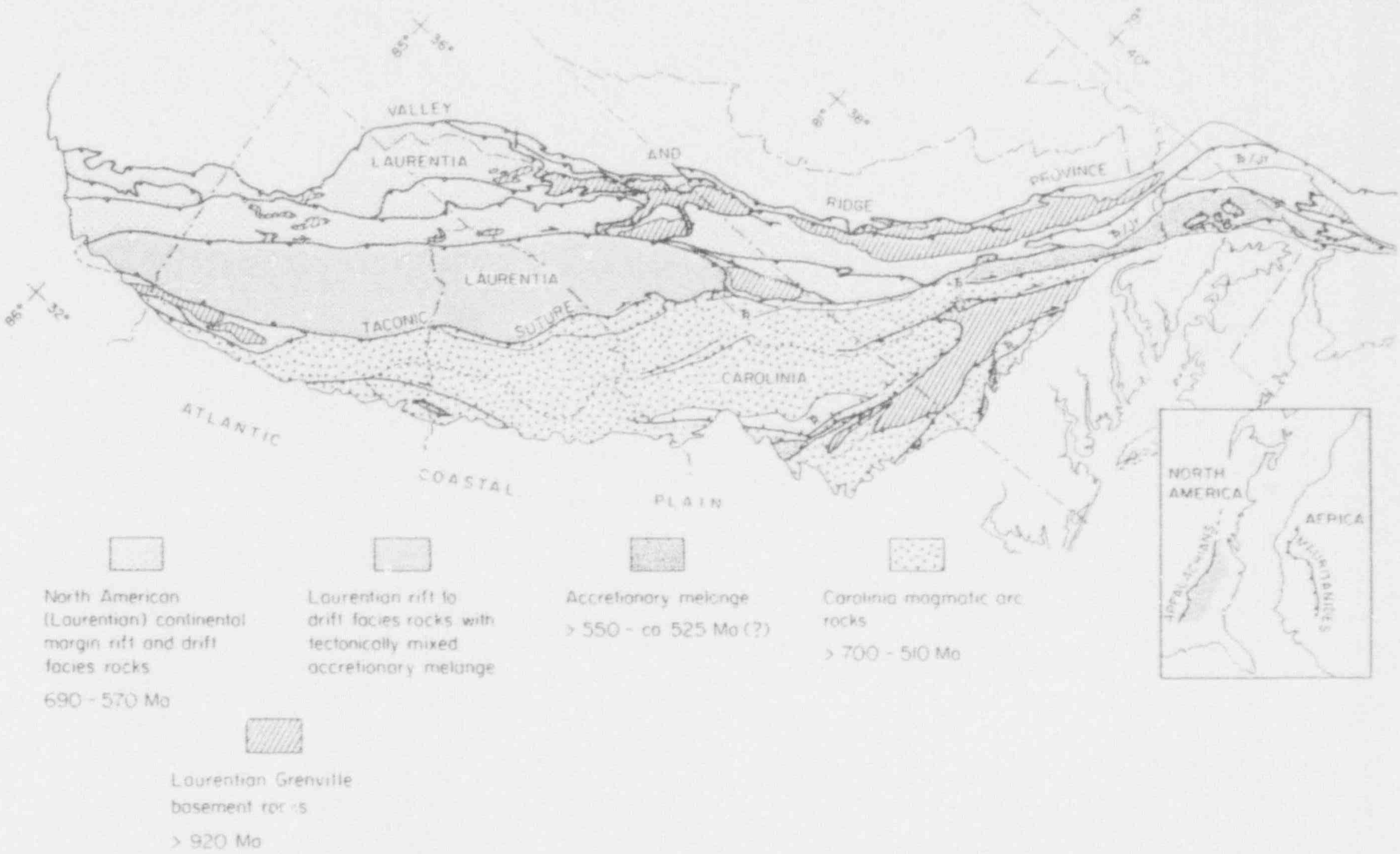

Figure 2. Tectonic map of the central and southern Appalachians. From Glover, 1989

chians.

Regional Tectonics

Blue Ridge Grenville Basement. Felsic and intermediate gneisses of the Grenville province comprise the oldest rocks found in the central and southern Appalachians. These billion-year-old rocks, the basement upon which the Laurentian part of the Appalachian orogenic system was assembled, crop out along the axis of the Blue Ridge in Virginia (Sinha and Bartholomew, 1984) and reappear in the eastern Piedmont where they are known as the Goochland "terrane" (Glover and others, 1978; Farrar, 1984). Although it is well established that the Grenville lies with profound unconformity (Nelson, 1932) below Late Precambrian conglomeratic sandstones (rift facies of Wehr and Glover, 1985) that comprise the oldest strata of the Appalachian system, these basement rocks remain the least understood of all geologic units exposed within the Appalachians. The recent state of Appalachian Grenville knowledge was summarized in a symposium volume (Bartholomew, editor, 1984).

In the central Virginia Blue Ridge the Rockfish Valley fault divides the Grenville basement into two massifs of contrasting lithology (Bartholomew, 1977; Bartholomew and others, 1981): the Pedlar massif west of the fault and the Lovingston massif east of the fault. The Pedlar contains granulite facies, massive pyroxene granofels and layered gneisses with a slight overprint of low grade metamorphic minerals. The Lovingston appears to represent a similar suite of rocks with a more intense greenschist metamorphic and deformational overprint of Paleozoic age. Bartholomew and others (1981), and Sinha and Bartholomew (1984) consider the Lovingston and Pedlar to represent massifs metamorphosed during the Precambrian at shallower, and deeper P-T conditions respectively, and to have been juxtaposed at their present structural levels during Paleozoic orogenesis. Evans (1984) made the interpretation that both massifs were originally at granulite facies during the Precambrian, and that the Lovingston massif was retrograded, largely to greenschist facies, during the Paleozoic.

Pettingill and others (1984) report Rb/Sr whole rock ages of orthogneisses ranging from about 1009 to 1021 Ma. Sinha and Bartholomew (1984) give zircon U/Pb ages for the Grenville orthogneisses in central Virginia ranging from 1130 to 1070 Ma. Detrital zircons suggest an older sediment source of 1870 Ma. The final metamorphism may have culminated at about 920 Ma.

Goochland Nappes of the Eastern Piedmont. These Grenville massifs (Figures 1, Plate 1), internal to the Appalachian orogen, comprise a sequence of units including from lower (older?) to higher (younger?) respectively the State Farm Gneiss, Sabot Amphibolite and Maidens Gneiss. The State Farm and possibly the Sabot were intruded by the Montpelier Anorthosite, which is similar to the Roseland Anorthosite that intruded the Blue Ridge Grenville basement just south of the latitude of this traverse (Plate 1).

The Goochland has been determined to be fault bounded along all of its contacts except where covered by early Mesozoic and younger sediments.

State Farm Gneiss (Brown, 1937; Goodwin, 1970, Poland, 1976; Reilly, 1980;

Farrar, 1984). The dominant rock type is a medium- to coarse-grained biotite-allanite monzogranite locally containing hornblende. At the type locality, a quarry on the State Farm, less deformed phases show relict plutonic textures and enclaves of more mafic rocks. Less deformed parts of the formation are massive, more deformed parts are layered. The monzogranite appears to locally grade into garnet-hornblende granodioritic to tonalitic gneiss. A relatively high titanium content in the State Farm is suggested by abundant clusters of titanite grains (Poland, 1976, Farrar, 1984). A.E. Gates, S.S. Farrar and J.G. Patterson (personal communication, 1985) report a mappable, tabular unit of pelitic garnet-biotite gneiss within the State Farm in the Hanover Academy and Montpelier quadrangles north of the James River. The State Farm appears to contain both metaigneous and metasedimentary protoliths.

Sabot Amphibolite: (Goodwin, 1970; Poland, 1976; Reilly, 1980; Farrar, 1984). The Sabot is dominantly a medium- to coarse-grained hornblende (locally with diopside cores) - plagioclase - quartz amphibolite with volumetrically minor, but abundant, thin interlayers of quartz - biotite - plagioclase and quartz - plagioclase. The amphibolite is a widely distributed tabular body that conspicuously outlines several domes in the Goochland massif along the eastern Piedmont in this part of Virginia (Plate 1).

A.E. Gates, S.S. Farrar and J.G. Patterson (personal communication, 1985) mapped a low-angle regional discordance between the base of the Sabot and the compositional layering in the underlying State Farm Gneiss. The origin of this discordance is unknown, however, it may be a fault or an angular unconformity. The upper contact of the Sabot seems to be everywhere conformable with overlying tabular units of gneiss and schist in the Maidens Gneiss. Most authors have suggested that its protolith may have been mafic volcanics of lava or pyroclastic origin.

Maidens Gneiss: (Poland, 1976; Farrar, 1984). This is an heterogeneous formation that structurally conformably overlies the Sabot (Plate 1). Its upper contact is unknown due to structural truncation or erosion. The dominant layered lithologies include garnet-biotite-quartz-plagioclase gneiss, biotite - quartz - plagioclase - K - feldspar augen gneiss, garnet - biotite - kyanite - K - feldspar - muscovite - plagioclase - quartz gneiss, biotite granitic gneiss, and lesser amounts of hornblende - diopside - plagioclase gneiss, scapolite - diopside - hornblende - K - feldspar - quartz - garnet gneiss and numerous thin calc - silicate layers. Poland (1976) concluded from a study of modal mineralogy, and the characteristics of zircon populations in the Maidens that it appeared to be a stratified volcanic and sedimentary formation. The Maidens is more feldspathic than either the Wissahickon or Lynchburg formations with which it has been compared (Poland, 1976; Reilly, 1980), and does not, compositionally or in facies succession, resemble other post-Grenville formations in the region (Glover and others, 1978). It does, however, bear some resemblance to the veined gneiss phase of the Grenville Baltimore gneiss.

Montpelier Anorthosite: (Clement and Bice, 1982; Bice and Clement, 1982). The Montpelier Anorthosite occurs along the northern edge of the State Farm dome about 16 km north of the James River where it intrudes the State Farm and may intrude the Sabot (Plate 1). The inner core of the anorthosite is coarse-grained rock composed of antiperthitic plagioclase, quartz, apatite, ilmenite, titanite, rutile and clinopyroxene partially altered to biotite and amphibole (A.E. Gates, S.S. Farrar and J.G. Patterson, personal communication, 1986). The outer zone is a foliated and lineated medium coarse-grained, recrystallized anorthosite consisting of plagioclase-microcline - quartz - ilmenite - titanite - rutile and clinopyroxene partially altered to biotite and amphibole (A.E. Gates, S.S. Farrar and J.G. Patterson, personal communication, 1986).

Metamorphism. The Goochland massif was metamorphosed to granulite facies (Farrar, 1984) at about 1 Ga (Glover and others, 1978) and then retrograded during Paleozoic metamorphic events to amphibolite facies. Relict core volumes of all formations in the massif still contain granulite assemblages in less than 10% of outcrops observed. Granulite facies assemblages include: 1) orthopyroxene + clinopyroxene + plagioclase, 2) orthopyroxene + garnet + plagioclase, 3) clinopyroxene + garnet + plagioclase. All include rutile and/or ilmenite, and all are ± quartz, K-feldspar, hornblende, and biotite. Pelitic assemblages may have K-feldspar + sillimanite + quartz ± garnet ± plagioclase. K-feldspar is perthitic orthoclase and may be the most common mineral.

According to Farrar (1984), the granulite assemblages in the Goochland are typical of assemblages forming in the range of 7.5 - 9 kb and 750°.- 800° C.

During Paleozoic metamorphic events K-feldspar + sillimanite were hydrated to muscovite + quartz + kyanite or staurolite.

The granodiorite gneiss at the State Farm quarry is composed of quartz (33%) + plagioclase (36%) + mesoperthite/microcline (13%)+ biotite (13%) + garnet (4%) + hornblende (3%) + clinopyroxene (1%) + traces of chlorite, titanite, magnetite, and zircon. The pyroxene has been replaced almost completely by coronas of hornblende around clusters of hornblende + quartz ± biotite ± garnet (Farrar, 1984).

Progressive dehydration of clinopyroxene + garnet granulite formed the present hornblende + plagioclase (Sabot) amphibolite with minor relict clinopyroxene (Farrar, 1984).

The Maidens appears to have typically formed from dehydration of K-feldspar-bearing granulites to produce gneiss with relict K-feldspar augen in a groundmass of plagioclase + biotite + quartz ± garnet ± hornblende. A more complete list of mineral assemblages in the Goochland massif may be found in Farrar (1984).

An Alleghanian amphibolite facies metamorphic event is indicated by $^{40}Ar/^{39}Ar$ on hornblende and biotite in the State Farm Gneiss in which cooling ages of 280-260 m.y. were obtained (Durant, and others, 1980; Farrar, 1984). Glover and others (1983) noted from regional considerations of metamorphic ages that the entire Piedmont experienced Ordovician (Taconic) metamorphism and probably much of it underwent "Acadian", ca. 360 Ma , metamorphism also. Thus the Alleghanian age of metamorphism may be only the last of several metamorphisms that affected the Goochland massif.

The age of the State Farm Gneiss at the State Farm quarry is 1031 ± 94 Ma by Rb/Sr whole rock analysis (Glover and others 1978, 1982), and "about 1 Ga" by zircon U/Pb (A.K. Sinha, personal communication, 1988).

The Goochland is considered to be part of the 1 Ga pre-Laurentian Grenville sequence because:

1) the entire Goochland has a similar metamorphic history including an early granulite facies event not recognized in adjacent terranes (Farrar, 1984);

2) the Goochland rocks bear a close similarity to the Grenville of the Blue Ridge,

especially in the apparently high titanium content of the State Farm, in the presence of anorthosite in both massifs, and in the granulite facies of metamorphism;

3) the rocks of th Goochland do not bear a resemblance, in lithofacies or facies succession, to other sequences in the region such as the Lynchburg or Wissahickon formations with which they have been compared;

4) Ordovician, Taconic, granulite facies rocks of the Wilmington Complex of Delaware and Pennsylvania, about 200 km to the north, are metamorphosed Cambrian volcanics unlike the Goochland rocks (Farrar, 1984). Cambrian volcanics above and below the Goochland nappes (Plate 1) are at low metamorphic grade, and

5) surface geology and crustal structure along the I-64 vibroseis line through the Goochland terrane is consistent with the Goochland being a nappe complex of the Laurentian Grenville basement emplaced during a dextral transpressional event in the late Paleozoic Alleghanian orogeny.

Late Precambrian/Cambrian Rifting and the Cambrian Rift-to-Drift Transition

Crossnore Volcanic - Plutonic Suite. Following the peak of Grenville metamorphism, at about 920 Ma, uplift and erosion deeply dissected the orogen over a period of about 230 m.y. During this time the Grenville was denuded to a depth of about 25 km (Herz, 1984), exposing granulite facies rocks.

At about 690 ± 10 m.y. ago (Odom and Fullagar, 1984, Rb/Sr ages from samples in the Mount Rogers - Grandfather Mountain area of North Carolina and southern Virginia) continental rifting began coevally with emplacement of the fluorite and sodic amphibole - bearing, peralkaline Crossnore plutonic-volcanic suite (Rankin, 1976). Non-marine and marine volcanic rocks and arkosic sandstones accumulated in rift graben. The youngest age (Rb/Sr whole-rock) of Crossnore plutonism in that area is 646 ± 9 m.y. for the Crossnore Granite itself. Odom and Fullagar found that earlier zircon U/Pb ages (Rankin and others, 1969) gave falsely older ages (820 Ma) because of contamination from old Grenville gneisses which they assimilated.

Several members of the Crossnore suite occur in the Blue Ridge of central and northern Virginia (Plate 1):

1) One of these, the fluorite - bearing Mobley Mountain Granite near Roseland, about 40 km south of Charlottesville, gives a Rb/Sr whole-rock age of 652 ±22 m.y. (Herz and Force, 1984)

2) Another, the Rockfish River pluton, located about 30 km south of Charlottesville, has yielded ages of 646 ± 55 m.y.(Mose and Nagel, 1984), and 630 Ma, (Mose and Kline, 1986)

3) A third, questionably the Robertson River granite, gave an age of ca. 650 Ma on

zircon $^{207}Pb/^{206}Pb$ analysis by T. Stern (reported in, Rankin, 1976). According to Lukert and Banks (1984), Stern's analysis was done on a riebeckite granite that intrudes the main body of the Robertson River pluton, which lies about 100 km north of Charlottesville. Lukert and Banks determined an age of 732 ± 5 m.y from a zircon U/Pb concordia intercept for the main body of the Robertson River. The zircon samples of Lukert and Banks did not appear to contain inherited older cores. Mose and Nagel (1984) determined a Rb/Sr whole-rock age of 646 ± 55 m.y. from samples spread over most of the length of the Robertson River, excluding the area of the Stern riebeckite granite. Subsequently they reported a refined Rb/Sr whole-rock age of about 650 Ma (Mose and Kline, 1986). Therefore, U/Pb and Rb/Sr ages are in disagreement by about 80 m.y. Until a more detailed zircon analysis of the Robertson River suite is undertaken to look more specifically for older inherited components, the Rb/Sr data seems more attractive. It is also worth remembering that Rb/Sr ages of deeply emplaced and slowly cooled plutons are commonly as much as 25 m.y. younger than the emplacement age because of late closure of the isotopic system. Thus, the youngest Crossnore granitoid plutons in the region of our traverse are thought to be about 650 Ma.

Although the Catoctin and all post 650 Ma igneous rocks were originally included in the Crossnore volcanic-plutonic suite by Rankin (1976), it now seems that an older granite and rhyolite - bearing suite of rocks lies unconformably below the Lynchburg and its southern equivalent the Ashe Formation. This granite-bearing suite is about 690-650 Ma and it may be best to confine usage of the term Crossnore to these older rocks. A similar argument has also been made by Badger and Sinha (1988).

Erosion has removed volcanic rocks associated with the Crossnore volcanic-plutonic suite over much of the Virginia Blue Ridge, and now cobbles of the Robertson River may be found in the basal conglomerates of the overlying Mechums River and Fauquier (Lynchburg) Formations (Lukert and Banks, 1984). This provides an older age limit of about 650 Ma for the Lynchburg Group..

Lynchburg Group. The Lynchburg Formation was named by Jonas (1927) for exposures along the James River near Lynchburg, Virginia. This sequence of rift-related; clastic rocks, basaltic volcanic rocks and shallowly emplaced ultramafic dikes and sills, crops out along the east flank of the Blue Ridge anticlinorium in Virginia (Figures 2, Plate 1). It non conformably overlies Grenville basement or, locally, rocks of the Crossnore Volcanic - Plutonic suite. The Lynchburg has been recently subdivided into five formations by Wehr (1985) in the Culpeper-Charlottesville region (Figure 3). In the Culpeper area the Group comprises a terrestrial, alluvial outwash deposit (Bunker Hill Fm.) at its base. The overlying formations, Monumental Mills, Thorofare Mountain, Ball Mountain, and Charlottesville include a deep water retrogradational fan sequence (Wehr, 1983). Details of the Bunker Hill, Monumental Mills and Thorofare Mountain Formations are briefly characterized here.

Bunker Hill Formation This formation consists of 0-1000 m of poorly sorted, medium-grained to granule feldspathic arenite with minor siltstone and mudstone. It is absent in the Rockfish River area south of Charlottesville. Facies analysis indicates deposition as a braided outwash plain adjacent to glaciated highlands composed largely

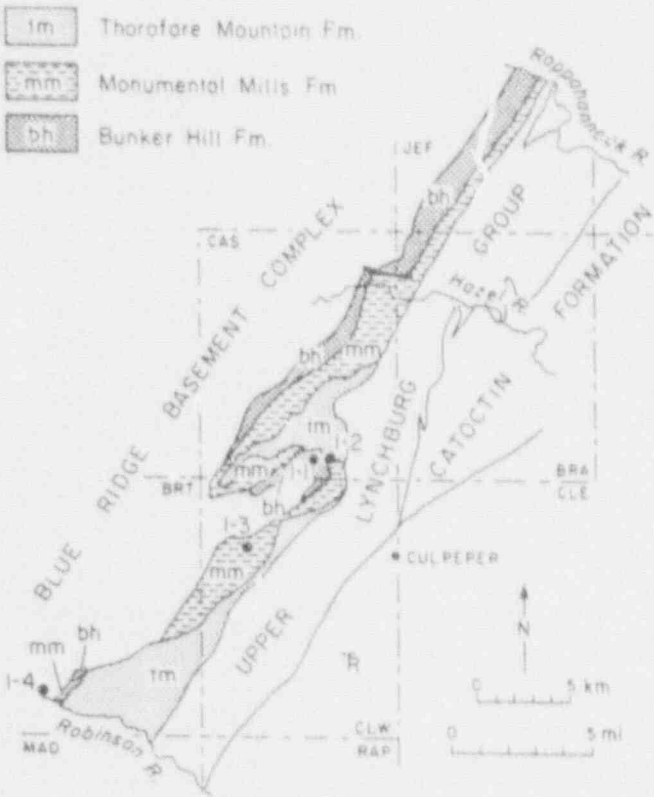

Figure 6. Subdivisions of the lower Lynchburg Group between the Rappahannock and Robinson Rivers near Culpeper, Virginia. From Wehr (1985).

Figure 3. Subdivisions of the lower Lynchburg Group between the Rappahannock and Robinson Rivers near Culpeper, Virginia. From Wehr (1985)

of Grenville basement.

Monumental Mills Formation. Wehr (1985) divided this formation into a lower sandstone member of thin bedded fine- to medium-grained, well-sorted sandstone and siltstone, and an upper member of thin bedded to laminated siltstone and mudstone. The outcrop belt of the Monumental Mills is 0-1500 m wide and thins toward the south. The Monumental Mills is absent or represented only by the Rockfish Conglomerate in the Rockfish River area south of Charlottesville. Facies analysis suggests a slope environment.

Rockfish Conglomerate. This Formation is a pebbly, feldspathic sandstone with conglomerate lenses that makes up the basal unit of the Lynchburg Group in the Rockfish River area south of Charlottesville (Plate 1). The Formation is about 500 m thick and consists of cobble conglomerate in the lower part grading upward into coarse-grained pebbly sandstone. The upper 20 m is graded thin-bedded sandstone with local occurrences of outsized clasts interpreted to be ice-rafted dropstones. The lower contact is with a mylonitic zone separating basement from the Rockfish. The upper contact is gradational into the lower Thorofare Mountain Formation.

According to Wehr (1985) most of the larger clasts are very coarse-grained light colored basement gneiss. The Rockfish also contains clasts of granite, biotite gneiss, fine-grained aplite (?), and dark siltstone. In thin section Rockfish sandstones contain detrital quartz and feldspar in a schistose matrix of quartz, plagioclase, mica and magnetite. Facies analysis (Wehr, 1985) of outcrops along the Rockfish River has shown that the outsized clasts are ice-rafted dropstones and indicates that the conglomerate was deposited as subaqueous glacial outwash.

Thorofare Mountain Formation. The Thorofare Mountain Formation is recognized from the Culpeper area to the Rockfish River (Figure 3, Plate 1). This formation consists of medium-grained to pebbly, poorly sorted feldspathic sandstone with minor conglomerate, siltstone and graphitic mudstone. Sandstones are massive to faintly stratified in beds a few cm to more than 8 m thick. Interbeds of coarsely laminated siltstone and graphitic mudstone are common, and these lithologies also occur locally as rip-up clasts in intraformational conglomerate. Facies analysis indicates that this sequence was formed in a deep water submarine fan.

Ball Mountain Formation. This sequence extends throughout the area of study by Wehr (1985) and occupies a belt 1-4 km in width. It consists of coarse-grained to pebbly quartz wackes and quartzites interbedded with laminated siltstone and graphitic mudstone. The upper 100 m is locally a graphitic schist named the Johnson Mill Member (Nelson, 1962). Over much of the area between Culpeper and the Rockfish River the Ball Mountain truncates underlying units and is either in unconformable or fault contact with them. In some places it is in conformable, and gradational, contact with the Thorofare Mountain. Facies analysis of the Ball Mountain shows that it has sedimentary characteristics similar to the underlying Ball Mountain Formation and was deposited by sediment gravity flows (Wehr, 1983,1985). The Johnson Mill Member at the top of the formation is euxinic which suggests abrupt cessation of influx of clastic material and basin-wide starvation following Ball Mountain sandstone deposition (Wehr, 1985).

Charlottesville Formation. The Charlottesville formation extends throughout the Culpeper-Rockfish River area. According to Wehr (1985) it comprises schistose siltstone and mudstone with isolated outcrops of medium- to coarse-grained, commonly amalgamated sandstone beds. Sandstone beds range from a few mm to about a meter in thickness. They tend to be massive, although grading, horizontal stratification, and complete Bouma T(a-e) sequences occur. The lower 1000 m of the formation in the Rockfish area is characterized by coarsely laminated to very thin bedded, fine grained sandstone and siltstone with prominent biotite porphyroblasts. Similar rocks occur more locally near Culpeper. Primary textures and sedimentary structures indicate deposition by turbidity currents in deep water.

Swift Run Formation. This formation occurs throughout most of the Culpeper-Rockfish River area, ranging from 0 - 5 km in width of outcrop belt (Plate 1).

On the west side of the Blue Ridge anticlinorium (Stose and Stose, 1946; Bloomer, 1950; Werner, 1966; Brown 1970) the Swift Run occurs in lenses as much as 400 m thick unconformable upon basement and grading upward by interleaving with the overlying Catoctin Basalt (Plate 1). Here it consists of cross-bedded arkose, conglomerate, mudstone and intercalations of mafic tuffs and lavas and is interpreted as alluvial in depositional environment (Gathright, 1976).

In the Culpeper-Rockfish River area, on the east side of the Blue Ridge, the Swift Run is conformable with the underlying Charlottesville Formation and is gradational over a short distance by interleaving with the base of the Catoctin Formation. In this area it contains, at the base, coarse-grained feldspathic sandstone; in the middle, greenstone, rare felsic volcanic rock, fine-grained sandstone, and graphitic mudstone; and at the top, coarse-grained blue-quartz sandstone and arkose interbedded with pale green mudstone and a few thin greenstone beds. In the Culpeper area many Swift Run sandstones are calcareous, and along the Hazel River tabular marble clasts as much as 45 cm in length occur in a coarse-grained sandstone matrix (Wehr, 1985). To the north of the Culpeper area thin lenses of marble are present below the Catoctin Formation (Furcron, 1939; Parker 1968), and these may be correlative with the limestone conglomerate in the Culpeper area.

Turbidites suggest that the Swift Run on the east side of the Blue Ridge is probably a deep water sedimentary gravity-flow deposit, in contrast to its non-marine nature to the west.

Catoctin Formation This formation was named by Keith, 1894. Metabasalts and minor intercalated siliciclastic rocks of the Catoctin Formation are abundant across the northward plunging nose of the Blue Ridge anticlinorium in southern Pennsylvania (Figure 2). From there to the south the Catoctin forms two belts of outcrop along the east and west flanks of the anticlinorium into central (Plate 1) and southern Virginia where it occurs intermittently. Thus the Catoctin is a key unit in relating the stratigraphy of the Valley and Ridge Province with that of the Piedmont.

The Catoctin comprises a sequence of greenschist-facies tholeiitic basalt lavas and minor breccias and tuffs intercalated with quartzose feldspathic sandstone and mudstone. The Catoctin is gradational over a short interval by interleaving with both overlying and underlying formations. In Pennsylvania minor rhyolite is intercalated with the basalt lavas. The total thickness of the formation may reach 1000 m (Gathright and others, 1977). Along the west flank of the Blue Ridge the Catoctin overlies the Swift Run Formation and is overlain by the Lower Chilhowee Group Unicoi/Weverton Formation. Rocks of the Unicoi and Swift Run are similar, and it is probable that in southern Virginia where the Catoctin is absent the Unicoi and Swift Run have been mapped together as Unicoi.

On the west flank of the Blue Ridge the Catoctin (including the enveloping Swift Run and Unicoi formations) is non-marine (Reed, 1955), and on the east flank the Catoctin and overlying Candler and underlying Swift Run formations are marine (Wehr and Glover, 1985). The transition from non-marine to marine takes place north of Culpeper along the east side of the Blue Ridge anticlinorium.

The Catoctin is a member of the Albemarle-Nelson suite as defined below. Blackburn and Brown (1976), Bland (1978) have shown by trace element geochemistry and petrochemistry that the Catoctin is a tholeiite related to rifting during the formation of Iapetus. Badger and Sinha (1988) dated the Catoctin by Rb/Sr whole rock and mineral isochron methods at 570 ±36 Ma. This age is consistent with the Early Cambrian and Early Cambrian (?) age deduced by Werner (1966) in central Virginia and by Simpson and Eriksson (in press) in southern and south-central Virginia from studies of the sedimentology and fauna (Simpson and Sundberg, 1987) of the rift related, basalt-bearing Unicoi Formation of the basal Chilhowee Group.

Evington Group. Rocks of the Evington Group overlie the Catoctin Formation (Plate 1), or where that is absent, the Lynchburg Group. The Evington sequence comprises the youngest Laurentian sequence known in the Piedmont of Virginia. Some of the most important earlier work may be found in Espenshade (1954), Brown (1958, 1970), and Redden (1963). These authors were uncertain about the order of stratigraphic succession in this complexly deformed and metamorphosed group of rocks. Patterson (1987a, 1987b, in press) revised the stratigraphic ordering based on mapping and structural studies in the Lynchburg area (Plate 1). Three facies sequences, proximal, distal and an eastern allochthon, were recognized. Detailed relations among these sequences are shown in Figure 4. The Slippery Creek and Mount Athos Quartzite pinch out eastward toward the distal facies where Joshua Schist was deposited directly on Candler. The Slippery Creek Greenstone and Mount Athos Quartzite pinch out to the northeast. Along strike to the southwest , the Slippery Creek pinches out, and the Mount Athos quartzite is underlain by the "Moon Mountain Greenstone" (informal name by Patterson, 1987). Still farther east, in the eastern allochthon, only Candler lithologies with interbedded greenstone and quartzite are present (Brown, 1958; Patterson, 1987a, 1987b, in press)

Candler Formation. This unit may be as much as 1.7 km thick and is dominantly siliciclastic in composition. The basal contact of the Candler is gradational over a short distance by interleaving with the underlying Catoctin Formation, or, where that is absent, is gradational with the Lynchburg Group. In the proximal facies sequence, the upper contact is gradational with the Mount Athos Quartzite. In the distal facies sequence pelites of the Joshua Schist overlie the Candler. The top of the Candler Formation in the eastern allochthon is not known.

The western, proximal, facies is composed of:

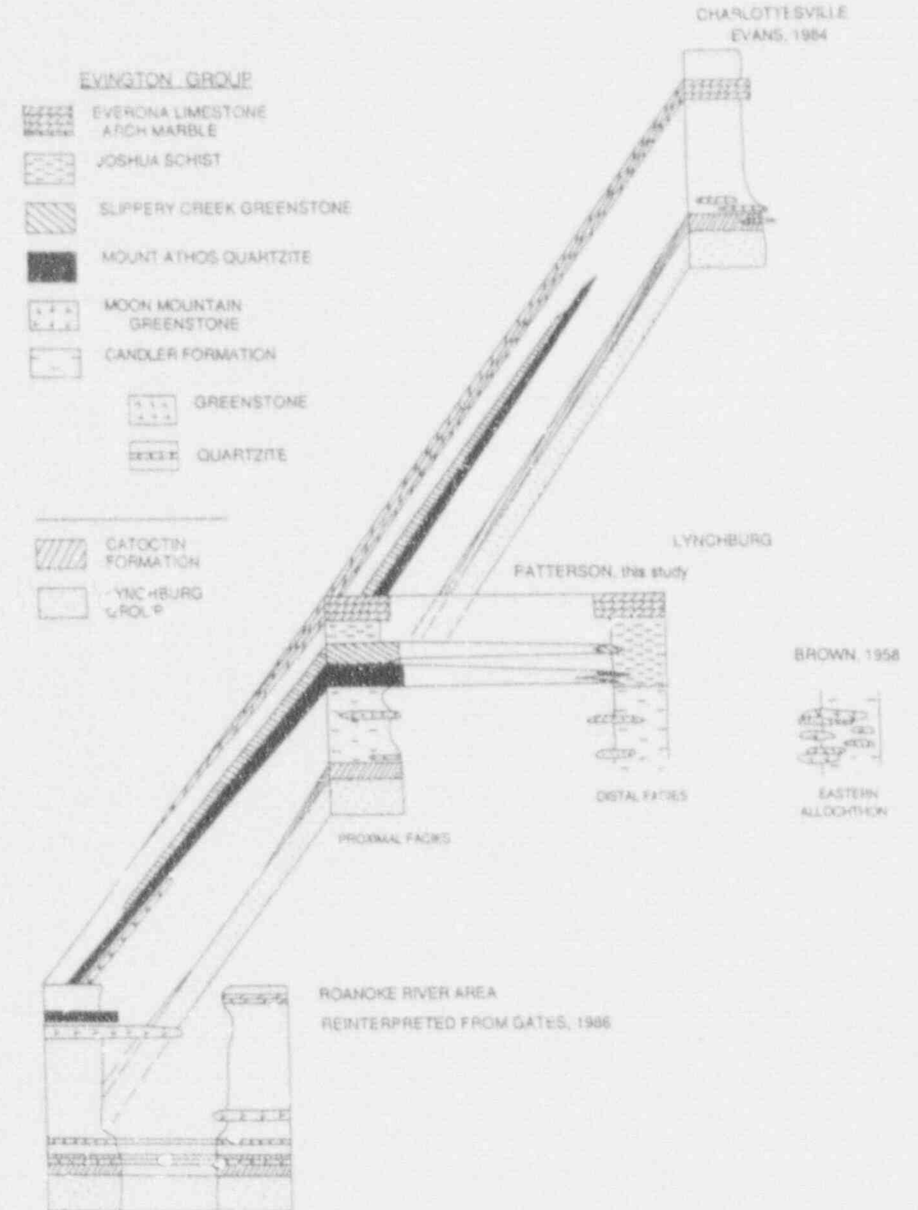

Figure 4. Stratigraphic relations in the Lynchburg Group near Lynchburg, Virginia. From Patterson (1987)

1) Sandy laminated schist: These contain sand laminae 1-2 mm thick, composed of quartz and minor feldspar. Pelite laminae are as much as 5 mm thick..

Local quartz-rich beds vary from 2.5 to 10 cm in thickness and rarely reach 50 cm. Quartz wacke: These beds may contain coarse to lower very coarse, blue quartz grains in a pelitic matrix. At one location, graded beds were found with sandstones averaging 5 mm thick, grading into pelites 5-10 mm thick. Compositionally this rock type varies from 30-50% quartz, 0-15% plagioclase, 20-30% biotite/chlorite, with local calcite.

Green chloritic phyllites and schists: These vary with increasing quartz content to a green sandy laminated lithology, which may contain as much as 50% chlorite. Thin (5 cm) layers of white marble are rare.

2) Blue weathering phyllite: This rock is composed of 10-45% chlorite, 25-50% muscovite, 5-35% quartz, and as much as 5% albite. Very rare millimeter thick sand laminations occur, and 1-2 mm diameter nodes of quartz sand are found locally.

3) Quartz arenite and subarkosic quartz arenite: This lithology may occur as lenses (75 x 530 m to 150 x 3700 m) and pods (as much as 8 m thick). Laminated lower- to middle-fine-grained quartz arenite occurs with micaceous quartzite and quartz muscovite schist. Fine grained arkosic quartz arenite occurs in medium to thin beds. Massive arkosic and subarkosic quartz arenites and matrix supported wackes contain feldspars (generally potassic) as much as 3 mm and blue quartz clasts as much as 4 mm in diameter.

4) Impure marbles and calc-silicate schists occur in lenses (75 x 115 m to 190 x 650 m), commonly along fault zones. The marble is dark grey to bluish, has a phyllitic sheen, and contains nodes and laminae of white calcite. Mica-quartz-feldspar, calcite, and mixed mica-quartz-feldspar-calcite layers, 1.5 to 5 m. thick, are gradationally interlaminated.

Still farther east, in rocks equivalent to the eastern allochthon of Patterson (1987), Brown (1958) mapped dominantly Candler facies lying above a domal structure cored by Lynchburg Group. Greenstones, commonly associated with quartz arenite, are scattered throughout this sequence which is truncated at it's top by a fault.

Patterson suggested that the alternating compositional laminations in the sandy laminated schist resulted from variable flow conditions. Phyllites (mudstones) may have been deposited from dilute turbidity currents or fair weather suspension fallout.

Map scale lenses of massive quartzose sandstones may have been deposited from sediment gravity flow. Some smaller lenses could be channel fill cut into the underlying muds.

Mount Athos Quartzite. This quartz arenite is the most resistant formation in the Evington Group and forms ridges which flank the James River Valley (Plate 1). The formation decreases in thickness and grain size toward the east. The Mount Athos is about 0.3 km thick and occurs only in the proximal facies of the Evington Group. The lower contact of the Mount Athos is gradational over a short distance by interleaving with the Candler, the upper contact is also gradational over a short distance into greenstone. Two lithologies are predominant in the western most proximal facies fault block:

1) coarse granule conglomerate with clasts of blue quartz and subordinate potassium feldspar, and

2) clast to matrix supported quartz wacke.

Primary sedimentary structures occur throughout the length of the Evington belt in central and northern Virginia. Bedding is planar to irregular, very thin to very thick. Cross stratification occurs in ripples 3 to 5 cm high. Small-scale, tabular-tangential cross stratification occurs. Trough cross strata 3 cm thick and 12 cm wide are developed in pebbly quartz arenite, with stratification by mica-rich layers. Graded bedding occurs in thicknesses varying from less than 5 to 30 cm thick.

According to Patterson (1987a) the graded bedding points to turbidity flow deposition. Massive coarse granule sandstone is structureless and may have been deposited from high density sediment gravity flow. Parallel-laminated sandstones with internal discordances relative to bedding planes, and planar bedded to lenticular sand bodies, are structures which typify hummocky cross stratification. Such structures are formed above storm wave base and below fairweather wave base by combined or oscillatory and unidirectional flow, under storm wave conditions. Tabular, tangential, trough cross stratification is produced by migration of dunes or ripples. Flowing currents which generate cross stratification are not restricted to any environment. Therefore, the Mount Athos is inferred to have been deposited by sediment gravity flows, with possibly minor sediment transport and reworking by storm generated currents.

Slippery Creek Greenstone. This metabasalt, approximately 2 km thick, contains the upper greenschist, epidote-amphibolite faces, mineral assemblage albite (15-35%) + quartz (0-14%) + hornblende (8-45%) + clinozoisite and/or zoisite (1-20%) + epidote (1-20%) + minor; titanite, magnetite and biotite. Relict plagioclase phenocrysts and amygdales are locally present. Volcaniclastic layers and quartz muscovite schist occur locally interlayered with the metabasalts. The lava sequence is depositionally conformable with the underlying Mount Athos. The upper contact was not observed in Patterson's area.

Bland (1978) gave trace element abundance data to support a rift origin and extrusion through continental crust for the Slippery Creek. The formation is therefore interpreted as a submarine lava related to rifting.

Just south of Patterson's area, near Oxford Furnace, the writer has seen enclaves that appear to be xenoliths of coarse-grained granite in the Slippery Creek. These are probably fragments of Grenville basement, suggesting that this part of the Evington Group was deposited on the continent of Laurentia, and not on oceanic crust.

Joshua Schist. This formation contains several siliciclastic lithologies and shows many well developed sedimentary structures. The lower contact with the Slippery Creek was not seen in Patterson's area. In places where the Slippery Creek is absent, the Joshua overlies the Candler gradationally. The upper contact is gradational into the overlying Arch Marble. The formation may be as much as 0.7 km thick in the area. The following rock types occur:

1) Quartz mica schist and phyllite with graded bedding. This facies is commonly graphitic. Graded bedding occurs with quartz sandstone laminae 1 mm thick capped by mica schist 0.2-0.3 mm thick. The mineralogy is; quartz (30-65%), muscovite (30-60%), graphite (0-25%), biotite (1-25%) with trace amounts of pyrite, apatite, zircon, plagioclase, tourmaline and titanite.

Soft sediment slump structures are preserved in the coarser graded beds.

2) Dark phyllite is very schistose and shows no primary sedimentary structures aside from bedding. Mineralogy is similar to that of the quartz-mica schist except that there is a lower quartz content. In thin section quartz rich laminae can be seen in the phyllite.

3) Green schist was found at two locations in the Joshua. This lithology contains: hornblende (39%), epidote (15%), biotite (15%), quartz (10%), plagioclase (15%), and minor amounts of clinozoisite and magnetite.

4) Conglomerate occurs locally in the Joshua. Angular to rounded clasts range from 1 mm to 5 cm x 1.5 cm. The clasts consist of quartz, plagioclase and potassium feldspar, micaceous quartz wacke, arkosic wacke, phyllite and dolomite. The conglomerate bodies are matrix to clast supported, with a matrix of fine grained quartz and mica.

5) Quartz wacke, quartz arenite, and calcareous quartz wacke occur in areally restricted lenses throughout the formation. Quartz arenite is rare, and occurs as small isolated outcrops of very fine grained quartzite.

Arch Marble. The Arch is generally laminated to thin bedded and locally massive. Color banding is dark and light depending on the amount of siliciclastic material in the layer. Layering commonly ranges from 0.2 mm to 5 mm in thickness. Locally graded bedding is preserved. The formation is about 0.2 Km thick in Patterson's area. The Arch is considered a deep water carbonate facies deposited by turbidity currents from sources nearer the shore face (Patterson, 1987b; Read, 1989).

Albemarle - Nelson Suite: Ultramafic Intrusive Rocks, and Mafic Dikes, Sills, Lavas and Tuffs; Late Precambrian to Early Cambrian (Post 650 pre 570 Ma)

A suite of mafic and ultramafic sills and dikes, including mafic lavas with minor felsic derivatives occur in the Lynchburg, Swift Run, Catoctin, Unicoi (Cambrian of western Blue Ridge), and upper Evington Group (Slippery Creek Greenstone). Ultramafic rocks are confined to the sequence below the Catoctin, and a set of hornblende gabbros may also be confined to the pre-Catoctin sequence. Most of these rocks were originally included in the Crossnore Plutonic - Volcanic suite of Rankin and considered to be about 820 Ma (Rankin and others, 1969; Rankin and others, 1973). Since then additional isotopic dating indicates that the felsic peralkaline plutonic and volcanic rocks of the type Crossnore are older than about 650 Ma. and unconformably underlie the Lynchburg/Ashe formations (See section on Crossnore Plutonic Volcanic suite.). This suggests that the Crossnore is distinct from the younger ultramafic and basaltic rocks and can be considered a sub-suite of the Late Precambrian Early Cambrian rift-related igneous rocks in the Blue Ridge and western Piedmont. In this report the younger mafic-ultramafic suite will be referred to the Albemarle-Nelson suite as herein modified from Burfoot (1930).

Within the region between Culpeper and Charlottesville (Figures 3, 5, 7) mapped by Wehr (1985) amphibolite dikes and sills are abundant in the basement and lower part of the Lynchburg and Swift Run sequence, but are found as well, though less commonly, in the upper part of the clastic sequence to a level just below the Catoctin greenstone. Mineralogy (Evans, 1984 of the dikes along the Rockfish River is:

plagioclase (An $_{25\text{-}35}$)+ epidote + hornblende + magnetite + quartz

All minerals are considered to be metamorphic. This would appear to be a medium grade amphibolite facies rock consistent with temperatures above 500°C. However, according t Evans (1984) the metamorphic facies of the surrounding Grenville biotite gneiss is greenschist, with garnet-biotite pairs implying a temperature of about 400° C. Evans noted that the grade of Paleozoic metamorphism decreased up section into the Lynchburg, Catoctin and Evington, none of which are reported to contain garnet.

Davis (1974), working near the same area, described a coarse-grained "hornblende metagabbro" with nearly equidimensional aggregates of metamorphic hornblende in a matrix of highly saussuritized plagioclase. Relict pyroxenes are altered partially or totally to hornblende, zoisite, magnetite and chlorite. Other metamorphic minerals are epidote, titanite, calcite, garnet, and rarely biotite. Davis did not discuss the conditions of regional metamorphism.

Reed and Morgan (1971) analyzed dikes of metabasalt in the Blue Ridge, northwest of the Rockfish area, and concluded that the compositions of the dikes were similar to those of the overlying Catoctin Formation. In Reed and Morgan's area as in the Rockfish area the Catoctin is a greenschist facies basalt derived from a dry pyroxene-bearing protolith with no evidence of hornblende in either mineral assemblage. Since Reed and Morgan's study all subsequent workers seem to have accepted that the amphibolite dikes on the southeast side of the Blue Ridge are also feeders to the Catoctin. Several geologic maps (Wehr, 1983; Brown, 1958) of segments of the belt over a distance of 120 km along strike, from Culpeper to Lynchburg, Virginia, consistently show amphibolite dikes and sills throughout the Lynchburg to within a few tens of meters of the Catoctin, yet the Catoctin is chlorite- and actinolite-bearing, and is without hornblende in these outcrops. Hornblende amphibolite occurs within 500 meters, stratigraphically below the Catoctin near Lynchburg, and the Catoctin is a biotite-bearing albite - actinolite schist of probable middle greenschist facies (Ping Wang, personal communication, 1988). If the amphibolite dikes are feeders of the Catoctin why don't they have a similar mineralogy where they are at levels of emplacement just below the Catoctin?

The Schuyler ultramafic body, in the Rockfish River area (Figure 6) is one of a number of thin, tabular ultramafic units emplaced dominantly in the upper part of the

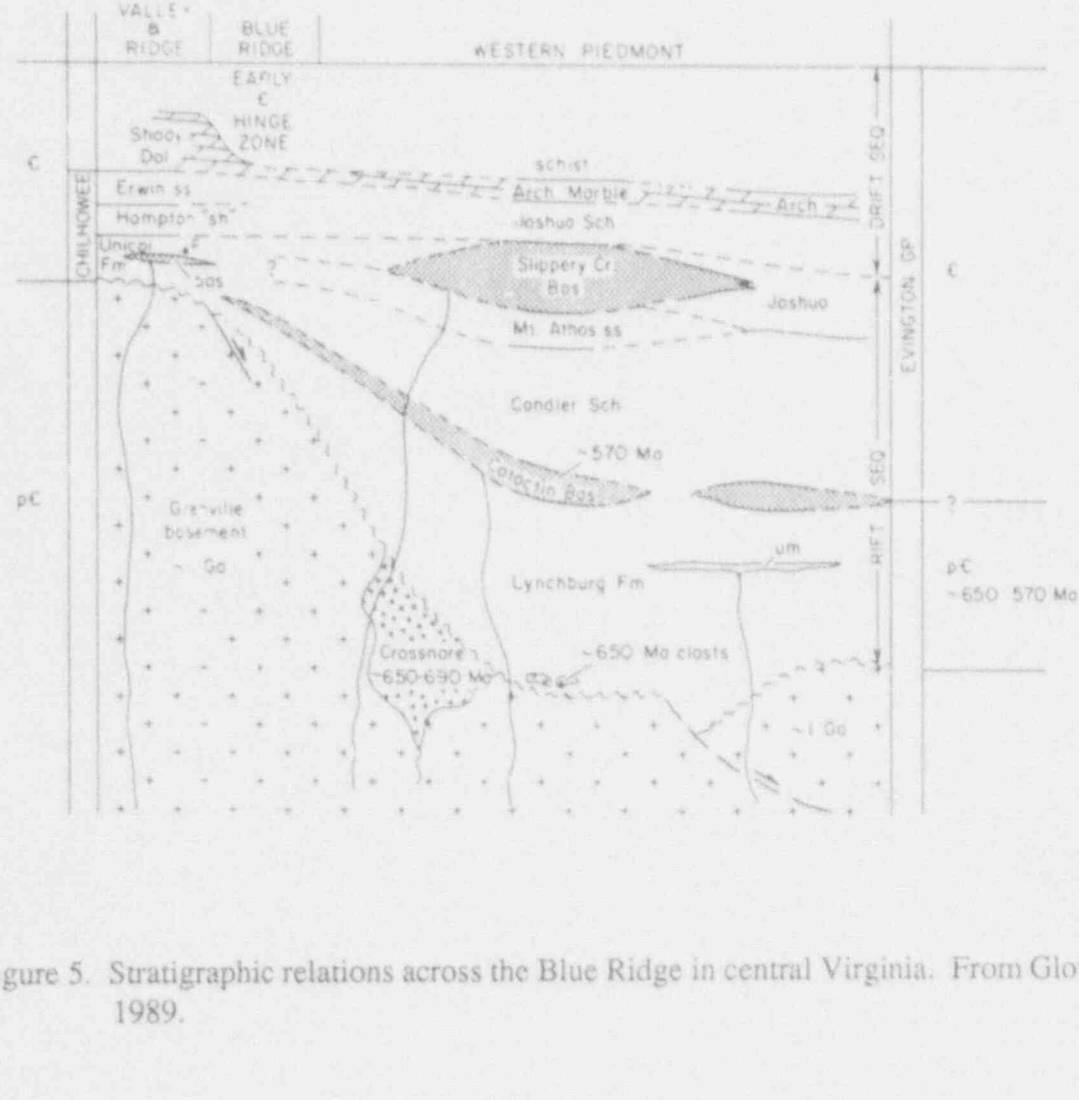

Figure 5. Stratigraphic relations across the Blue Ridge in central Virginia. From Glover 1989.

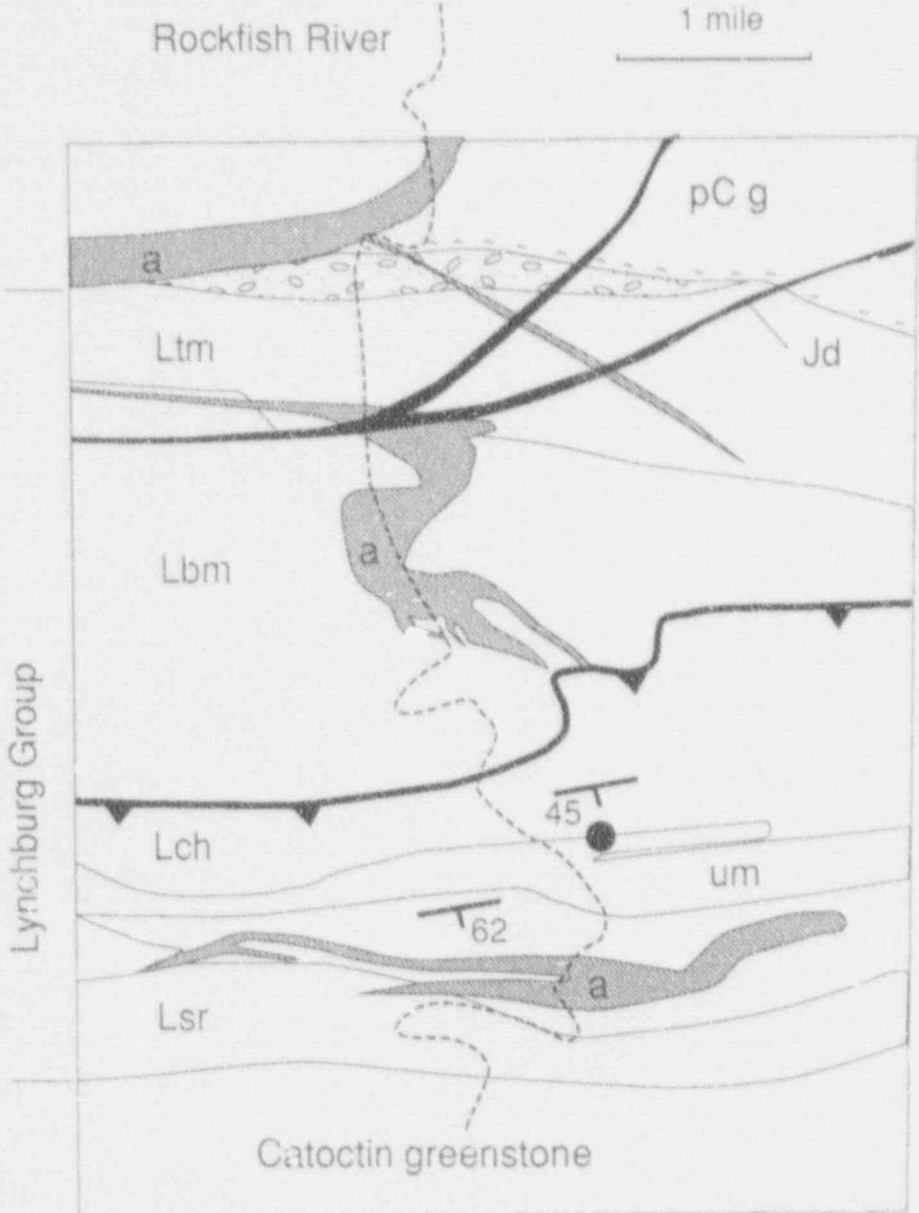

Figure 6. Ultramafic sill contact at Schuyler, VA. Lynchburg formations: Lsr = Swift Run; Lch = Charlottesville; Lbm = Ball Mountain; Ltm = Thorofare Mountain; a = amphibolite; um = ultramafic complex. From Wehr, 1983.

Lynchburg and Swift Run formations along the east side of the Blue Ridge in Virginia. They comprise the Albemarle - Nelson soapstone belt (Burfoot, 1930). The ultramafic association includes, in decreasing order of abundance, amphibolite-chlorite schist, serpentinite, soapstone, and altered peridotite (Burfoot, 1930; Hess, 1933; Brown, 1958; Nelson, 1962; Misra and Keller, 1978). Hess (1933) concluded that the parent material was peridotite and feldspathic peridotite (picrite). An intrusive contact is visible between thin-bedded Charlottesville Formation and the Schuyler ultramafic sill at Schuyler. Hess (1933) found the following sequence of rock types in the Schuyler body: 1) at the base, ultramafic rock, talc-chlorite-actinolite-calcite; 2) in the middle, gabbroic rock with hornblende and actinolite assemblages; and 3) silicic rocks with quartz-albite-microcline-chlorite-hornblende assemblages. This suggested, to Hess, that the sill had differentiated in place. Brown (1958) thought that the ultramafics might be extrusive rocks, as this would explain the localization parallel to bedding, and association with the Catoctin lavas. So far we have not seen evidence of this.

In the Culpeper to Schuyler region (Figure 7) the ultramafic rocks are confined to the upper part of the Lynchburg Group and Swift Run Formation. None are found above the base of the overlying Catoctin Formation or within the still younger Evington Group. In the Lynchburg area, 100 km south of Charlottesville (Plate 1), ultramafic rocks occur throughout all but the lowest part of the Lynchburg (Brown, 1958) where the contact relations reveal them to be intrusive sills (Ping Wang 1988, personal communication).

Differentiation relations between hornblende gabbro and ultramafic rocks suggested for the Schuyler sill (Hess, 1933) and confinement of hornblendite dikes and sills as well as the ultramafics to stratigraphic levels no higher than the Catoctin Formation, suggest that the hornblende - actinolite gabbro and ultramafics may ifferentiates of a common subcrustal magma (see also Bloomer and Werner, 1955) of ynchburg and/or Swift Run age, that is, Late Precambrian-Early Cambrian, 650 - 570

Whatever the future may provide about the details of the ultramafic - mafic assemblages described above, they appear to be part of the late Iapetan rift sequence which extends upward and includes the Slippery Creek basalt of the Evington Group (see below) as well as basalts in the Unicoi Formation in the western Blue Ridge. They are not part of an ophiolitic melange as implied by Hatcher (1987), Hatcher and others(1989), Rankin and others (1989), Horton and others (1989), and Keppie and Dallmeyer(1989).

Correlations with the Valley and Ridge. The Catoctin and Swift Run formations form a common stratigraphic datum on both sides of the Blue Ridge in central and southern Virginia (Plate 1). This has long been recognized as an important starting point for correlation of the strata across the Blue Ridge (Bloomer and Werner, 1955, Brown, 1970, Patterson, 1987).

Glover and Costain (1984) and Wehr and Glover (1985) have shown that the Blue Ridge province is the thrust-decapitated crest of the early Paleozoic hinge zone of Laurentia. Late-Early Cambrian through Early Ordovician strata west of the Blue Ridge crest belong to the shallow water drift sequence, and correlative rocks east of the Blue Ridge are deep water distal shelf and slope deposits (Brown, 1970; Wehr and Glover, 1985; Patterson, 1987; Glover, 1989).

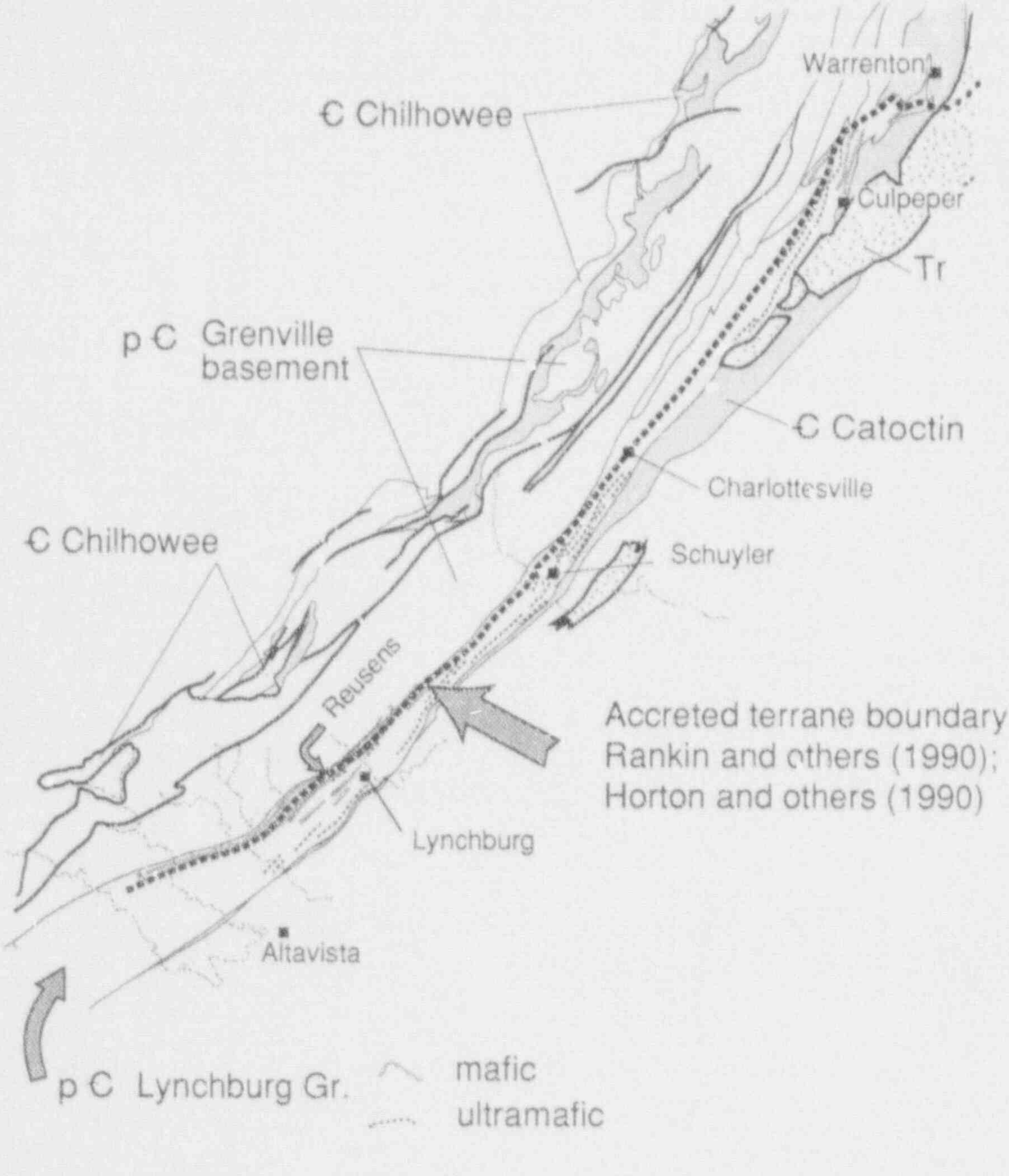

Figure 7. Virginia Blue Ridge, showing ...rane boundary of Rankin (1990) and Horton and others (1989).

On the west side of the Blue Ridge the Catoctin is overlain by the Chilhowee Group and includes, from oldest to youngest, the Unicoi / Weverton, Hampton / Harpers and Erwin / Antietam formations. The second named in each couplet is the commonly used term for equivalent formations north of central Virginia.

In southern Virginia the Unicoi Formation, at the base of the Chilhowee, is a sequence of dominantly non-marine feldspathic sandstones, conglomerates and basalts formed during the later part of the rift stage that led to the development of Iapetus (Simpson, 1987; Simpson and Eriksson (in press). As noted previously, the Unicoi south of the Catoctin pinch out on the west side of the Blue Ridge, probably includes rocks equivalent to the Swift Run which occurs below the Catoctin north of that pinch out.

The Unicoi is overlain by the dominantly progradational (Simpson, 1987), and compositionally more mature, quartz arenitic Hampton and Erwin Formations of the middle and upper Chilhowee. Simpson (1987) and Simpson and Eriksson (in press) place the rift to drift transition at the top of the Unicoi. Overlying the Chilhowee is the Shady Dolomite which records continued progradation of the drift sequence culminating in development of a rimmed shelf (Read, in press).

This information can be used to provide a [illegible] roved correlation with the deeper water facies of the Evington Group, which lies above the Catoctin on the east side of the Blue Ridge (Figures 5,Plate 1,). In the [illegible]aterson (1987) preferred model of this correlation the rift-to-drift transition is placed just above the Slippery Creek Greenstone, the youngest lava in the sequence. Thus, the Swift Run (?), Catoctin, Candler, Mount Athos, and Slippery Creek should all be approximately correlative with the Cambrian and Cambrian (?) Unicoi Formation. The Joshua Schist should be correlative with the Hampton, Erwin, and possibly part of the Slippery Creek, and the Arch Marble with the Shady Dolomite. The Chilhowee and Shady are Early Cambrian (excepting possibly the lower Unicoi), therefore the Evington Group should be entirely Early Cambrian also. The upper part of the Evington Group is truncated by faults and no younger Laurentian strata are known in the western Piedmont.

Cambrian through Early Ordovician Drift (Passive Margin) Stage, Laurentian Continent

Subsidence of the platform during drift resulted in the accumulation of about 3.5 km of clastics and carbonates now exposed in the Valley and Ridge and western Blue Ridge Provinces of Virginia. Retrogression, and Rome trough rifting in the continent, dominantly west of the Valley and Ridge, resulted in about 0.5 km of Rome Shale ponded west of the Shady Dolomite shelf rim (Read, 1989). This was followed by deposition of about 1.6 km [illegible] shallow water carbonates and shales (Elbrook Dolomite/Conasa[illegible]ga Shale Group [illegible]d overlying Knox Dolomite Group) capped in most places by a regional unconformity of early Middle Ordovician age.

Outer shelf and slope deposits younger than the Arch Marble, Early Cambrian Shady Dolomite equivalent, are unknown east of the Blue Ridge in Virginia.

Depositional model for Late Precambrian to Early Cambrian Laurentian Continental Margin

690-650 Ma - Late Precambrian Early Rift Stage: rifting of Grenville basement; emplacement of Crossnore Volcanic Plutonic suite; uplift and erosion, local grabens preserved Crossnore volcanic rocks and associated non-marine sediments, Crossnore plutons exposed in basement

650-570 Ma - Late Precambrian to Early Cambrian Late Rift Stage: glaciation at about 650 Ma; continued rifting; development of hinge zone west of Lynchburg near center of Blue Ridge basement; deposition of Lynchburg periglacial braided alluvial fan overlain by Lynchburg retrogradational slope and deep basinal marine fan sequence; sediment source dominantly from Grenville basement to the west; emplacement of: mafic dikes, sills, and lavas(?), and ultramafic dikes and sills (by injection of olivine-rich crystal mush derived from fractional crystallization in a previous chamber, or other dynamic crystal accumulation process?) in the Lynchburg and Swift Run; emplacement of mafic dikes and lavas in Catoctin; deposition of Unicoi non-marine sediments and basalt lavas west of hinge zone on west flank of Blue Ridge; deposition of marine slope and basinal Candler and Mount Athos Formations of Evington Group east of hinge zone (east flank of Blue Ridge); marine eruption and deposition of Slippery Creek basalt on Mount Athos quartz arenite; formation of oceanic crust east of Slippery Creek; end of rift stage

570 - ca. 550, Early Cambrian Drift Stage: deposition of Hampton/Harpers shale and quartz arenite west of hinge zone on west flank of present Blue Ridge; coeval deposition of Joshua turbidites east of hinge on east flank of present Blue Ridge; followed by buildup of Shady rimmed reef margin (Read, 1989) on hinge and coeval deposition of deep water limestone (Arch marble) east of hinge zone

ca. 550 - ca. 540, Late Early Cambrian Overlapping Rift of Laurentian Eastern Interior and Continued Laurentian Drift: rifting of Laurentia in eastern interior, regression of sea and deposition of Rome shale behind Shady reef rim

ca. 540 - 490, Late Early Cambrian - Early Ordovician Drift: buildup of largely carbonate bank west of hinge zone into Early Ordovician time

Alternate tectonic models that have been proposed

Hatcher (1987, Figures 2, 3; Hatcher and others, 1989) proposed an extension of the Hayesville - Fries fault, which he considers to be the Penobscot-Taconic suture, into the Virginia Blue Ridge and Piedmont. By this reconstruction the Lynchburg Group is shown as allochthonous upon the Grenville basement and Catoctin Formation (personal communication May, 1988), and the surface of thrusting is the Penobscot-Taconic suture. Hatcher's interpretation is contrary to the geologic relations described herein. The basal conglomerate of the Lynchburg Group in northern Virginia (Figure 5) contains cobbles of granite derived from immediately underlying Crossnore granites in the Grenville basement, therefore no large distance thrusting is indicated. The Catoctin overlies the Lynchburg and is continuous with the Valley and Ridge sequence on the west side of the Blue Ridge; here also no large scale thrust fault is indicated. As noted previously, the Lynchburg on the east side of the Blue Ridge is in depositional contact, not fault contact, with the overlying Catoctin.

An interpretation similar to that of Hatcher (1987; Hatcher and others, 1989) has also been made by Horton and others (1987) and by Horton and others (1989) who have named the Lynchburg-Ashe-Tallulah Falls-Wedowee strata along the east flank of the central and southern Appalachian Blue Ridge the Jefferson terrane. A similar interpretation is also shown by Keppie and Dallmeyer (1989). The same objections apply to all of these reconstructions.

Late Precambrian/Early Ordovician History of the Exotic Carolina Terrane

Chopawamsic Formation Type Area. Volcanic rocks known as the Chopawamsic Formation crop out in a NNE-striking belt in the central Piedmont of Virginia (Plate 1). The formation was named by Southwick and others (1971) for rocks cropping out in Stafford and Prince William counties in northern Virginia. The type section occurs along Chopawamsic Creek on the Quantico Marine Base in the Joplin, Virginia 7.5 minute Quadrangle. The formation in that area consists of, "... (1) metamorphosed medium- to thick-bedded mafic to intermediate volcanic rocks derived from andesitic to basaltic flows, coarse breccias, and finer tuffaceous clastic rocks; (2) metamorphosed medium- to thick-bedded felsic volcanic rocks derived from flows and associated volcaniclastic accumulations; and (3) metamorphosed thin- to medium-bedded volcaniclastic rocks of felsic to mafic composition, locally containing beds of non volcanic quartzose metagraywacke, green to gray phyllite, and felsic to mafic flows. Units 1 and 2 grade vertically and laterally into unit 3 and appear to be tongues or lenses within a complex volcanic-sedimentary pile" according to Southwick and others (1971). The thickness of the Chopawamsic in the type area is 2-3 km. The lower contact was not seen, but its position was inferred within a meter-wide covered interval and the contact was judged by them to be sharp in one area and interleaved in another. Relations between correlative rocks in Maryland (Crowley, 1976) suggests to me that if the diamictite at Chopawamsic is the Maryland diamictite, the contact may be a fault in northern Virginia, or, that metavolcanic rocks of northern Virginia were misidentified as "Wissahickon" diamictite. The upper contact was found by Southwick and others (1971) to be gradational by interleaving with the overlying Quantico Slate. In contrast, Pavlides (1976) working to the south, near Fredericksburg, found an unconformable relation between the Quantico (Arvonia equivalent) and the underlying Chopawamsic; this relation is in accord with the relations seen along our traverse (Brown, 1969).

Isotopic ages (zircon) indicate that the Chopawamsic is about 550 Ma, or Early Cambrian (Pavlides, 1981, p. A6).

Chopawamsic Formation along the James River in Virginia. Smith,and others (1964) dropped the earlier name "Peters Creek Quartzite" in favor of "metamorphosed volcanic and sedimentary rock unit" of the Evington Group. They subdivided the Chopawamsic into three map units:

1) a predominant phase, very fine - grained plagioclase - quartz gneiss and sericite phyllite, either may have local abundance of quartz and feldspar crystals; felsite porphyry, and local quartzite and phyllite.

2) a mafic phase, amphibole schist and gneiss (locally with amygdales), biotite-chlorite schist, and plagioclase-chlorite-epidote rock, and,

3) grossly interlayered felsic and mafic rocks.

Brown (1969), mapped this formation to the west of the Arvonia syncline (Plate 1) as "metavolcanic rocks" of the Evington Group. He found greenstones derived from mafic volcanics, and porphyritic rocks of dacitic composition interlayered with feldspathic metasedimentary rocks. Lying to the east of the Arvonia syncline is the Hatcher complex which Brown (1969) named for plutonic granite, granodiorite, and quartz diorite injected into amphibolite and mica gneiss. At the southern end of the Hatcher he recognized "rocks of uncertain age" some of which resembled the Arvonia and underlying metavolcanic rocks west of the Arvonia syncline. Subsequent authors (for example, Conley, 1978; Pavlides, 1981) have considered the amphibolite of the Hatcher complex and the "rocks of uncertain age" to be higher grade parts of the Chopawamsic Formation.

Tectonic interpretations. Pavlides (1981), working to the south of the type area, near Fredericksburg (Plate 1), concluded from trace element geochemistry that the Chopawamsic on the west side of the Quantico syncline was a tholeiitic island arc suite with associated calcalkaline rocks. The more mafic Ta River suite, on the east side of the Quantico syncline, has affinities (seven analyses) with oceanic basalt. The Ta and Chopawamsic were never seen in contact.

Pavlides (1981) suggested that the Ta River was a more oceanward facies of the Chopawamsic volcanics and that the subduction zone therefore dipped westward under the Chopawamsic arc. A marginal, back-arc, basin was thought to separate the Chopawamsic arc from the Laurentian continent. The small number of Ta River analyses, scatter outside of the defining trace element fields, and lack of contact relationships between the Chopawamsic and Ta River weaken the paleotectonic conclusions. Pavlides also recognized that he couldn't confirm that the Chopawamsic and Ta River were coeval, and would concede that the Ta River might be thrust over the Chopawamsic. Reconstruction of the development of the ancient Laurentian margin, as outlined in this paper, indicates that rifting ended and drift began about 570 Ma, Early Cambrian, probably before much of the ca. 550 Ma Chopawamsic volcanic rocks were erupted. If the Chopawamsic arc developed along the eastern edge of Laurentia, as proposed by Pavlides, much pyroclastic material would have fallen into the Laurentian Early Cambrian drift sequence, but it is not there. Rifting to form a backarc basin occurs in modern arcs along the axis of the arc so that the basin evolves with a dead arc behind it and an active arc adjacent to the subducting margin. Closure of the backarc basin would require a jump in subduction position and a reversal of polarity according to the Pavlides model. In this case the resulting structural sequence from west to east would be: 1) an early, pre rift, half arc of calcalkaline rocks that interfingered westward with Late Precambrian and Early Cambrian upper Lynchburg, Swift Run, Catoctin, Evington Group, Unicoi, Hampton, Erwin, and Shady strata. 2) an accretionary melange (Shores and Hardware) of the backarc basin sequence with calcalkaline pyroclastic and epiclastic rocks from the margins of the basin floor, and basin floor basalts. and 3) calcalkaline volcanic rocks of the outer proposed active arc (Chopawamsic) thrust over the accretionary melange. The foregoing is not in accord with the geologic framework developed in this paper. Furthermore if the Chopawamsic arc developed adjacent to the Laurentian continent in the Early Cambrian some evidence of carbonate reefs might be expected from the vicinity of volcanic islands in a subequatorial sea, but carbonate detritus does not seem to occur in these rocks. If on the other hand as proposed below, the Chopawamsic volcanics are part of Carolinia (Figure 2), they would have been at high latitudes during this time and carbonate would be rare or absent.

In 1976 W.R. Brown recognized the tectonic melange nature of the rocks lying between the Chopawamsic volcanics and "Evington Group" rocks to the west. Thus the Chopawamsic volcanics and the melange could no longer be considered part of the Evington Group and a collision zone or suture was implied. Bland and Blackburn (1980) characterized the Chopawamsic volcanics as part of the younger Carolina slate belt of Glover (1974), and determined on the basis of trace element studies that they were dominated by low-K tholeiites. In this sense they differed from the older Carolina slate belt (Glover, 1974), which they characterized as a calc-alkaline sequence of volcanics. Building on the models of Rodgers (1972) and Glover and Sinha (1973), Bland and Blackburn (1980) suggested two possible models, each identifying the melange as as an ocean-floor off-scraping, and the locus of eastward subduction of oceanic crust below the Chopawamsic volcanics, which bordered a marginal basin off of Laurentia. In one variant of the models the Chopawamsic is a separate island arc that subsequently collided, by eastward subduction, with the Carolina slate belt (older slate belt). A problem with this model is that the younger slate belt (including the Chopawamsic Formation) was deposited unconformably upon the older slate belt (Glover, 1974; Harris and Glover 1985, 1988; see also below) and not thrust upon it as this model requires. In the second variant of the model the Chopawamsic represents a later episode of subduction and volcanism superimposed on the older slate belt. This model is similar to the one developed in this paper.

Pre- and Post - Virgilina Deformation, Carolina Slate Belt Sequences. In 1973 Glover and Sinha discovered an orogenic event, the Virgilina deformation, in the Carolina slate belt sequence at Roxboro, northern North Carolina. At that location an older sequence of volcanics and epiclastic rocks was folded and faulted at about 600 Ma., and was subsequently intruded by the Roxboro Metagranite at 575 ± 20 Ma. Glover (1974) speculated that the deformation would have produced an unconformity and that the [illegible] gently folded sequence near Asheboro in central North Carolina had probably been deposited above this unconformity. The concept of pre- and post-Virgilina sequences in the slate belt was reinforced by Briggs and others (1978) with the determination that the Roxboro Metagranite was a very shallowly emplaced pluton that was probably an eruptive source for part of the younger volcanic sequence. This concept was challenged by Wright and Seiders (1980), in a study of the central North Carolina sequence, who proposed three possibilities: 1) The stratigraphic sequences of the two areas (Roxboro and Albemarle NC) are partly correlative. The Virgilina deformation was synchronous with deposition of the upper part of the central North Carolina sequence, but the deformation did not extend into the central North Carolina area. 2) The stratigraphic sequences of the two areas are correlative, and the Virgilina deformation was younger than the central North Carolina sequence but was weak or absent in that area. 3) The central North Carolina sequence is entirely younger than the Virgilina deformation, and the volcanic rocks may represent an extrusive phase of plutonism of the Roxboro-Durham area (the Glover speculation). Wright and Seiders favored possibility # 1 as the most likely relationship, this was largely based on their belief that the well bedded Tillery Formation of the Albemarle area was the same as the well bedded Aaron Formation of the Roxboro area. In 1988 Harris and Glover presented evidence for the Virgilina unconformity in the Albemarle area and showed that the Aaron was part of the older sequence unconformably below the younger Tillery-bearing sequence of the slate belt (Figure 8). They also suggested that the *intra* arc basin sequence (largely epiclastic deep water turbidites of the Aaron Formation), and the strongly bimodal nature of most of the younger volcanism, may be analogous to the proto-Gulf of California or to transcurrent pull-apart basins. Perhaps then the Virgilina deformation represents an oblique collisional deformation (transpressional and transtensional) that occurred on Carolinia prior to its collision with Laurentia.

Chopawamsic, James Run, Carolina Slate Belt, Kings Mountain, Charlotte Belt, Raleigh Belt, and Eastern Slate Belt: All Parts of the Carolina Terrane in the Southeastern U.S. Piedmont. In 1972 Higgins proposed the existence of a long belt of metavolcanic rocks, the "Atlantic seaboard volcanic province", that extended in the Piedmont from Georgia to New York during the Late Precambrian to Early Ordovician. Subsequent literature largely ignored this in favor of multiple arcs and complex collision scenarios. The crustal profile and field data presented here indicates that Higgins was essentially correct in relating all of the volcanics to a single terrane.

The James Run/Chopawamsic volcanic rocks occur in the eastern Piedmont of Delaware and Maryland and extend into the north central Piedmont of Virginia (Figures 2, Plate 1). They were overthrust from the east by the Laurentian Goochland basement nappe (Figures 2, Plate 1) during a late Paleozoic Alleghanian dextral transpression event (Glover and Gates, 1987). By reconnaissance and local detailed mapping they have been traced into southern Virginia where they comprise part of the Charlotte belt. Volcanic rocks of the Eastern slate belt and Raleigh belt are also clearly part of the Carolina slate belt sequence (Farrar, 1985). Detailed mapping across the Charlotte belt in southern Virginia (Figure 1, Plate 1) has not revealed any suture within the sequence, a sequence which has been recognized for more than a decade as higher grade volcanic rocks equiva-

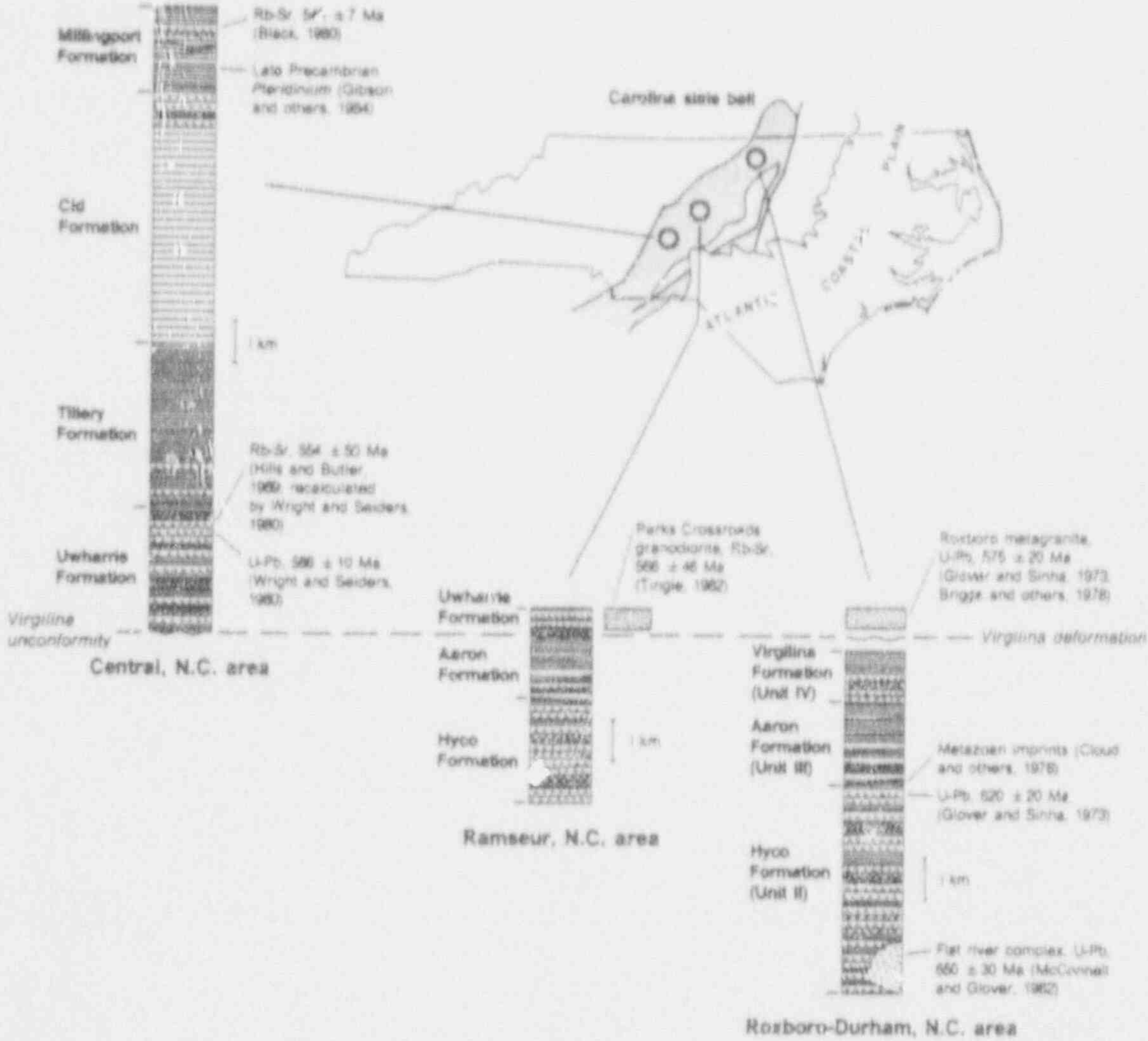

Figure 8. Correlation of Carolina slate belt stratigraphy in North Carolina. From Harris and Glover, 1988. The Chopawamsic Formation in northern Virginia is correlative with the post Virgilina sequence of North carolina.

lent to the Carolina slate belt (Glover and Sinha, 1973). The Cambrian age and chemistry (Bland and Blackburn, 1980) of the James Run/Chopawamsic volcanic rocks is similar to the younger Carolina slate belt sequence (Cambrian-Latest Precambrian) shown in Figure 8, which unconformably overlies an older Late Precambrian volcanic slate belt sequence deformed during the ca. 600 Ma Virgilina deformation (Glover, 1974; Briggs, Gilbert and Glover, 1978; Bland and Blackburn, 1980; Harris and Glover, 1985, 1988).

Kings Mountain belt rocks of the Carolinas (Figures 1, 2) bear some similarities to the Chopawamsic and Arvonia sequences along strike in central Virginia. Both sequences contain volcanic rocks, ultramafic rocks, quartz sericite schist, quartzite, calc-silicates (limy mudstones), marble, and graphite schist (Gates, 1981; Horton and others, 1981; Horton, 1983). If the proposed correlation turns out to be correct, the Kings Mountain sequence would range into the Cambrian and possibly have some post-suture Ordovician Arvonia infolded/faulted into it. The Kings Mountain and Chopawamsic rocks might be part of the Cambrian sequence found in the slate belt in South Carolina, which also grades upward into a less volcanic and more epiclastic sequence containing quartz rich sandstones. Similar quartz sandstones also occur in the Eastern slate belt of North Carolina. The Kings Mountain and Chopawamsic both occur along the Taconic suture and were probably dropped downward along early Mesozoic reactivation of the suture as indicated by the occurrence of Triassic basins and late brittle fractures having zeolitic assemblages along it.

Therefore, all of the volcanic sequences in the preceding can be considered to comprise segments of a single terrane, *Carolinia* (use modified from Secor and others, 1983), that collided with Laurentia during the Taconic (Figure 2). The youngest stratified rocks known in Carolinia are the Asbill Pond and Richtex formations in South Carolina which are Middle Cambrian (about 530 Ma.) Secor and others (in press). Both of these formations contain pyroclastic rocks. Volcanism probably persisted into the Middle Ordovician because Ordovician plutons are known in the Piedmont and bentonites occur in the Taconic clastic wedge in the Valley and Ridge.

Events Leading up to Middle and Late Ordovician Collision (Taconic Orogeny) between Carolinia and Laurentia.

Shores Melange. Brown (1976; Brown and Pavlides, 1981) first recognized the significance of the structures in the outcrops along the James River at Shores, Virginia (Plate 1), which Brown named the Shores complex melange. Glover and others (1982) gave seismic reflection evidence for a fault boundary between the Chopawamsic Formation and the Shores melange. Evans (1984) gave evidence for a fault contact with the Hardware sequence on the west. Evans (1984) described the Shores as a polydeformed amalgam of quartzo-feldspathic epidote-chlorite gneiss, epidote-chlorite migmatitic gneiss, and hornblende-epidote-albite schist (greenstone). The rock types are heterogeneously mixed on a scale of meters to tens of meters. In many outcrops greenstone occurs in blocks as much as several meters across, enclosed in epidote-chlorite gneiss or migmatitic gneiss. According to Evans (1984), quartzo-feldspathic epidote-chlorite gneiss is characterized by metamorphic segregation layering defined by quartz-albite and epidote-chlorite-magnetite-titanite layers.

Migmatitic gneisses are gradational from non-migmatitic gneisses and contain quartz-plagioclase-muscovite lenses which appear to have crystallized from a melt.

Greenstones are infolded with the surrounding gneisses and are locally intruded by tonalite veins. Some greenstones were coarse-grained gabbros others were fine-grained and may have been extrusive rocks.

Both the quartzo-feldspathic gneisses and the migmatitic gneisses may have been largely of graywacke protolith.

Petrologic and igneous fabric analysis by Evans (1984) indicates that the Shores reached medium- to high-pressure epidote amphibolite conditions. Temperatures on the order of 630°C and pressures of at least 6-7 kb were inferred under which metamorphism of the non-migmatitic gneisses and incipient melting of the migmatitic gneisses occurred.

Lower greenschist overprinting (quartz + albite + epidote + chlorite + muscovite ± magnetite) involved hydrothermal alteration and oxidation of the earlier metamorphic assemblages.

Brown (1986) described the Shores complex at Shores in detail and gave its regional setting as a major zone of thrusting and obduction.

The metamorphic and deformational history of the melange is distinctly more complex in temperature, pressure and structural development than that of the Hardware terrane, to the west, upon which the Shores is thrust and this has importance in determining its early history.

Origin of the Shores Melange. Previous studies (Bland, 1978; Bland and Blackburn, 1980) have shown that the greenstone blocks in the Shores melange have the geochemical signature of ocean floor basalts. The work of Evans indicates that the Shores was metamorphosed under conditions differer an the rocks upon which it was overthrust to the west. Higher grade metamorphism f ved by lower grade metamorphism is a common sequence in melanges of accretionary prisms. Recent work by Cloos (1982, 1984) and Cloos and Shreve (1986) suggests how this sequence of metamorphism and deformation may come to be. Cloos and Shreve discuss five possible types of flow patterns in melanges. Their types D (composed of slope cover, offscraped sediment, offscraped melange, and underplated melange) and E (composed of slope cover, offscraped melange, and underplated melange) seem to fit the sequence in the Shores best, because these are the only ones in which once more deeply buried material may return toward the surface during accretion. In these models a metamorphic aureole in melange is formed at the base of the hot overriding plate. Return circulation may develop in the underlying cooler and fluid-rich subducting sediments which plucks blocks of the metamorphic aureole and carries them back toward the surface. While this is an attractive hypothesis it should be kept in mind that the second (and later?) overprints on the melange may have occurred during regional metamorphisms related to tectonic burial.

Hardware Metagraywacke. This unit was named by Evans (1984) for a graded metagraywacke sequence that lies between a fault bounding the Shores Melange and the Mountain Run fault (a name which has priority over the Buck Island fault zone of Evans) bounding the Evington Group (Plate 1). Within the Mountain Run fault zone Evans found very thin bedded, fine grained graywackes thought to be distal turbidites related to the Hardware Metagraywacke. These cover a narrow area and are not shown on Plate 1.

Hardware metagraywackes (Evans, 1984) are quartzose chlorite schists and phyllites with laminations 1 mm to 1 cm in thickness. Grain size is fine- to medium-sand with local pebbly lenses. Local allochthonous blocks of metamorphosed mafic igneous rocks occur in the Hardware. Detrital components of the metagraywacke include subequal amounts of quartz and plagioclase, tourmaline, epidote, magnetite, titanite, and rare clasts of metamorphic muscovite (after detrital K-feldspar?). Included lithic fragments are; dacitic tuff, gabbro, and granite (quartz with zircon and biotite inclusions and remnant perthitic feldspar). These detrital components suggested to Evans (1984) a source to the east in the Chopawamsic volcanics, Shores Melange and Goochland nappes. Subsequently Glover and others (1987) have shown that the Goochland was emplaced during the Late Paleozoic, thus it probably was not the source of the granitic fragments. Timing relations suggest that Carolinia is the probable source. However, the Chopawamsic, Shores and deep water Laurentian rift/drift sediments remain plausible sources for the Hardware.

Age and correlation of the Shores Melange and Hardware Metagraywacke. The Shores and Hardware rocks are undoubtedly part of the complex melange sequence described by Drake and Morgan (1981), Drake and Lyttle (1981), Drake (1985), and Drake, (1987) along strike in northern Virginia (Figure 2, Plate 1). This sequence consists of three tectonic motifs (Drake, 1987), each motif including an allochthon of overlying deep water turbidite sedimentary rocks underlain by a precursory melange. The higher two motifs have ultramafic and mafic blocks in melange; the lowest has only basaltic blocks. All three allochthons are overlain unconformably by a turbidite sequence (Popes Head Formation) containing some units that may be mafic and felsic ashfall tuffs. The three motifs show more deformations and an additional metamorphism not experienced by the Popes Head.

All motifs, as well as the Popes Head, are intruded by the synkinematic(?) Occoquan Granite. Seiders and others (1975) dated the Occoquan by the zircon U/Pb method at about 560 Ma. Mose and Nagel (1982) dated the Occoquan by the Rb/Sr whole rock method at 494 ± 14 Ma. Because the Occoquan was emplaced in hot rocks undergoing metamorphism, the Rb/Sr age could be younger than the emplacement age as a result of slow cooling delaying closure of the isotopic system. Abundant experience in the Appalachians suggests that about 25 m.y. should be added to the cooling age to approximate the emplacement age in such cases. Therefore the Occoquan was probably emplaced about 525 Ma, possibly at 560 Ma assuming no inheritance of older lead in the zircons that were dated.

Additional age constraint on the Hardware comes from the identification of probable Chopawamsic volcanic fragments erosionally introduced in to it (Evans, 1984). The Chopawamsic has been dated at about 550 Ma (Pavlides, 1981). Thus at least part of the Hardware must be younger than 550 Ma but probably older than 525 Ma. Because the Shores was also a source for the Hardware it must be somewhat older than the part of the Hardware for which it was a source. It seems probable then that much of the Shores/ Hardware is about 550-525 m.y. old (Middle to Early Cambrian) and that parts could be older.

Regional Tectonic Interpretation. In northern Virginia Drake (1987) considers the three motifs (motif = precursor melange overlain by deep water turbidites), previously mentioned, to constitute three terranes amalgamated by suturing into one and overlain by the Popes Head Formation. Subsequently, during the Taconic, he suggests that they collided with Laurentia. The pre-Popes Head, and the post-Popes Head/pre-Occoquan deformations Drake considered to be records of the Cadomian (western Europe) and Penobscot (New England Appalachians) orogenies respectively.

Studies along the James River suggest rather, that the three motifs of Drake are segments of an accretionary melange brought ashore during the Taconic orogeny. Where it has been possible to identify source in these deposits it seems to be from the east. The Popes Head may be a forearc basin deposit containing some pyroclastics from the magmatic arc. If this hypothesis is true, the deformations Drake correlated with widespread orogenies around the Atlantic may be incorrect. Rather, it is suggested that the pre- and post-Popes Head deformations represent different stages in the deformation of a complex accretionary wedge that developed over a long interval of time, beginning offshore from Laurentia.

Plutons in the Melange Units. Several gabbroic and granitic plutons occur in the melange (Plate 1), and some of the granitic plutons have ages of ca. 500 Ma. This dates them as possibly having been generated over oceanic crust before the Taconic collision with Laurentia . Much work needs to be done on petrogenesis of these intrusives. There are similar occurrences in Kodiak Island, Alaska, where granitoid and gabbroic plutons were apparently generated in the accretionary wedge above oceanic crust. Perhaps they can be explained as a result of the subduction of an active spreading ridge which could emplace basaltic magmas into the melange and also create secondary granitic melts from the melange itself.

Taconic Orogeny.

The Taconic orogeny was the earliest to affect the Laurentian margin, it must therefore be related to the first suture found outboard of Laurentia. The previous discussion supports the Shores Melange as marking the suture and the Chopawamsic volcanic rocks of the exotic Carolina terrane as remnants of the colliding continent. Collision between Carolinia and Laurentia occurred during Late Cambrian through Late Ordovician time. The initial collision decapitated a slice from the Laurentian hinge zone, and this slice was the ancestral Blue Ridge (Glover and others, 1983; Glover and Costain, 1984; Wehr and Glover, 1985). Erosion breached the ancestral Blue Ridge down to Grenville basement and fragments of the western platformal rift and drift stratigraphy down to the basement are preserved in the Late Ordovician Fincastle Conglomerate of the Taconic foreland basin near Roanoke Virginia (Karpa, 1974). Fragments of gneissic lower Chilhowee (Unicoi?) in the Fincastle indicate that Taconic metamorphism in the hinge and continental slope sequence was already well advanced by Late Ordovician, Caradocian, or about 450 Ma. Glover and others (1983) summarized the evidence for Taconic metamorphism which ceased by cooling during thrust-driven uplift over most of the Piedmont and Blue Ridge at about 480 Ma. or Early Ordovician. Further deformation and filling of the foreland basin continued until about 440 Ma or Late Ordovician time in the central and southern Appalachians.

Other Tectonic Interpretations

Hatcher (1987, Hatcher, 1989) includes the Lynchburg, Swift Run, Hardware Shores and Chopawamsic strata in his Piedmont terrane. In this paper however, it has been shown that: the Lynchburg, Swift Run and Catoctin are Laurentian rift stage strata deposited on continental crust; the Hardware and Shores are tectonic melange and deep water sediments of oceanic and continental derivation; and the Chopawamsic volcanic rocks are part of Carolinia, a terrane that collided with Laurentia during the Taconic. Therefore it seems unlikely that Hatcher's Piedmont terrane, as presently constituted, is a valid tectonic unit. Similar criticism applies to the models of Rankin and others, 1989, and to Horton and others, 1989.

The model presented herein is also strongly at varian... with Hatcher's (1987) concept of the timing of regional metamorphism and collision in the central and southern Appalachians. Much of the problem, as shown in this paper, lies in our differing views on the number of terranes and their ages of collision.

Latest Ordovician, Silurian and Early Devonian Drift.

During and following erosional reduction of the Taconic Mountain system, Carolinia and Laurentia (Figure 2) apparently drifted together as a single continent for about 30 m.y. During this time subsidence and perhaps transform motion parallel to the collisional axis allowed successor basins to accumulate quartz arenite and carbonate on the platform (present Valley and Ridge Province). In the Piedmont, over the eroded roots of the Taconic Mountains, the Arvonia Formation and correlatives furnish a record of Paleozoic sedimentation following the Taconic orogeny.

Arvonia Formation (Watson and Powell, 1911). This formation (Plate 1) is a laminated to thin bedded quartz-muscovite-graphite schist or phyllite with lesser amounts of biotite, chlorite, magnetite, plagioclase, pyrite, carbonate minerals, and local grains of tourmaline and zircon; garnet occurs in the eastern exposures (Smith and others, 1964; Brown, 1969, 1970). The base of the formation is unconformable upon the Rb/Sr 454 ± 9 Ma. (Mose and Nagel, 1984) Columbia Granite and upon the U/Pb ca. 550 Ma. Chopawamsic volcanics. The top of the formation is the present erosion surface.

Quartz arenite occurs locally at the base, especially where it is in unconformable contact with the Columbia Granite (Taber, 1913).

Along the James River the Bremo quartz arenite and quartz pebble arenite occurs in the middle and lower part of the formation (Brown, 1969, 1970).

The Buffards Conglomerate Member, placed at the top of the formation by Brown (1969, 1970), now appears to be a localized unit near the base of the Arvonia according to new mapping by Evans and Marr (1988). Buffards crops out about 10 miles SSW of Arvonia contain massive conglomerate consisting of well rounded pebbles and cobbles of quartzite, mafic and felsic volcanic pebbles and cobbles in a quartzo-feldspathic sandstone matrix. The conglomerate is interleaved with graded graywacke and dark phyllite.

Age and correlation of the Arvonia. Tillman (1970) studied the trilobites that occur in the Arvonia and summarized the age as Middle or Late Ordovician. Earlier identifications of all of the Arvonia fossils available in the late 1940's yielded a probable Maysville stage of the Late Ordovician (Stose and Stose, 1948). As noted above, the formation rests unconformably upon the 454 ± 9 Ma. Columbia Granite (Mose and Nagel, 1984). Considering that the Columbia was intruded into the low grade Chopawamsic volcanic rocks during or late in their Taconic metamorphism it is unlikely that the granite cooled below the retention temperature for the Rb/Sr isotopic system immediately. Thus, the emplacement age is probably 10-20 m.y. older, or Middle Ordovician. These rough brackets on the age of the Arvonia and Columbia Granite imply that the unconformity itself is probably Middle but not Late Ordovician (Taconic) in age (see discussion of the age of the Quantico and Arvonia below).

The Arvonia has been correlated with the Quantico Formation, a similar black phyllite or schist, which crops out in a syncline to the east of Arvonia and extends from the James River to northern Virginia (Pavlides, 1980, and references therein). Pavlides and others (1980) clarified the age of the Quantico in northern Virginia, assigning a Late Ordovician or Silurian(?) age to it. Therefore, the age of the Quantico and Arvonia might be in part Middle Ordovician, *is* in part Late Ordovician or Silurian. The age of the basal unconformity then is probably Middle to Late Ordovician.

Tectonic Interpretation. Because Carolinia and Laurentia were sutured by the Taconic orogeny, the Arvonia and Quantico represent an overstepping sedimentary unit that was probably deposited across the suture. The unconformity at the base of the Quantico/Arvonia represents the Taconic unconformity in the Piedmont.

The basal sandstone of the Arvonia must be in part, at least, of shallow water deposition. The rest of the Arvonia/Quantico sequence seems to be a quiet, probably deep water sequence into which pelagic and turbiditic sedimentation occurred. The Buffards Conglomerate is interbedded with graded quartzo-feldspathic sandstone. Both facies are interleaved with Arvonia black slate. We interpret these conglomerates and sandstones to be debris flows and turbidites, respectively. Seiders and others (1975) also report graded sandstones with flute casts (turbidites) in the correlative Quantico Formation in northern Virginia.

If the Arvonia/Quantico sequence is of Late Ordovician age, it cannot be part of the Martinsburg clastic wedge on the west side of the Blue Ridge, because the Martinsburg shoreline had already prograded across the ancestral Blue Ridge and into the foreland basin by that time. This problem requires further research. A scenario consistent with the present sparse data base might be as follows: The Arvonia/Quantico may be post-Taconic, Silurian, deposited on the eroded roots of the Taconic mountains. Rapid development of quiet, or deep water conditions implies rifting (transtensional? Had the dextral transform movements of the Acadian and Allaghanian already begun?) and subsidence into a marine trough. This interpretation may be supported by the presence of turbidites and debris flows in a basin accumulating euxinic facies and would be consistent with the margins of the basin being at considerable distances from the fine -grained sequence now preserved.

Early Devonian - Early Mississippian Acadian Orogeny: A Manifestation of Oblique Collision

Time constraints on Acadian deformation along the James River traverse are poor and most of our knowledge of Acadian events comes from outside the area (Glover and others, 1983). Post Late Ordovician - Silurian (?) Paleozoic sedimentation is unknown in the Piedmont of Virginia, North Carolina and South Carolina.

In northern Virginia Pavlides (1982) has shown that metamorphism and deformation in the Falls Run Granite Gneiss occurred after 410 Ma and before intrusion of the oldest Falmouth rocks at about 322 Ma. Glover and others (1983) reviewed the evidence for the Acadian orogeny throughout the central and southern Appalachians (Figure 9) and concluded, from stratigraphic ages of the clastic wedge in the Valley and Ridge of Virginia, that the orogeny extended from the Middle Devonian (385 Ma) to Early Mississippian (360 Ma). Numbers in parentheses result from the new time scale (Palmer, 1983) which became available after Glover and others (1983). Isotopic data in the Piedmont suggest that the orogeny culminated at about 360 Ma (Early Mississippian). Acadian metamorphism and ductile deformation are mostly confined to an overprint on the more widespread Taconic metamorphism (Figure, 9). Acadian activity is manifest in the west central part of the crystalline terrain in the southern Appalachians and along the central and eastern part of the exposed Piedmont in Virginia (Figure 9).

Rodgers (1967) noted the differences in age of Acadian clastic wedges between New England and the Central Appalachians. Glover and others (1983) questioned whether the difference in timing of orogeny from north to south was diachroneity in the same orogeny or two separate events. In New England Acadian clastic wedges are Early and Middle Devonian, in New York to northern Virginia they are Middle and Late Devonian and in southern Virginia and Tennessee the clastic wedges are Latest Devonian to Early Mississippian (Ettensohn, 1987, and references therein). Ettensohn (1987) proposed oblique dextral collision of Avalon terranes with promontories on the North American continent to explain the southwestward migration of clastic wedges (Figure 10). Ferrill and Thomas (1988) recently proposed, on the basis of evidence for an Early Devonian wrench basin in the Talladega belt of Alabama and Georgia, that oblique faulting also extended into the Alabama segment at that time. According to their modification of Ettensohn's model, wrench and transpressional stress varied along the orogenic system with time as shown in Figure 9.

The oblique collision model of Ettensohn (1987) and modifications by Ferrill and Thomas (1988) seems plausible. For the central and southern Appalachians, however, it could not have been the initial collision with Carolinia ("Avalonia") because, as previously shown, Carolinia collided with Laurentia during the Ordovician Taconic orogeny. A more likely hypothesis is that the amalgamated terranes of Laurentia and Carolinia collided with southern Europe and west Africa. Middle Devonian metamorphism and deformation are recorded in France and Morocco (Robinson and others, in press).

Along the James River traverse, during the Acadian, dextral transpression probably reactivated many of the Taconic faults and undoubtedly moved the ancestral Blue Ridge closer to its present position.

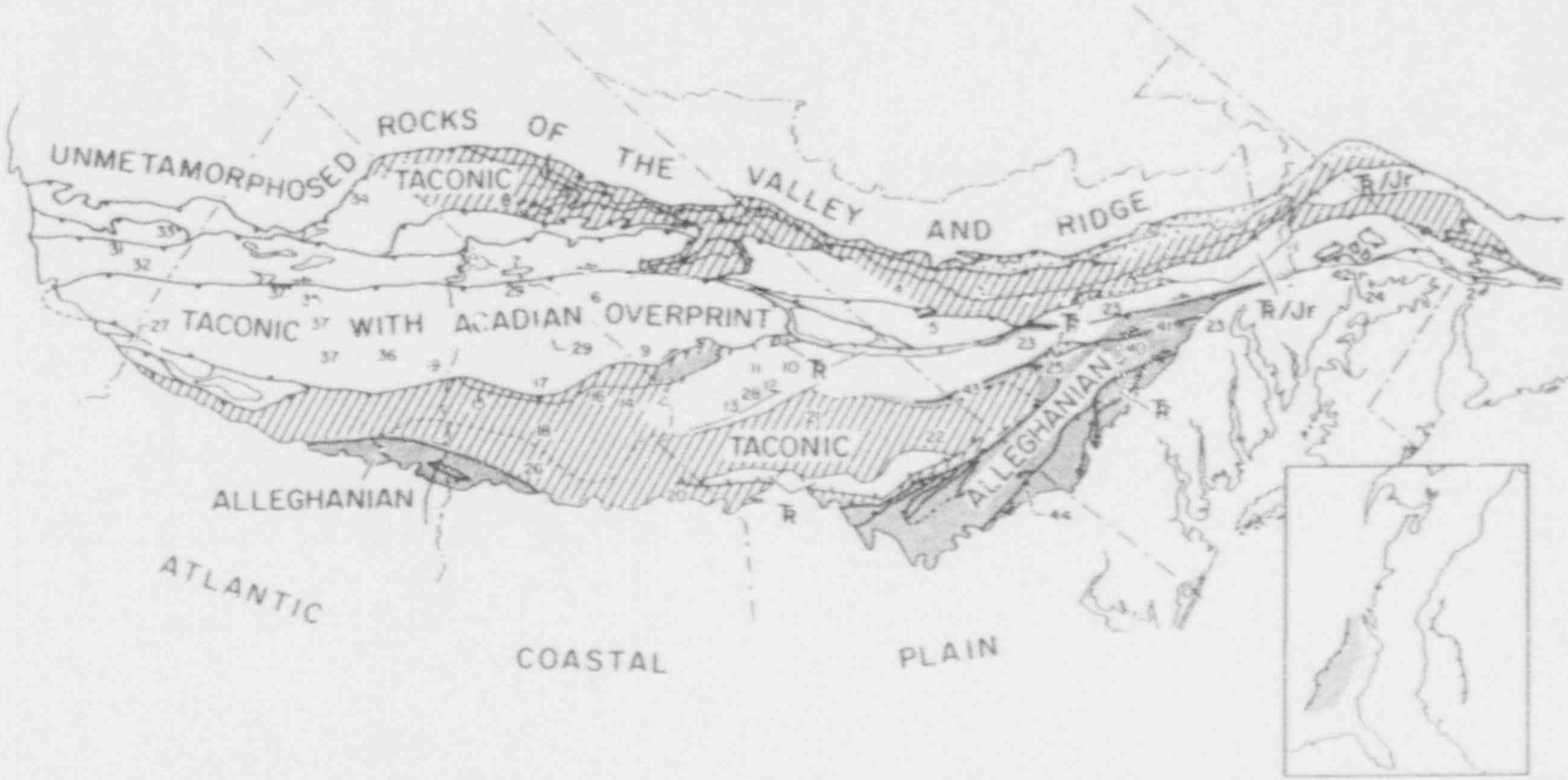

Figure 9. Ages of metamorphism and ductile deformation in the central and southern Appalachians. From Glover 1989.

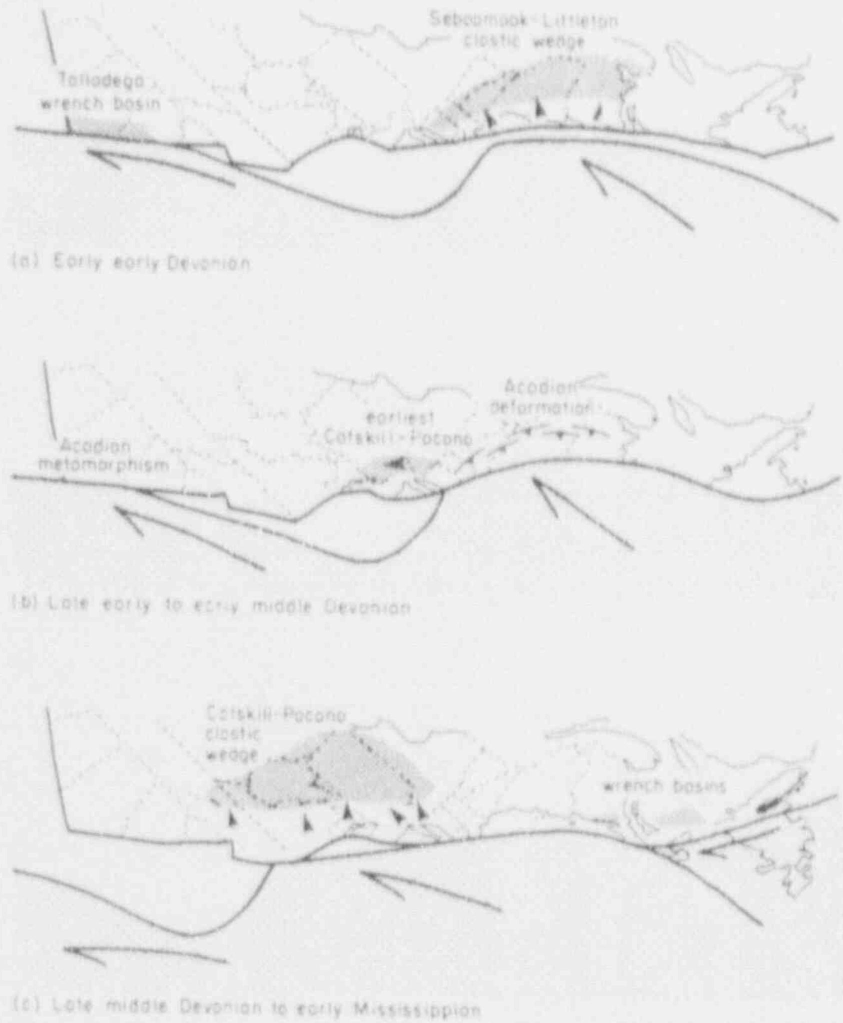

Figure 10. "Sequential maps showing Acadian oblique convergence along Appalachian margin of North America. Sites of synorogenic accumulation (arrows mark directions of sediment dispersal) and tectonic elements of the North American plate. Arrows on other plate show sense of motion between plates." From Ferrill and Thomas, 1988, p. 607.

Other Tectonic Models. Hatcher (1987) considers the suturing of Avalonia (Carolinia) to North America to be an Acadian event and to have occurred along the western boundary of the Kings Mountain belt in the southern Appalachians. In Virginia this boundary on his map (Hatcher, 1987, Figure 2) passes east of the Chopawamsic Formation and is overthrust by the Goochland basement nappes. The ultimate basis for Hatcher's ages of collision-boundaries seems to be the need to match a sequence of three orogenic events with a succession of terranes eastward (outward) from the North American craton. In this paper it has been shown that the western (Taconic) suture is mostly misplaced by Hatcher in Virginia, and that the collision with Carolinia occurred during the Taconic (see also Glover and others, 1983).

Early Mississippian/Permian Alleghanian (Hercynian) Orogeny

In the Valley and Ridge of Virginia and West Virginia the Greenbrier Limestone, Merrimacian/Visean in age, was deposited upon the Acadian clastic wedge in a brief interval about 340 m. y. ago. The Greenbrier is immediately overlain by the Alleghanian clastic wedge which ranges into the Permian, or to about 250 Ma. This brief pause in clastic deposition probably represents a change in style of deformation, perhaps the passage of a non-compressional transform junction with southern Europe and Africa. In any event it hardly seems reasonable to consider it more than a brief lull, or change in style, in an otherwise continuous orogeny.

Glover and others (1983) found that the metamorphic thermal peak in the eastern Virginia Piedmont was reached about 280 Ma (Early Permian), and that ductile deformation over most of the eastern Piedmont ceased about 250 Ma (Late Permian). Thus deformation in the Piedmont was synchronous with deposition of the clastic wedge in the foreland basin.

The Petersburg Granite of the eastern Piedmont, near Richmond (Plate 1), is Late Mississippian, 330 ± 8 Ma according to Wright and others (1975) and was deformed intensely along its western margin by the dextrally transpressive Hylas mylonite zone (Bobyarchick and Glover, 1979; Gates and Glover, 1989).

Gates and others (1986) have shown that late Paleozoic dextral transpressional (ductile) faulting, parallel to the orogenic axis, occurred in the crystalline terrain throughout the length of the Appalachians. They inferred from isotopic dates that this occurred from 324 Ma to 285 Ma (Early Mississippian to Early Permian). The 324 Ma older bound is a cooling age on hornblende (Glover and others, 1983) and the actual age of initial deformation is undoubtedly older. Similarly, the 285 Ma age is now superceded by more recent findings, mentioned above, that suggest dextral transpression continued until about 250 Ma.

The I-64 seismic profile (Plate 1) shows the structural relation of North American Grenville basement to the overthrust Carolinia rocks in the eastern Piedmont of Virginia. In the Goochland nappe, foliation, lithologic layering, and mylonite zones are all parallel as a result of the final deformation during the Alleghanian orogeny. Surface studies indicate that these three parallel features of the layering are recorded in the reflections on the seismic record as shown below the surface in the Goochland nappes on the profile.

Therefore the dextral transpression that formed these structures occurred along moderately eastward dipping zones that, as shown by the profile, reached into the middle and lower crust. The emplacement of the Goochland nappes and the doming of the Carolinia cover therefore is a consequence of Alleghanian dextral transpression. Alleghanian amphibolite facies metamorphism that retrograded the Grenville granulite mineral assemblages occurred at about 5-7 kbars (Farrar, 1984) equivalent to uplift of 15-20 km during the deformation. Subsequent erosion has breached the Carolinia cover in Virginia and parts of North Carolina (Figure 2). Now the Grenville basement rocks in the nappes plunges gently southward forming the Raleigh belt of North Carolina (Figures 1, 2). It seems likely that the entire southern Piedmont is underlain by North American Grenville basement.

Alleghanian regional metamorphism and ductile deformation is largely confined to the eastern Piedmont of the central and southern Appalachians (Figure 9). Localized zones of Alleghanian deformation are imposed on Alleghanian granites along the Brevard zone in Georgia (Glover and others, 1983, and references therein) and on the High Shoals Granite at the northern end of the Kings Mountain belt (Figures 1, 2) in North Carolina (Horton and others, 1987). These occurrences represent areas of ductile deformation along or near fault zones that were reactivated and locally intruded by granites during the Alleghanian; undoubtedly more will be found. Ductile deformation of Alleghanian age also occurs in central Virginia where reactivation of the Taconic suture in Alleghanian time produced mylonites of this age (Gates, 1981, Gates and others, 1986). Although Alleghanian ductile deformation is found along major reactivated faults in the central and western Piedmont (Figure 9) the eastern Piedmont contains the only large areas of regional ductile deformation and metamorphism known at this time. Therefore, the pattern of regional metamorphism and ductile deformation in the Piedmont and Blue Ridge (Figure 9) supports our conclusion (Glover and others, 1983) of a general southeastward migration of thermal metamorphic events with time. This seems in accord with the conclusion that collision zones also become younger eastward as a natural consequence of the succession in collision described herein.

The timing of the development of the principal areas of Acadian and Alleghanian metamorphism and ductile deformation in the central and southern Appalachian Piedmont was contemporaneous with the formation of their respective clastic wedges in the Valley and Ridge. During deposition of the Acadian clastic wedge parts of the Blue Ridge province (those parts not overprinted by Acadian ductile deformation, Figure 9) were transported westward as largely rigid blocks deforming the shelf and Taconic clastic wedge strata and edge of the Acadian foreland basin in front of them. During the deposition of the Alleghanian clastic wedge the Blue Ridge, central and western Piedmont blocks moved in a dominantly rigid state deforming the shelf, Taconic and Acadian foreland basin strata ahead of them. Foreland basin strata deformed by pre Alleghanian orogenies were overridden during the Alleghanian so that most of the deformation now seen in the Valley and Ridge is of Alleghanian age. Because of this sequence of events the Blue Ridge and associated faults that carry metamorphosed terrain over unmetamorphosed Valley and Ridge strata in the southern Appalachians is commonly thought to be Alleghanian. The sequence of events outlined above suggests that these faults had their inception much earlier during the Taconic collision.

Origin of Middle and Late Paleozoic Plutons

Middle- to late-Paleozoic granitic rocks of the Piedmont were largely generated during times of crustal thickening by Acadian and Alleghanian transform, transtensional and transpressional collision. Possibly some 410-385 Ma (Early Devonian) gabbroic, syenitic and granitic plutons in the central Carolinas correspond to a non-compressive interval of time and their compositions suggest that they may have had an transform or transtensional origin in accord with the timing and nature of the Acadian orogeny outlined above

The Acadian and Alleghanian granites of the Piedmont are mostly monzogranites, but range through thondhjemite to monzonite to syenogranite (Speer, Becker and Farrar, personal communication, 1984). Only about 10% of the coeval intrusives are gabbro. Thus the suite is strongly bimodal and lacks diorite. A magmatic arc origin is unlikely.

Crustal thickening during compressive and transpressive events provides a number of attributes that promote generation of melt. Frictional heat accumulates and lower crust and upper mantle are depressed into higher temperature zones. The transcurrent component of movement may more locally create higher and lower pressure zones within the lithosphere. Crustal shortening may result in delamination and sinking of lower lithosphere, and allow upwelling of hot asthenosphere to the base of the crust. Rapid upward transport during nappe stacking, and isostatic rebound during quiet times between compressive events may make decompression melting possible.

Early Mesozoic Rifting Precursor of Atlantic Ocean Basin Opening.

Numerous rift basins of Late Triassic and early Jurassic age exist in the Piedmont of the central and southern Appalachians. Dikes of Jurassic diabase (dolerite of European usage ?) record the beginning of the generation of the Atlantic Ocean and the current plate tectonic regime. They are beyond the scope of this report and are mentioned only to provide an end point for the evolution of the Appalachian orogenic system.

Conclusions

Rifting of Grenvillia began about 690 Ma and continued to about 570 Ma (Early Cambrian). At least two stages can be seen in this protracted extensional event: The first stage, 690-650 Ma, was characterized by rifting, eruption of the tholeiitic and peralkaline Crossnore plutonic-volcanic suite, and accumulation of non-marine clastic sediments and volcanic rocks in graben. Glaciation occurred at the beginning of the second stage (and perhaps at the end of the first stage), and widespread subsidence occurred along the east flank of the present Blue Ridge. Subsidence led to the development of a retrogradational braided submarine fan over attenuated continental crust as the proto margin of Laurentia evolved. The second stage of rifting was accompanied by intrusion of basaltic dikes and sills and eruption of lava. Mafic/ultramafic dikes and sills were injected during the early part of the interval. Rifting was complete at 570 Ma. and the new ocean basin Iapetus was initiated.

The drift stage began at 570 Ma and continued to about 490 Ma. (Early Ordovician). During this time the Laurentian continent drifted in tropical seas and accumulated nearly three kilometers of quartz arenite and carbonate on its shallow platformal edge.

Somewhere, at high latitudes in the Iapetan Ocean, the microcontinent Carolinia was being mantled by calcalkaline volcanic rocks as the ocean closed between it and Laurentia. The oldest known Carolinia volcanism is older than 700 Ma. A little before 600 Ma. the volcanism was interrupted by the Virgilina deformation which folded and faulted the older volcanic sequence. Deep submarine basins (transtensional(?) graben) filled with turbidites formed over the core region of the older arc. This deformation and sedimentation pattern suggests a change in direction of plate movement followed by transpressional-transtensional collision. When volcanism resumed after the Virgilina deformation it was more nearly bimodal in composition. The younger period of volcanism lasted from about 600 Ma until about 510 Ma. Collision with Laurentia occurred between 510 and 490 Ma., probably during the Early Ordovician.

Sedimentary and tectonic melanges accumulating in the trench and accretionary wedge were brought ashore during the Taconic collision. Ages of crosscutting intrusives and source materials suggest that much of this melange was deposited about 550-525 Ma. However, melanges undoubtedly began to form as early as 700 + (?) Ma. because the magmatic arc was already well developed by then.

During the Taconic collision the hinge zone of the Laurentian continent was sliced off and thrust cratonward as the proto Blue Ridge, while the foreland basin filled with clastic sediments culminating in deposition of the thick Martinsburg and Juniata formations.

After the Taconic collision the North American continent drifted at low latitudes for about 30 m.y., collecting compositionally mature shallow water clastics and carbonate sediments along its platformal margin. Minor rifting may have occurred in the Piedmont.

Collision with South America(?) Africa and southern Europe probably began during the Silurian or Early Devonian. This collision was initially strongly dextral-transpressional in New England but dominantly dextral-transform in the central and southern Appalachians. By the Middle and Late Devonian the region of intense dextral transpression had moved into the central Appalachians. During the early Mississippi (Greenbrier time) the central Appalachians were probably experiencing transform motion. Transpressional motion was dominant in the central and southern Appalachians during the Alleghanian (Hercynian) orogeny. The Acadian and Hercynian orogenies appear to be parts of a single protracted oblique collisional event (should we resurrect the term "Appalachian orogeny" with Acadian and Alleghanian/Hercynian phases?). Possibly the obliquity of the collision angle is the reason why subduction was not aborted and collision ceased much sooner.

Rifting began again during the Late Triassic and by Middle Jurassic the present Atlantic ocean had begun to form.

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Part B: Velocity and Q studies in central Virginia

By John K. Costain

Purpose of study

In a general sense, different rock types have characteristic seismic velocities; the higher the velocity, the higher the value of the quality factor, Q (Waters, 1978). The objective of this study was to identify subsurface rock types in the structurally complex crystalline Piedmont of Virginia by measurements of velocity and Q from reflection seismic data acquired on lines NRC2A1-1, NRC2A1-2, NRC2A1-3, NRC2A1-4, and NRCRRT-1. If Q could be determined, then using published values for general relationships between Q and velocity, rock velocity could be inferred. Velocity determinations from stacking velocities can be inaccurate and ambiguous because of residual statics problems, choice of the reference datum for data processing, reflections from out of the plane of the section, etc.

Intrinsic damping (i.e., Q) is difficult to measure. A review of various laboratory and field techniques as well as results was given by Toksoz and Johnston (1981). If reflection seismic data are used, then many factors can affect the measurement of Q if ratios of the amplitudes of seismic wavelets are used, or if ratios of their spectral components are used. These factors have been summarized repeatedly in the literature (i.e., Toksoz and Johnston, 1981). What is needed is a method that is relatively independent of source/receiver coupling, is independent of spherical spreading, independent of reflection or transmission coefficients, and insensitive to residual statics shifts beneath either the source or receiver — in other words, a method that depends upon shape instead of amplitude. Ecevitoglu (1987) and Ecevitoglu and Costain (1988) published a unifying approach to the numerical modeling of intrinsic damping that facilitates observations of its effect on the shape of the propagating seismic wavelet, and allows for a direct determination of Q; this was the method used herein to analyze the reflection seismic data from the crystalline Piedmont of Virginia.

The quality factor, Q, is greatly affected by the presence of free water. For example, the Q of a low-porosity (1-2%) olivene basalt without hydrated mineral phases ranges from about 100 in normal laboratory air to over 2,000 when outgassed at moderate temperatures in a high vacuum. Birch and Bancroft (1938) showed that Q (at least in the kilohertz frequency range) of many igneous rocks increased from values of a few hundred to values of about 2000 at depths of about 10 km. Exposure to even laboratory air drastically reduces the value of Q from values greater than 2000 to around 100 (Tittman 1981).

Theoretical background

Ecevitoglu (1987) and Ecevitoglu and Costain (1987) formulated an exact expression to describe the effects of absorption and body wave dispersion on the shape of a seismic wavelet for which the attenuation coefficient, a(n), is an arbitrary function of the frequency, n. They showed that absorption-dispersion pairs can be computed using the discrete numerical Hilbert transform, and that approximate analytical expressions requiring the selection of arbitrary constants and cutoff frequencies are no longer necessary. For constant Q, the dispersive body wave velocity, p(n), is

$$p(\nu) = \frac{p(\nu_N)}{1 + \frac{1}{2Q}\frac{H(-\nu)}{\nu}} \quad ,$$

where H denotes numerical Hilbert transformation, p(n) is the phase velocity at the frequency n, and p(nN) is the phase velocity at Nyquist. From the above equation, it is possible to estimate Q in the time domain by measuring the amount of increase, DW, of the wavelet breadth after a traveltime, Dt, by

$$Q = \frac{2\Delta\tau}{\pi\Delta W} \qquad (2)$$

Aki and Richards (1980, pp. 172-177) summarized difficulties with analytic expressions for phase velocity that involve frequency limits of integration that extend to infinity. In order to calculate the "real, physical" phase velocity at some specific frequency, then application of the Hilbert integral for frequency limits from zero to infinity will result in an unbounded phase function that implicitly includes a linear phase (due to traveltime), a dispersive phase, B(n), (due to body wave dispersion), and a "hidden" phase (due to approximations that must be made for frequency limits of infinity). A graphic summary of the phase definitions of Futterman (1962), Strick (1970), and Kjartansson (1979) are given in Ecevitoglu and Costain (1988). The purely dispersive phase spectrum, B(n), is shown in Ecevitoglu and Costain (1988), i.e., with the linear phase due to traveltime subtracted.

Aki and Richards (1980, Eq. 5.74) noted that

$$A(x) = A_0 e^{-\left[\frac{\omega x}{2cQ}\right]}$$

implies

$$\frac{\omega}{C_\infty} + H[\alpha(\omega)] = 2Q\,\alpha(\omega)$$

where H denotes numerical Hilbert transformation; there is no Hilbert transform pair for which this relation is satisfied with constant Q. Such problems can be overcome by using a discrete Hilbert transformation with a finite upper frequency limit instead of integration with an upper frequency limit of infinity, and by recognizing that the total phase spectrum can be split into a linear-with-frequency nondispersive phase defined by the traveltime of a reflected event, plus a purely dispersive phase spectrum that is associated with body wave dispersion brought about by causal absorption. Aperiodic dispersive phase terms are unbounded; therefore, they always implicitly subtract, as in Futterman (1962), or add, as in Strick (1967), or both add and subtract (i.e., "bend" a(n) versus n as in Kjartansson, 1979) some amount of pure, undesirable, time delay. The expressions for the absorption coefficients of Azimi et al. (1968) are convex upward because they have to satisfy the Paley-Wiener condition. This condition restricts the permissible choices of a(n) versus n to those that increase slower than the first power of the frequency as the frequency goes to infinity. This restriction is removed for the case of real data and the realities of a Nyquist frequency by invoking periodic theory in which Kolmogorov's condition is instead satisfied.

The Hilbert transform as given in Lee (1960, Equation 65) is used to formulate the relation between the amplitude and phase spectra of an absorptive filter:

$$B(\nu) = -4\int_0^\infty \sin(2\pi\nu t)\,dt \int_0^\infty \ln[A(\nu')]\cos(2\pi\nu' t)\,d\nu' \qquad (3)$$

where A(n) and B(n) are the amplitude and phase spectra, respectively, of the absorptive filter. According to the first assumption:

$$A(\nu) = e^{-\alpha(\nu)x} = e^{-b\nu}$$ and

$$\ln[A(u)] = \ln\left[e^{-bu}\right] = -bu \qquad (4)$$

Substituting (4) in (3) we obtain:

$$B(\nu) = 4\int_0^\infty \sin(2\pi\nu' t)\,dt \int_0^\infty b\nu' \cos(2\pi\nu' t)\,d\nu' \tag{5}$$

For an upper limit of integration of ∞ in (5), the rightmost integral becomes infinite. This is the difficulty encountered by earlier workers who used "aperiodic theory" (i.e., n '= ∞) when a(n)a = bng with g = 1. To overcome this difficulty, they chose g close to 1 (say, g = 0.9). This approach has been called (Strick, 1967) "power-law attenuation," and a(n) is chosen such that Q is almost constant over the frequency range of interest. Instead of relaxing Q and allowing it to become "slightly" frequency dependent, Ecevitoglu and Costain (1988) proceeded directly with the integration and selected some arbitrary Nyquist cutoff frequency.

Let q(n) be the total (dispersive phase spectrum plus the linear-with-frequency phase corresponding to nondispersive traveltime) phase spectrum:

$$\theta(\nu) = B(\nu) + 2\pi\nu\tau \tag{6}$$

where B(n) is defined here as the pure body wave "dispersive phase" and 2pnt is the phase that corresponds to some pure traveltime, t. Thus,

$$\theta(\nu) = 2\pi\nu\tau = 2\pi\nu \frac{a}{p(\nu)} \tag{7}$$

where t is now total time (the sum of the traveltime plus a frequency-dependent time delay due to body wave dispersion), a is the travel distance, and p(n) is the dispersive phase velocity. Let

$$\tau = \frac{a}{p(\nu_N)} \text{ and } b = \frac{\pi a}{Qp(\nu_N)} \tag{8}$$

where t is the traveltime, p(nN) is the phase velocity at the Nyquist frequency and is also the p-wave velocity of the medium without absorption, and a is the travel distance.

Although a linear-with-frequency attenuation is used here as an example, the numerical approach presented is appropriate for any behavior of a(n) versus n. This means that the inverse problem, that of determining a(n) versus n from Hilbert transformation of the phase spectrum as derived from real data, will reveal the nonlinear dependence of a(n) versus n if it is present in the data.

The phase spectrum, B(n), for linear-with-frequency absorption as obtained by the exact numerical procedure is:

$$B(v) = H[\ln A(v)] = H[\ln e^{-bv}] = bH(-v) \tag{9}$$

Here, H stands for numerical Hilbert transformation. From (6), (7), (8), and (9) the phase velocity is:

$$p(v) = \frac{p(v_N)}{1 + \frac{1}{2Q}\frac{H(-v)}{v}} \tag{10}$$

and thus every value of dispersive velocity can be computed from (10) from n = 0 to n = nN, inclusive. There are no arbitrary constants to choose. The effects of different absorption levels on the dispersive phase and impulse response are shown in Figure 11 for distances of 2, 4, 6, and 8 km from the vibrator source. (Note that the absorption coefficient is multiplied by the travel distance, a.) Observe the scaling and broadening effects on the causal absorption impulse responses (lower). As a(n)a versus n becomes steeper, then the dispersive phase B(n) becomes larger so that the peak of the pulse in the time domain is gradually delayed.

The percent, D, of body wave dispersion from n = 0 to n = nN is

$$D = \frac{p(v_N) - p(0)}{p(v_N)} \times 100 \tag{11}$$

and it is thus possible to determine a value for D over any frequency bandwidth.

From Equation (10), the phase velocity at n = 0 (see Ecevitoglu, 1987) is:

$$p(0) = \frac{p(v_N)}{1 + \frac{2}{\pi Q}}$$

and body wave dispersion, D, over the entire bandwidth from n = 0 and n = nN for a linear-with-frequency absorption coefficient is

$$D = \frac{p(v_N) - p(0)}{p(v_N)} = \frac{1}{1 + \frac{\pi}{2}Q} \tag{12}$$

From Equation (12), the value of D is independent of any frequency cutoff, and depends

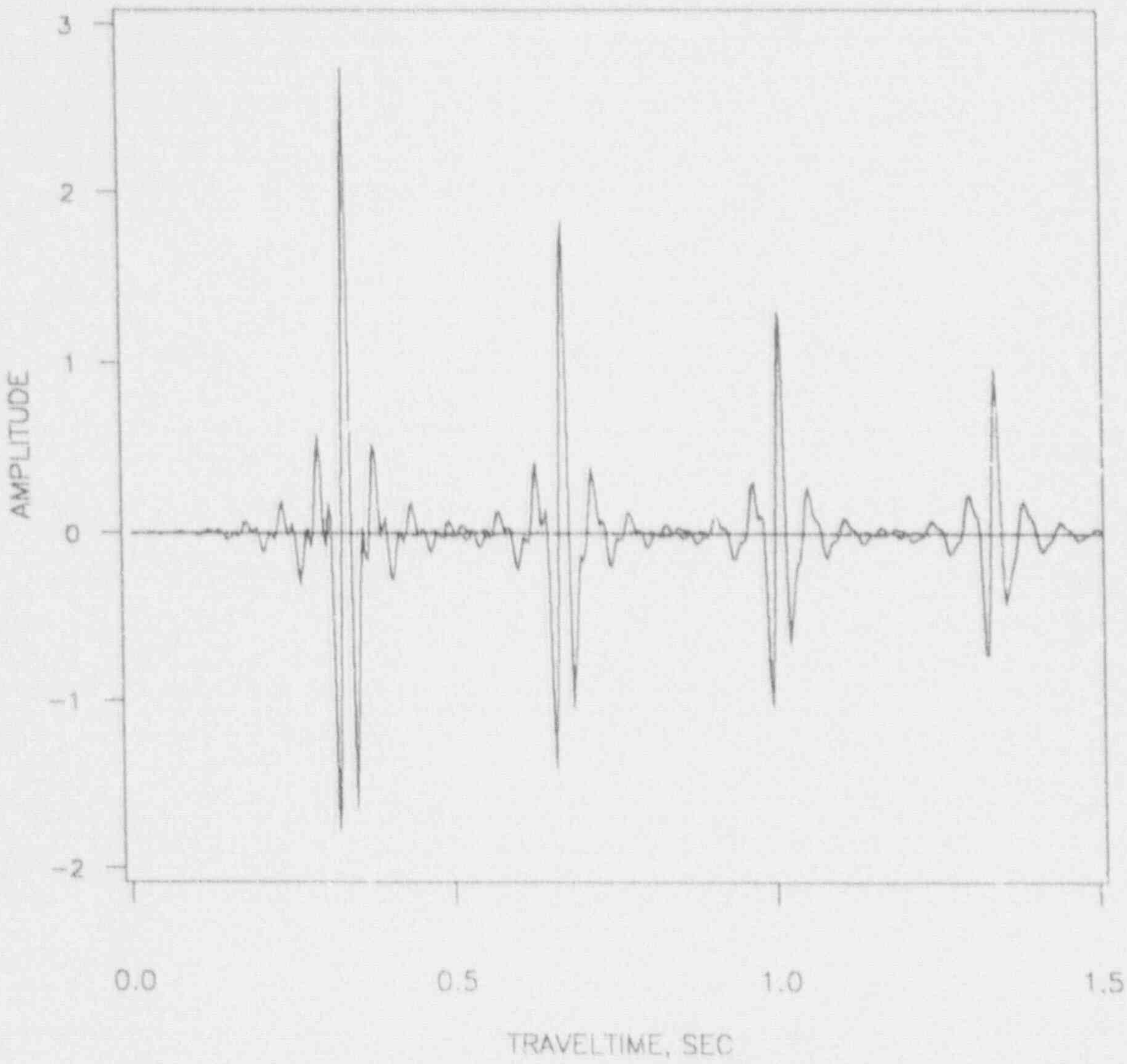

Figure 11. Four synthetic vibroseis wavelets attenuated by intrinsic damping in a medium of Q = 100: The wavelets are recorded at distances of 2, 4, 6, and 8 km from the source. The Klauder wavelet at the source is approximately symmetrical about a central peak. Note the loss of symmetry of the Klauder wavelet due to causal attenuation. Such wavelets are used to determine Q by measuring the change in time between the two lowest peaks (this change is DW in Equation 13) as the reflected wavelet is recorded at successively greater distances from the vibrator source. For example, in Equation (13), DW is the amount of wavelet spreading between the second and first reflected wavelets, and t is the difference in traveltime between the second and first recorded wavelets.

only on Q.

We now compute the difference in dispersive time delay, DW, between frequencies n = 0 and n = nN. Both of these frequencies traveled the same distance a. Therefore,

$$a = p(v_N)\tau = p(0)\,(\tau + \Delta W)$$

where t is the pure traveltime. Then

$$\Delta W = \frac{p(v_N) - p(0)}{p(0)}\,\tau$$

From the expression for body wave dispersion, we have

$$\frac{p(v_N) - p(0)}{p(0)} = D\,\frac{p(v_N)}{p(0)} = \frac{1 + \frac{2}{\pi Q}}{1 + \frac{\pi Q}{2}} \approx \frac{2}{\pi Q}$$

Therefore,

$$\Delta W = \frac{2\Delta\tau}{\pi Q}$$

or

(for constant Q)

$$Q = \frac{2\Delta\tau}{\pi\Delta W} \qquad (13)$$

Equation (13) makes possible time-domain measurements of Q. Dt is the traveltime (or the difference in traveltime between an event arriving at different receiver locations) and ΔW is the amount of wavelet breadth increase during the time Dt. Small values of body-wave dispersion (D) for high-Q rocks require resampling the data in order to see the increase in wavelet breadth, DW. This is discussed further in the Procedures section.

Futterman's (1962) velocity dispersion expression superimposed in a (necessarily) piecewise manner upon the exact curve of Ecevitoglu and Costain (1988) was published by Ecevitoglu and Costain (1988) who generated a dispersive velocity curve for nN = 125 Hz and Q = 250 from 0 to 125 Hz and computed pieces of Futterman's phase velocity curve using the same value of Q. Excellent agreement with Futterman was obtained. It is not possible to superimpose Futterman's entire results with a single selection of his constants, c0 (km/sec) and n0 (Hz). In Equation (10), there is no arbitrary constant other than Q that governs the shape of the velocity dispersion curves computed from discrete

Hilbert transformation; p(nN) is just a scale factor.

Procedure

The input data for the velocity-Q studies are raw, unstacked traces from the field tapes that have not been pre-whitened or filtered. Application of Equation (13) requires

1. A reflected event with a good signal-to-noise ratio,

2. No interference between adjacent reflections. This constraint is not as serious as it might seem, because it is only the central part of the Klauder wavelet (the part with the two side lobes on each side of the central peak) that is used in the analysis (see Figure 1).

Correlated shot records used for subsequent conventional stacks are inspected for isolated reflections that can be followed across the record. The uncorrelated data is then recorrelated, without vibroseis whitening (Çoruh and Costain, 1983), from the original field tapes.

Three methods of analysis were used:

1. Follow a reflection on a common-source record,

2. Follow (the same) reflection in a CDP gather,

3. Measure DW between two different reflections on the same record, but at different times.

Although all methods were examined, the last was preferred because of the common source-coupling and relatively short offset (70 m) for high-speed rocks (6 km/sec). For high-Q rocks, DW in Equation (13) will be less than the original sampling interval (4 ms) when the data were acquired. The data, f(t), must therefore be resampled at a smaller sampling interval (say 0.25 ms) by application of the Fourier transform pair:

$$F(\omega) = \int_{-\infty}^{\infty} f(t)e^{-i\omega t}dt$$

$$f(t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} F(\omega)\, e^{-i\omega t} d\omega$$

where t is now assigned values 0, 0.25 ms, 0.5 ms, etc. If this is not done for crystalline rocks, it is not possible to measure the amount of wavelet spreading, DW, using Equation (13). In order to estimate the appropriate resampling interval, Equation (13) can be evaluated for different assumed values of Q and different traveltimes, t, and wavelet-spreading, DW. Results are shown in Figure 12. For high values of Q it can be seen that spreading is a fraction of a millisecond; the new location of the peak (or trough) of the wavelet after spreading would therefore not be detected without resampling. It was concluded that resampling at 0.25 ms was adequate for the values of Q anticipated for the crystalline rocks in central Virginia. For example, values of t for traces at different offsets (70-meter spacing), x,
for an event reflected from a depth of 6 km would be

$$\tau_{(x=0)} = \sqrt{\frac{x^2}{v^2} + t_0^2} = 2 \text{ sec}$$

$$\tau_{(x=70)} = \sqrt{\frac{x^2}{v^2} + t_0^2} = \sqrt{\frac{70^2}{6000^2}} = 2.0000 \text{ sec}$$

$$\tau_{(x=1680)} = \sqrt{\frac{x^2}{v^2} + t_0^2} = \sqrt{\frac{1680^2}{6000^2} + 2^2} = 2.0195 \text{ sec}$$

$$\tau_{(x=1750)} = \sqrt{\frac{x^2}{v^2} + t_0^2} = \sqrt{\frac{1750^2}{6000^2} + 2^2} = 2.0212 \text{ sec}$$

where t0 is vertical traveltime (x = 0). Between adjacent receivers (at x = 1680 meters), Dt is 1.7 millisecond.

It should be noted from Figure 2 that a saprolite layer 50 m in thickness (Q = 50, v = 1 km/sec) will have the same effect on the shape of a vibroseis wavelet as 3000 meters of crystalline rock (Q = 500; v = 6 km/sec). For this reason, an attempt was made to determine refraction intercept times for head waves from the base of the saprolite in order to estimate relative saprolite thickness. The results were not definitive, however, because of the poor quality of the first breaks on the vibroseis data.

Line NRC2A1-4 provided reliable reflection continuity and offered the most opportunity for tracking clean waveforms across at least a portion of a common-shot

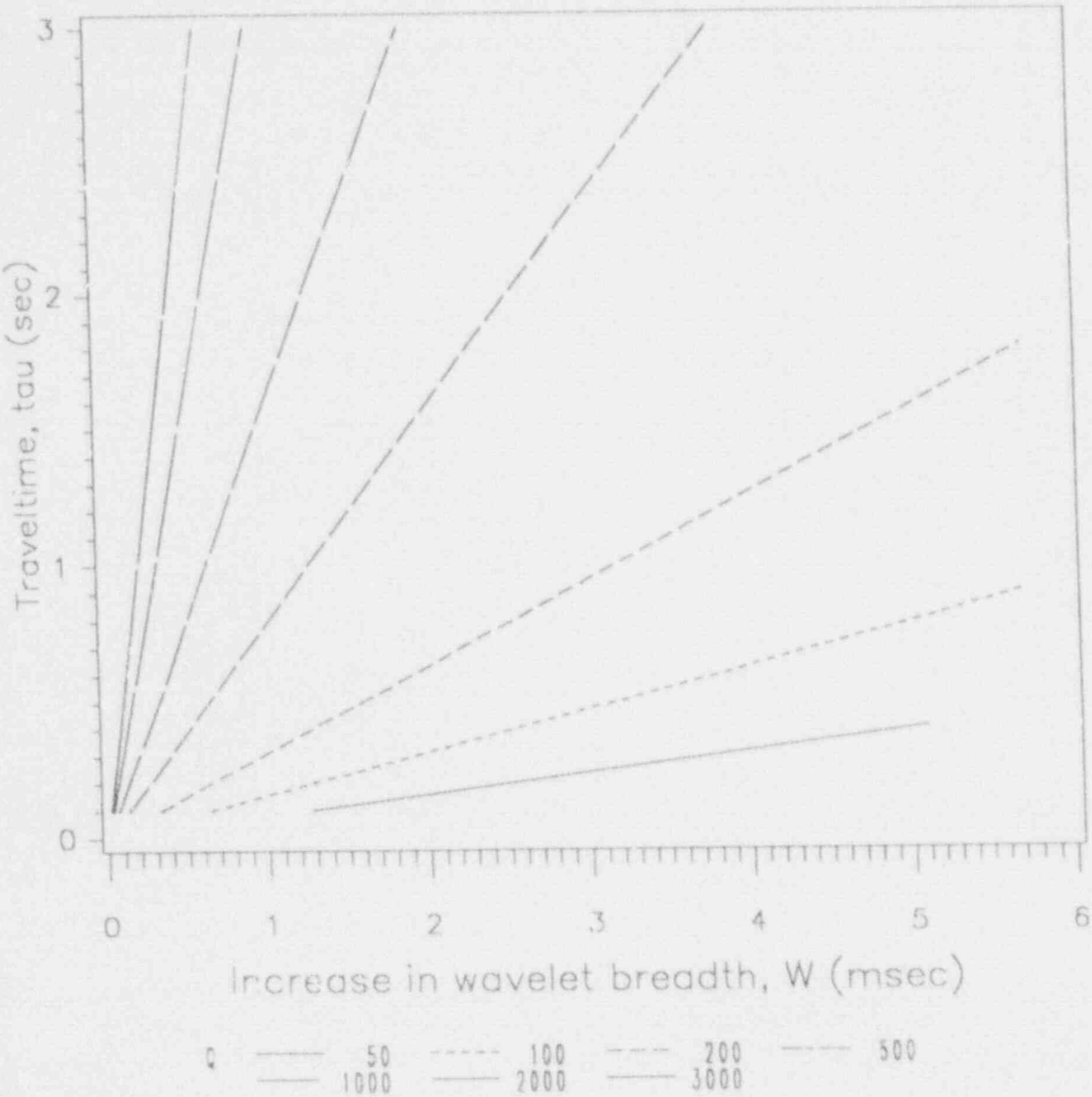

Figure 12. Effect of the quality factor, Q, on wavelet shape (spreading) as a function of traveltime, t, and wavelet spreading DW, for a vibroseis Klauder wavelet: Traveltime is in seconds; wavelet spread is in milliseconds. A Klauder wavelet at the source is symmetrical about a central peak. The increase in time between the two side lobes of the Klauder wavelet is DW. For high values of Q (low intrinsic damping), changes in DW as traveltime increases are small, and will generally be less than one sample interval (4 ms for the NRC data) from trace to trace. To measure Q, therefore, might require a Fourier interpolation to a smaller sample interval.

record. Reflection continuity with a high S/N ratio across the entire record was rare. For high-Q rocks this means relatively little spreading associated with normal move-out; however, this is not a problem if a deeper reflection on the same record can be compared with a shallow one. Comparisons of wavelet spreading between a shallow and a deep reflection on a portion of the same shot record provided the most convincing results.

A representative interval (tape D0215, Shot 23) recorrelated without vibroseis whitening, is shown in Figure 13 and Figure 14 from 0-1.7 sec and from 2.2-3.1 sec, respectively. This illustrates an example of the desirable signal-to-noise ratio for this type of analysis. The idea is to examine enlarged portions (circled in the figures) of the data and track the waveform spreading, DW, from trace to trace. From this spreading, Q can be determined, and a rock type inferred.

A representative example of resampling a Klauder wavelet to a considerably smaller interval is shown in Figure 15. The idea is to find the maxima of the peaks of the Klauder wavelet so that DW can be measured.

Signal-to-noise ratios on field records of the other lines were judged to be too low for reliable determinations of Q, or else the values of Q turned out to be negative, probably due to interbed multiples. The geology is complex. The S/N ratio improves after stack; however, the basic waveform shape for application of Equation (13) is lost.

The data from Line NRC2A1-4 was judged to be satisfactory for the analysis. Over short distances, the data were at least as good as that in the only other location in crystalline terrain where this method has been successfully applied. Values of Q determined from Equation (13) side-lobe to side-lobe times at 1.5 sec were in the range 29-34 ms (average=32 ms); at 2.5 sec, this period was 32-34 ms (average=33 ms) From Equation (13), Q = 650.

Conclusions

The generally observed lack of spreading of Klauder source wavelets observed on the unprocessed reflection seismic data from Line NRC2A1-4 in Virginia is interpreted to be due to a high Q (very little intrinsic damping), implying a relatively dry crust (at least as sensed by seismic waves of wavelength 100-400 meters) in this part of central Virginia. Because exposure even to laboratory air drastically reduces the value of Q from

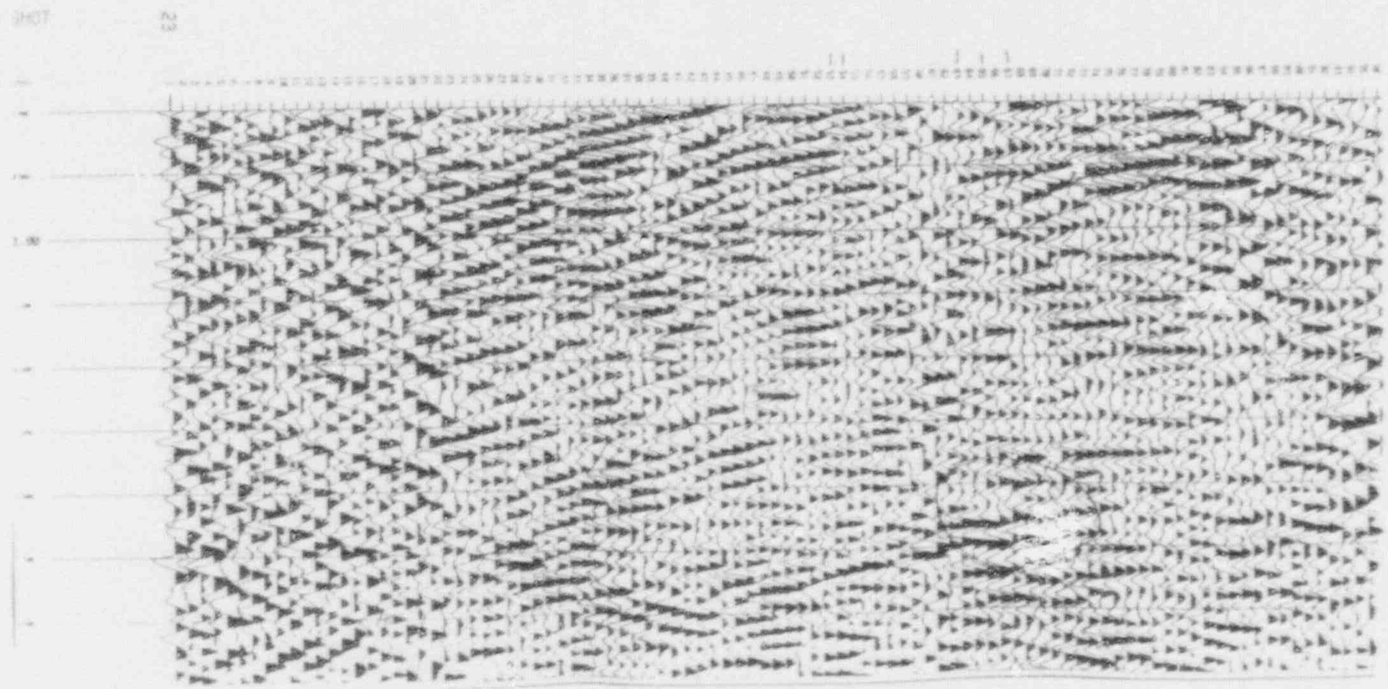

Figure 13. Sample shot record used in analysis: Line NRC2A1-4, shot 23, time 0.8-1.7 sec.

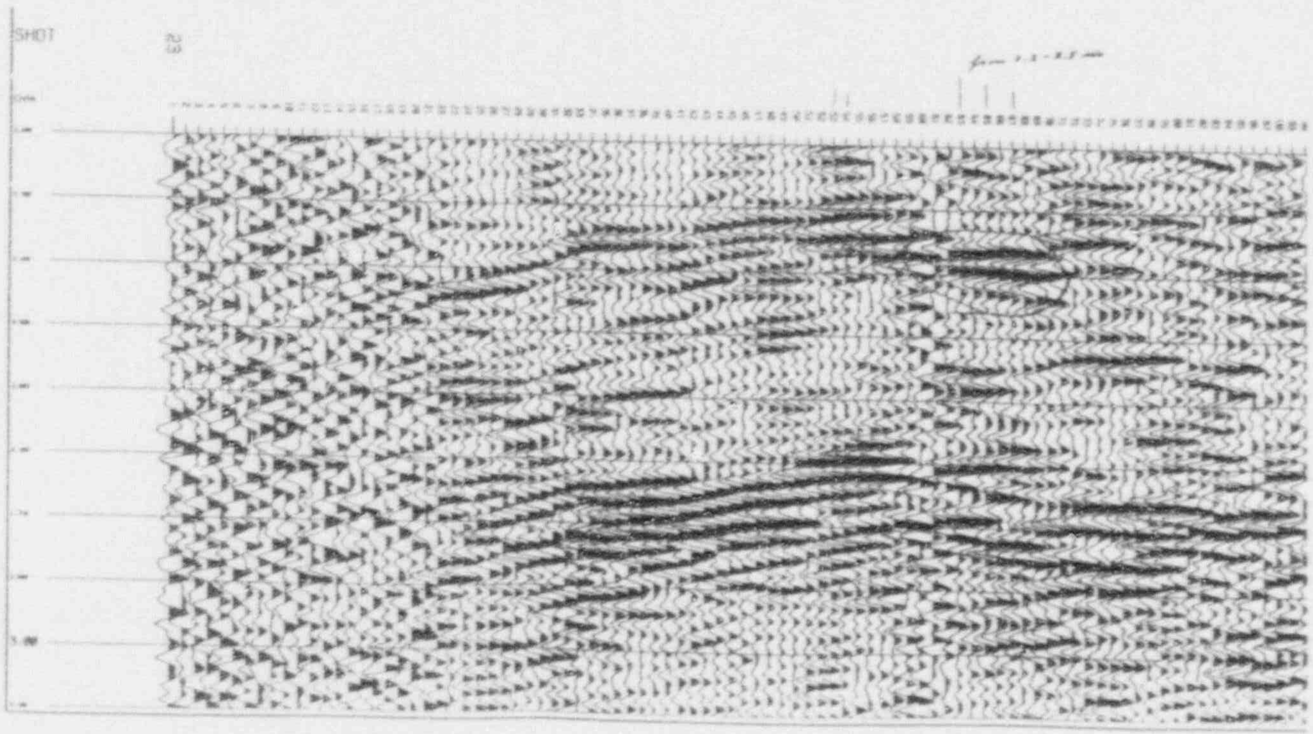

Figure 14. Sample shot record used in analysis: Line NRC2A1-4, shot 23, time 2.2-3.0 sec.

Figure15. Example of resampled portion of trace used in analysis: Resampled at 0.5 ms.

values greater than 2000 to around 100 (Tittman, 1981), the results obtained in central Virginia imply that the gross mineralogy at depth is free of hydrous phases. This interpretation does not preclude a dry fractured crust with water-filled fractures, however, because seismic wave lengths in the bandwidth 14-56 Hz do not see individual fractures. The results summarized above are not inconsistent with the hydroseismicity process. At this time, there is not enough data to prove that fluid-wall reactions associated with hydrolytic weakening of fault asperities eventually result in rock weakening, which leads to an earthquake, although that is the hydroseismicity hypothesis. If the crust is dry (high Q) to begin with, such a process might be more efficiently accommodated in a dry fractured crust, stressed close to failure, and in contact with meteoric water. This result of high Q estimated for the upper crust at the location of Line NRC2A1-4 in central Virginia might also offer some clue as to rock type at this location.

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Part C. - Seismic reflection data acquisition and interpretation in the central Virginia Seismic zone (Task 2, Subtask 2A-1-1, -2, -3, -4.).

By Lynn Glover, III, Cahit Çoruh, Alexander E. Gates, Stewart S. Farrar, Judith Patterson, J.K. Costain, and G.A. Bollinger

Introduction

The following profiles have been processed by the Automatic Line Drawing (ALD) process developed by Cahit Çoruh (Çoruh and others, 1988). ALD's emphasize relative reflectivity in the reflection data and they take the place of conventional line drawings for use in interpreting subsurface structure.

Reflections and zones of coherent reflection in these profiles caused by variations in impedance contrast may have several geologic origins, one of which is not likely for this area:

1) gas or fluid layers in metamorphic rocks - unlikely in large quantities
2) impedance contrast between geologic bodies of different composition -likely
3) impedance contrast between strongly foliated and/or mylonitic zones and less foliated host rock - likely
4) combinations of 2 and 3 above where rock layers of different compositions have been transposed into mylonitic zones of intense ductile shear - likely

All of the last three geologic origins are well known from surface studies in the Piedmont and Blue Ridge of the Appalachians, and therefore contribute to the production of zones of coherent reflections shown in the profiles. Velocity and Q studies have the potential to discriminate between some of the origins enumerated above, but ideal conditions for using these means are not common in the deformed and metamorphosed rocks of the Piedmont and Blue Ridge. Therefore there remains an inherent ambiguity in the exact interpretation of the reflectors except where they can be tied into surface rocks.

Most of the lines drawn on the ALD's appear to represent planes of discordance where reflectors or packages of reflectors are truncated, ie the Hylas mylonite/fault zone on I-64 and NRC - 10. In some [illegible]s these have been traced into known mylonite or fault zones at the surface. In other cases mafic-rich gneissic layers at the surface correspond to strong reflectors or zones of reflectors at depth, ie. mafic gneisses on NRC 10.

Data Acquisition and Processing

Eastern Piedmont segment of U.S. I-64 profile (Figure 16)

This profile is introduced first because it is part of a long and continuous one in which the nature of many reflections are known from surface extrapolation. This is important in relation to interpreting the reflections at depth in the shorter profiles carried out for this program adjacent to I-64 (Plate 1). This part of the profile is located over the

Figure 16. Eastern Piedmont segment of U.S. I-64 profile

Goochland Group just west of Richmond. It crosses, in dip section, the southernmost of three *en echelon* domes that are each cored by the State Farm Gneiss.

Geological interpretation: Surface geology shows the 1 Ga Maidens Gneiss, Sabot amphibolite and State Farm Gneiss of the Goochland nappes thrust over the Cambrian Chopawamsic volcanics by dextral transpression (Spottsylvania zone) parallel to the orogenic axis. These nappes include the Petersburg Granite and its Goochland host in the eastern part of the profile. Structure of the Maidens in the western part of the profile is moderate to steeply dipping, and very discordant to the subhorizontal attitude of the Chopawamsic volcanics as shown in the profile. Thus the Spottsylvania is a fault of major vertical and horizontal displacement located at the contact of the Goochland and Chopawamsics. The fault is recognized in the subsurface by the discordances it created between other reflectors, and locally by reflecting surfaces parallel to it.

The Hylas mylonite zone is another of these ductile dextral transpression zones shown in the profile that can be extended well into mid-crustal levels. Together these subparallel ductile fault zones form a family of faults that dip moderately into the subsurface and merge somewhat discordantly into the "lower laminated crust" below about 6 seconds two-way time.

Dashed lines are used to emphasize antiformal intervals of reflectors that are truncated by eastward dipping transpressional faults. A dome cored by the State Farm has been mapped at the surface and exhibits structure similar to that shown by the antiformal reflector intervals. The geometry of the antiforms and their relation to the transpressional faults indicates that they were formed in the transpressional process that created the faults.

Between 6.5 and 7 seconds horizontal reflectors predominate in the western half of the profile. At the western end of the profile a very reflective interval at 12 - 13.5 seconds ascends eastward to less than 10.5 seconds. The base of this interval corresponds to Moho (Pratt and others, 1988). Between the mid crustal and lower crustal reflective zones less well developed subhorizontal and gently to moderately east dipping reflectors occur (first generation reflectors). The gently east dipping reflectors merge asymptotically with the mid crustal horizontal zone and probably also with the lower crustal reflective zone. The more moderately east and west dipping reflectors (second generation reflectors) are superimposed on the mid to lower crustal reflectors just described and are clearly younger.

The geometry of the lower crustal first generation reflectors has no parallel in stratigraphic geologic frameworks, nor does it resemble a simple sill/dike relationship. It does however have similarities to C (shear band) and S (schistosity) structures in ductily deformed massive rocks. In this analogy the mid and lower crustal reflectors are the shear bands and compositional layering, and the gently dipping reflectors in between are schistosity and compositional layering. The production of schistosity and shear bands also tends to transpose the compositional layering into parallelism with the schistosity or shear bands. These observations are true for the metamorphic rocks now exposed at the surface, and they present a logical framework for interpreting the deeper structure as well.

Earthquake foci: These foci have been projected as much as 30 km along the structural strike into the plane of the section. In a few instances the foci are directly under or adjacent to the profiles. The distance can be seen from Figure 17. The foci have vertical error bars in the range of ±1 - 5 km which makes it impractical to try to relate them to a particular fault within the profile. It is a wonder of the data base that nearly all of the foci, when plotted at the mid point of their error range, fall on or very near faults in the section. Even more enigmatic is the observation that their first motion fault planes are dominantly high angle strike slip surfaces strongly discordant to the faults interpreted here.

NRC - 10 profile (Figure 18)

This profile is situated over the middle of three *en echelon* domes just west of Richmond (Plate 1) and just north of the I-64 profile. The seismic traverse (Plate 1, Figure 17) is a somewhat U-shaped one with the arms of the U oriented NE along the strike of the geology. All of the data has been projected (binned?) so that the profile is a dip section oriented NW.

Geological interpretation: Surface geology shows a low amplitude dome in the Goochland Group cored by the State Farm Gneiss. The surface structure is concordant with the reflection data in the first second or so of two-way time. Surface layers are mafic and felsic gneisses including conformable relict layers of granulite gneiss. Thus the reflections come from original compositional layering and from some granulite lenses conformable to the original layering that would be of high velocity because of their dehydrated, unretrogressed mineralogy. On the east side of the profile this gently dipping Goochland structure is broken by moderately east dipping mylonites of the Hylas zone. East dipping mylonite zones are clearly seen transecting the gently dipping reflections below stations 300 and 346. At stations 200 and just west of station 300 these ductile faults are found at the surface.Other parallel shear zones occur at 1.4 seconds below station 200, at 2.6 seconds below station 260, and elsewhere. Some late (Mesozoic?) extensional dip slip movement is suggested by the apparent offset of reflective intervals at 1 - 1.5 seconds below stations 150 and 200. Surface studies of the Hylas zone (Bobyarchick and Glover, 1979, Gates and Glover, 1989, and Glover, 1989) show that the Hylas and many other compressional mylonite zones in Virginia have experienced Mesozoic extensional reactivation.

Earthquake foci: Neither faults or other structures are well defined by the data below about 3 seconds. Earhtquake foci fall in an interval between about 4 and 5 seconds that is particularly nondescript.

NRC 2A-1-2 profile (Figure 19)

This is another dip section located 7 km NE of NRC - 10 (Plate 1). It spans a down-fold between the northern two of three domes.

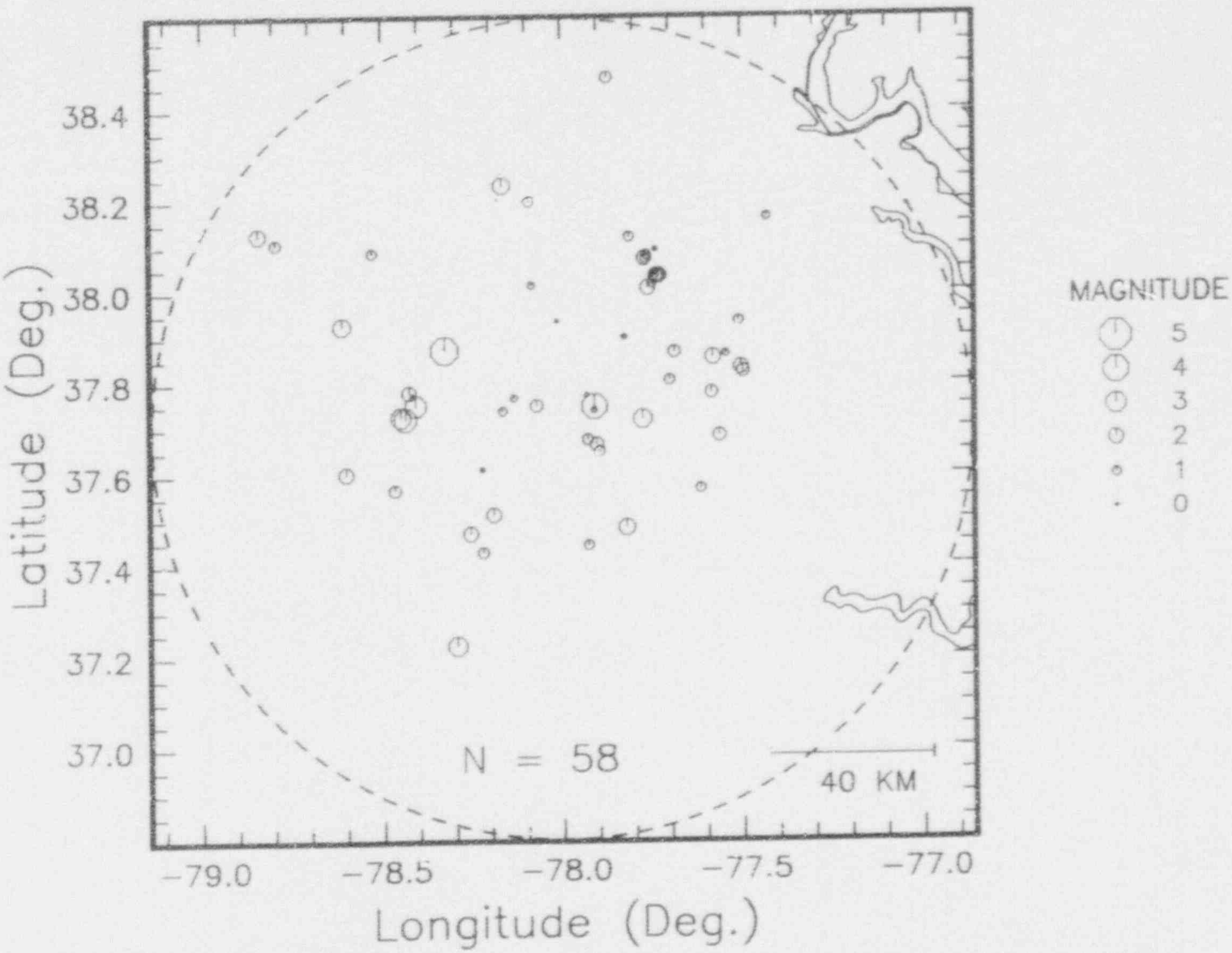

Figure 17 a. Seismicity in the central Virginia seismic zone

INSTRUMENTALLY LOCATED REGIONAL AND LOCAL EARTHQUAKES FOR VIRGINIA

Lab.-Reg.	Origin Time (UCT) Year Mo Dy	Hr/Mn:Sec	Hypocenter Location Lat-N	Long-W	Depth	Location Parameters NSTA	P/S	GAP	DMIN	RMS	SQD	Error Ellipse Proj. (ERH1,AZ1;ERH2; ERZ;Q)	Magnitude mb/Md/I	Srce
4 -CV	1974 06 18	15:53:	38-54.76	77-45.73	1.3					0.1		(0.5,360; 0.5; 3.6;)	/ 1.0/	2
7 -CV	1974 11 07	21:30:57.3	38-14.26	78-09.62	27.5	8	6/10	235	33	0.4	C|D	(4.1,+10; 1.8; 2.9;B)	3.4/ 2.7/5	1
8 CV	1975 04 12	13:30:34.1	37-49.26	77-41.98	7.3	12	12/11	324	14	0.1	B|D	(1.2,-14; 0.8; 4.1;B)	/ 1.2/	1
11 -CV	1975 08 15	13:42:	38-12.26	77-43.73	0.1					0.0		(0.1,360; 0.1; 1.0;)	/ 1.5/	2
12 -CV	1975 09 07	19:53:	38-11.53	77-43.70	0.6					0.0		(0.1,360; 0.1; 0.3;)	/ 2.1/	2
13 -CV	1975 09 18	18:56:	38-11.57	77-43.77	0.7					0.1		(0.2,360; 0.2; 1.5;)	/ 1.7/	2
15 -CV	1975 12 28	02:32:	38-18.72	77-52.24	16.1					0.1		(0.7,360; 0.7; 1.4;)	/ 1.4/	2
20 -CV	1976 07 19	13:58:	38-11.51	77-44.18	1.9					0.0		(0.1,360; 0.1; 2.6;)	/ 1.6/	2
22A -CV	1976 10 30	09:32:49.8	37-35.56	77-53.62	15.0	13	9/12	336	31	0.2	C|D	(3.0,-74; 1.5; 4.4;B)	/ 1.0/	1
23B -CV	1976 10 30	10:57:18.7	37-40.44	77-55.62	15.0	14	13/14	324	30	0.2	C|D	(2.2,-33; 1.4; 1.0;A)	/ 1.3/	1
26 -CV	1976 12 02	18:25:	38-05.71	77-25.72	5.2					0.1		(1.8,360; 1.8; 1.5;)	/ 0.8/	2
28B -CV	1977 02 27	20:05:35.5	37-55.51	78-36.51	5.5	6	5/ 6	153	34	0.3	D|C	(4.8,-13; 1.8; 3.9;C)	2.5/ /4	1
29 -CV	1977 03 06	01:28:	38-00.94	77-44.79	0.4					0.0		(0.2,360; 0.2; 1.0;)	/ 0.7/	2
30 -CV	1977 04 20	03:19:	38-12.13	77-44.41	1.2					0.1		(0.1,360; 0.1; 0.7;)	/ 0.8/	2
31 -CV	1977 04 24	02:51:	38-10.42	77-45.57	0.3					0.1		(0.2,360; 0.2; 0.4;)	/ 2.0/	2
41 -CV	1978 10 29	12:22:42.9	38- 2.92	78- 5.00	10.4	3	3/ 3	242	24	0.0	B|D	(0.7,-45; 0.1; 1.4;B)	/ 0.6/	1
42 -CV	1978 11 15	08:33:47.6	37-40.86	77-33.68	13.4	5	4/ 4	186	50	0.2	B|D	(0.9,-45; 0.8; 1.6;B)	/ 1.6/	1
43 -CV	1978 11 26	03:04:51.3	37-22.63	78-13.18	6.4	10	9/ 7	103	28	0.2	B|C	(0.6,-81; 0.3; 1.2;A)	1.3/ 2.2/	1
49 -CV	1980 04 26	13:59:54.8	37-42.09	77-34.94	5.2	5	9/ 9	204	35	0.3	B|D	(0.9,-50; 0.5; 1.6;B)	/ 1.6/	1
53 -NA	1980 08 04	10:13:32.7	38- 4.22	77-46.01	5.5	8	8/ 7	116	7	0.1	A|B	(0.4,-38; 0.3; 0.8;A)	/ 0.9/	1
57 -NA	1980 09 26	01:31:57.6	38- 4.29	77-46.14	3.8	7	7/ 5	116	4	0.2	C|B	(0.7,-33; 0.4; 2.7;C)	/ 1.9/	1
57A -NA	1980 09 26	05:04:15.7	38- 5.48	77-44.26	4.2	3	3/ 3	191	4	0.2	D|D	(3.8, 45; 1.8; 4.5;C)	/ 0.2/	1
59 -NA	1980 10 31	22:40:28.5	38- 7.16	77-48.68	2.4	4	3/ 3	166	6	0.1	C|C	(3.4,-48; 0.5; 4.3;C)	/ 1.1/	1
63B -CV	1981 01 21	16:29:58.1	37-44.31	78-24.86	7.5	7	6/ 3	174	24	0.1	B|C	(0.6, 43; 0.4; 2.0;B)	/ 0.2/	1
64A -CV	1981 02 11	13:44:16.4	37-43.17	78-24.09	6.3	14	14/ 9	78	29	0.2	B|C	(0.5,-82; 0.4; 1.4;B)	3.4/ 2.4/4	1
64B -CV	1981 02 11	13:50:31.5	37-44.97	78-24.10	10.0	12	12/ 7	115	26	0.2	B|C	(0.7, 29; 0.5; 1.1;A)	3.2/ /4	1
64C -CV	1981 02 11	13:51:38.7	37-43.46	78-26.85	8.5	9	8/ 6	128	29	0.1	B|C	(0.5, 26; 0.4; 1.2;A)	2.1/ 2.2/3	1
66 -CV	1981 04 09	07:12:54.4	37-26.83	77-49.27	0.7	11	8/10	129	43	0.3	B|C	(0.6, 88; 0.4; 1.5;B)	/ 2.3/	1
69 -CV	1981 04 16	13:49:20.5	37-36.65	78-13.30	13.5	3	3/ 3	161	23	0.1	D|D	(5.5,-65; 8.3; 2.4;D)	/ 0.2/	1
71 -CV	1981 07 30	11:58:48.5	38-11.47	78- 5.38	4.1	10	8/10	170	30	0.3	C|C	(1.4,-17; 0.6; 3.8;C)	/ 1.1/3	1
77 -CV	1982 01 13	13:16:25.0	37-14.93	78- 4.21	9.2	10	6/ 9	170	6	0.1	A|C	(0.4,-65; 0.3; 0.5;A)	/ 1.6/	1
78 -CV	1982 01 18	06:11:41.1	37-53.96	77-49.35	5.8	9	8/ 5	179	10	0.1	B|C	(0.9,-36; 0.4; 2.2;B)	/ 0.3/	1
81A -CV	1982 05 04	14:54:02.2	37-31.94	78-27.90	2.4	6	5/ 6	140	40	0.2	C|C	(1.8,-84; 0.5; 4.3;C)	/ 1.5/	1
82 -CV	1982 05 06	07:18:10.9	37-51.23	77-34.71	10.0	10	10/ 8	173	17	0.2	B|C	(0.5,-73; 0.4; 1.2;A)	/ 2.1/1	1
84 -CV	1982 06 16	18:40:58.7	38- 7.57	78-50.34	11.0	6	6/ 5	124	37	0.1	B|C	(0.5, 8; 0.3; 0.9;A)	/ 2.0/1	1
86 -CV	1982 06 25	23:03:47.0	37-49.93	77-30.10	13.1	9	9/ 9	166	23	0.1	A|C	(0.5,-22; 0.4; 0.9;A)	/ 1.9/	1
87 -CV	1982 09 20	12:15:32.0	37-49.56	77-29.77	10.1	10	10/10	168	29	0.2	B|C	(0.5,-82; 0.4; 0.9;A)	/ 1.6/	1
100 -CV	1983 08 10	12:29:34.2	37-46.72	78-25.46	11.3	8	8/ 6	119	23	0.3	C|C	(2.0,+28; 0.7; 2.8;C)	/ 1.8/	1
108 -CV	1984 04 12	23:46:30.6	37-56.79	78-00.70	07.4	4	4/ 4	217	13	0.0	B|D	(0.8,321; 0.1; 0.7;B)	/-0.7/	1
109 -CV	1984 05 25	11:35:35.0	38-26.37	78-41.46	38.4	5	5/ 5	328	33	0.1	C|D	(1.8,-11; 0.4; 2.8;C)	/ 1.5/	1
111 -CV	1984 08 17	15:05:46.0	37-52.30	78-19.53	08.6	13	13/ 3	105	17	0.2	B|B	(1.0,+15; 0.4; 1.7;B)	4.2/ 4.3/5	1
113 -CV	1984 10 31	08:57:40.7	37-56.11	77-30.41	14.7	9	8/ 9	202	17	0.2	B|D	(0.9,-27; 0.5; 0.9;A)	/ 1.1/	1
115 -CV	1984 12 02	12:29:34.1	37-24.32	77-56.57	4.5	6	3/ 6	258	42	0.2	C|D	(0.6,+60; 0.4; 4.1;C)	/ 1.1/	1
117 -CV	1985 04 22	18:22:16.0	37-36.13	78-35.91	4.5	8	8/ 7	111	44	0.2	C|D	(1.1,+90; 0.5; 4.1;C)	/ 2.2/	1
123A-CV	1985 09 02	04:34:03.9	37-43.47	78-07.97	8.2	3	2/ 3	353	4	0.1	D|D	(3.1,+31; 1.9; 2.9;C)	/ 0.4/	1
123B-CV	1985 09 02	04:58:00.5	37-44.31	78-09.83	10.0	5	2/ 5	356	6	0.2	C|D	(1.4,+21; 1.2; 2.1;B)	/ 0.3/	1
129A-CV	1986 07 20	19:24:04.2	38-05.16	78-31.49	12.4	3	3/ 3	292	13	0.1	B|D	(1.0,-20; 0.9; 1.0;B)	/ 1.1/	1
138 -CV	1987 04 11	03:30:20.5	37-39.91	77-54.16	6.7	8	8/ 8	136	23	0.2	B|C	(0.4,+84; 0.4; 1.4;B)	/ 1.7/	1
139 -CV	1987 04 14	03:09:00.5	37-51.71	77-32.57	11.7	6	3/ 6	256	24	0.3	C|D	(2.5,+44; 1.3; 4.2;C)	/ 0.8/	1
148 -CV	1988 08 27	16:52:29.5	37-43.03	77-46.50	14.2	7	6/ 7	158	31	0.3	B|C	(0.9,+62; 0.7; 1.4;C)	3.1/ 2.3/3	1
152 -CV	1989 05 06	09:49:26.2	37-17.66	78-17.59	8.8	8	8/ 6	137	6	0.2	B|C	(0.4,+26; 0.8; 1.2;A)	2.8/ 2.5/3	1
157 -CV	1990 06 17	11:23:58.3	37-28.12	78-15.31	1.1	10	10/10	87	31	0.2	B|C	(0.6,-78; 0.3; 1.2;A)	/ 2.0/	1
161 -CV	1990 11 06	08:00:51.3	37-35.07	77-36.86	12.5	6	6/ 6	188	50	0.2	B|D	(0.3,+79; 0.5; 1.1;B)	/ 1.1/	1
162 -CV	1990 12 14	08:32:15.0	37-51.96	77-41.17	15.6	6	6/ 6	268	17	0.3	C|D	(1.5,-54; 1.4; 2.2;B)	/ 1.5/	1
164 -CV	1991 01 11	21:01:59.0	37-30.42	78-11.41	6.6	8	8/ 8	132	32	0.2	B|C	(0.7,+90; 0.4; 2.0;B)	/ 2.0/	1
166 -CV	1991 03 15	06:56:08.3	37-44.78	77-54.53	15.5	7	7/ 6	116	10	0.2	B|B	(0.8,+66; 0.5; 2.5;B)	3.8/ 3.3/	1
167 -CV	1991 03 15	07:05:54.3	37-44.29	77-54.55	10.4	6	5/ 6	196	19	0.3	C|D	(1.2,-6; 0.8; 4.5;C)	/ 0.6/	1
168 -CV	1991 03 20	21:09:35.9	37-46.32	77-55.63	14.7	5	4/ 3	257	14	0.3	D|D	(4.6,-28; 1.6; 2.7;C)	/ 0.1/	1

There are 58 earthquakes in this list.

Abbreviations for Regions:
CV: Central Virginia Seismic Zone,
NA: North Anna, Virginia, area.

: Events that have been relocated for special studies. Calculated using either: a new technique (JHD, JED, HYPOELLIPSE, etc.) or a different velocity model.

Sources and/or References
1) VTSO records, NRC reports, SEUSSN Bulletins, or G. A. Bollinger personal files.
2) Dames and Moore, 1977, "A Seismic Monitoring Program At The North Anna Site In Central Virginia - January 24, 1974 through August 1, 1977. Submitted to VEPCO, 1977.

Figure 17b. Instrumentally located earthquakes for central Virginia

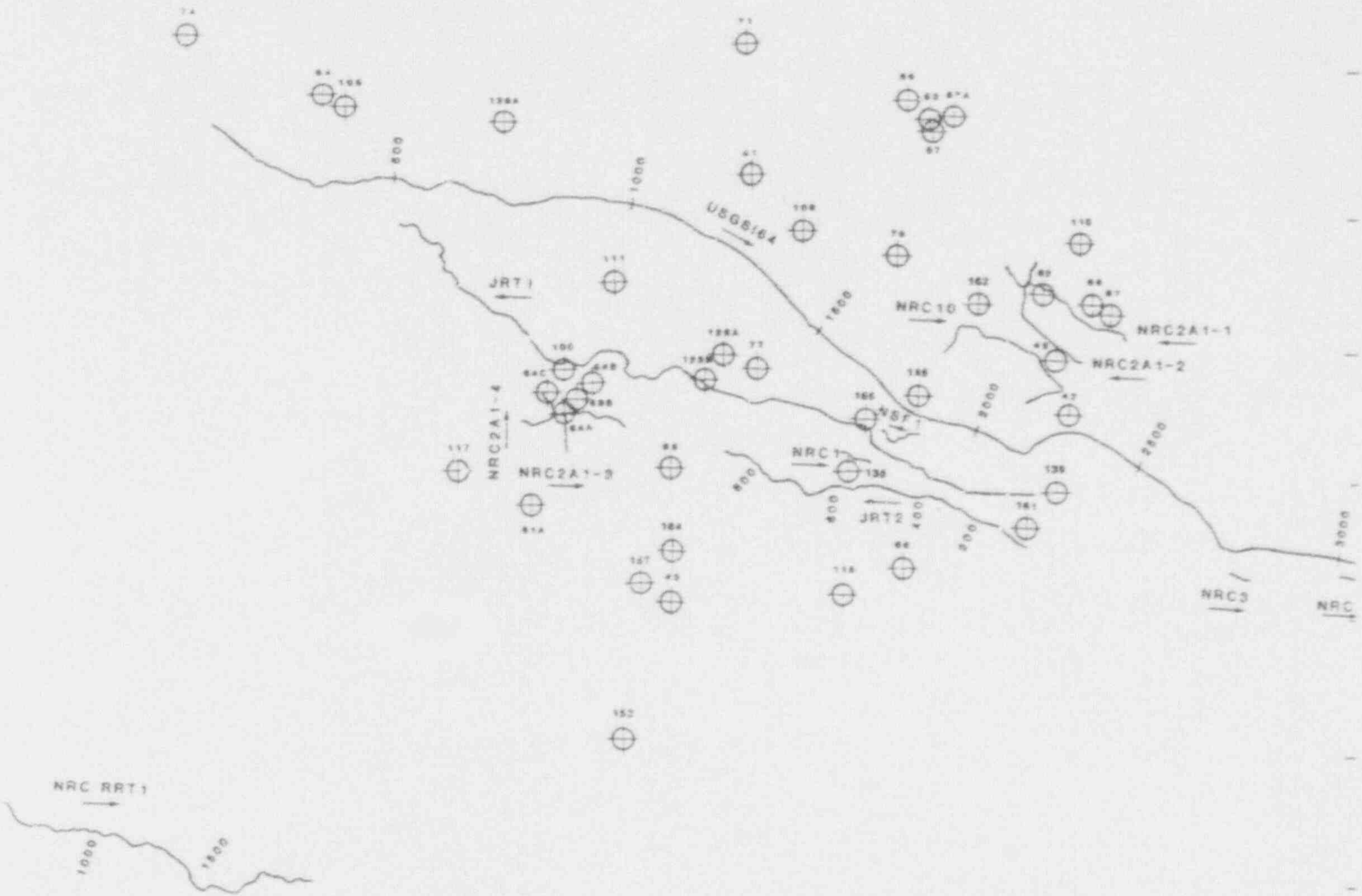

Figure 17c. Earthquake locations relative to vibroseis profiles.

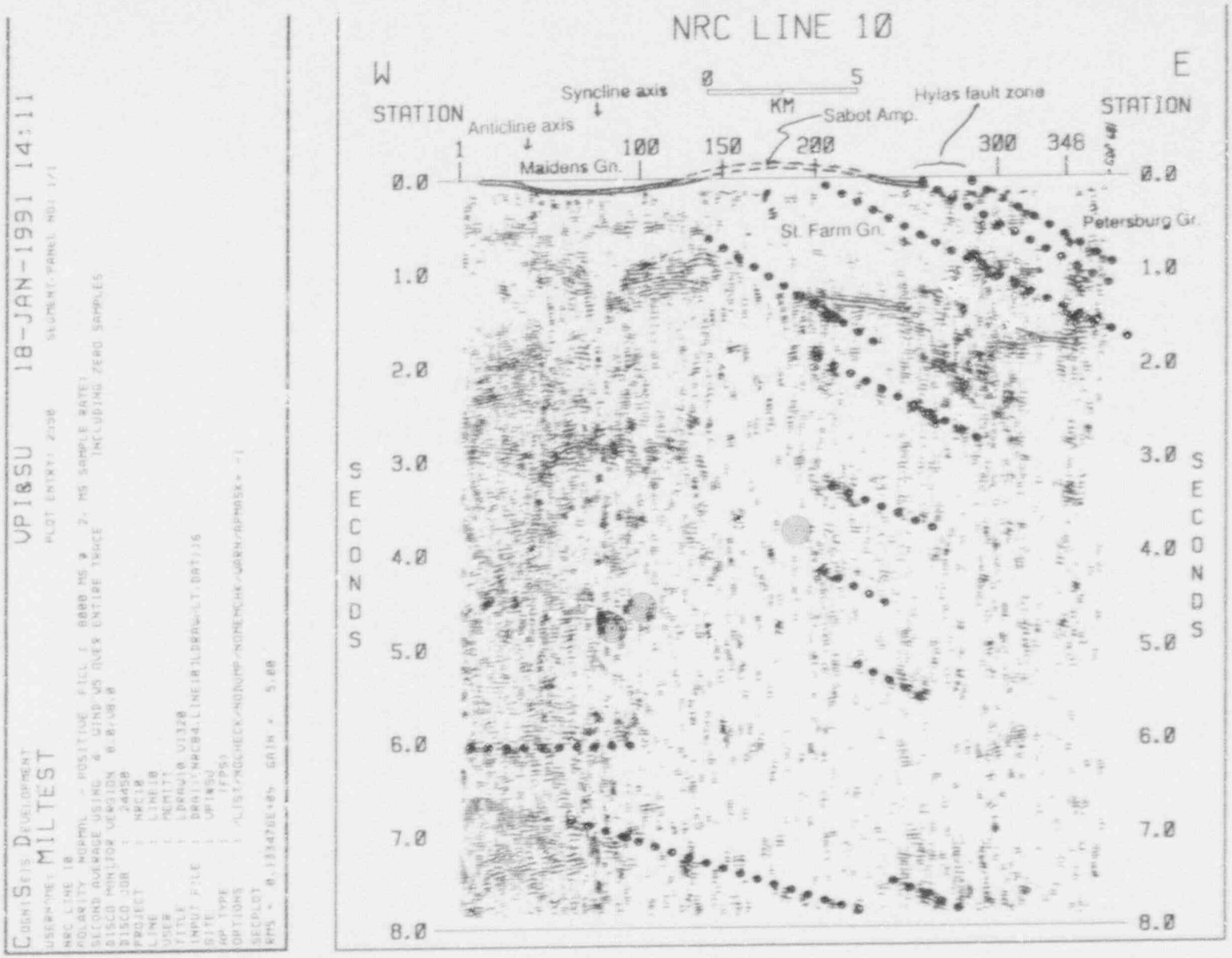

Figure 18. NRC line 10 profile

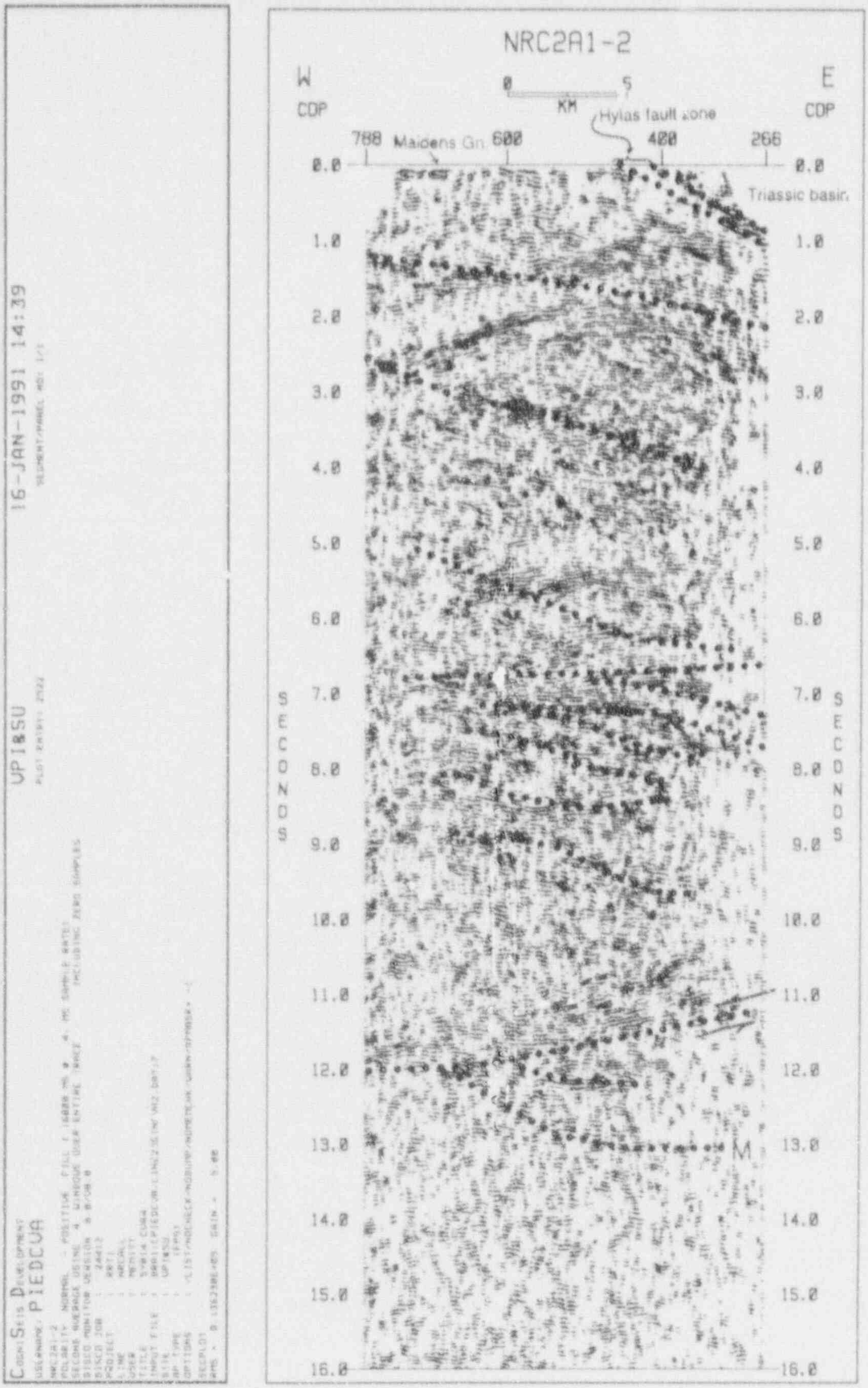

Figure 19. NRC 2A1-2 profile

Geological interpretation: At the surface nearly horizontal Maidens Gneiss of the Goochland Group occupies the western 2/3 of the profile. On the east the Maidens is truncated by the east dipping Hylas zone. Moderately dipping reflectors of the fault zone are well displayed along the subsurface projection of the Hylas. During Early Mesozoic extension the Hylas was reactivated as a normal fault (Bobyarchick and Glover, 1979) and gently dipping Triassic sandstones now occur east of the Hylas border fault. Several west dipping reflectors (stratigraphy) and east dipping faults can be seen in the profile. At about 7 seconds the mid crustal reflecting zone is quite prominent. Below it are some moderately east dipping trends that merge upward into the mid crustal zone and downward into the lower crustal zone and M discontinuity without disrupting them. The lower laminated crust shows structure between 11 and 13 sec. One interpretation is that the M discontinuity slopes westward between 11 and 12 seconds. However, additional laminated structure occurs between 11.5 and 13 sec in the east central part of the profile. An alternative interpretation would be that this is not the M discontinuity between 11 and 13 sec. but is a ductile extensional fault as shown on the profile.

Earthquake foci: Earthquake foci project into the profile between 3 and 3.3 sec.. There is little obvious structure in the profile at the focal positions but they would be very close to the fault in I-64 as shown projected into this profile. The west dipping reflector between 2 and 3 sec. is also probably the same reflector at that position in I-64.

NRC 2A-1-1 profile (Figure 20)

This is a dip section located about 4 km NE of NRC 2A-1-2 (Plate 1). NRC 2A-1-1 crosses the southern end of the northernmost dome on the ma

Geological interpretation: This profile is very similar to NRC 2A-1-2. The profile crosses the Maidens, Sabot and State Farm gneisses forming a dome which is truncated by the Hylas mylonite and fault zone on the east.. West dipping reflectors that end at the east dipping Hylas zone in this profile may be Triassic sedimentary layers or basement gneiss. The "brightness" of these reflectors suggests that they are part of the metamorphic basement. Weakly expressed discordances near 4 to 5 sec. suggest east dipping faults offsetting a west dipping reflector. The west dipping reflector just below 2 sec is probably the same as that in I-64 and 2A-1-2 at a similar time/depth. The mid crustal, 6 to 8 sec. reflective zone is well developed. Between 8 and 10 sec. a few east dipping discordances appear in a field of otherwise faint horizontal reflectors(?). The lower crustal reflective zone dips west from 11 to 12.5 sec along a possible fault. This truncates slightly east dipping reflectors which extend down to over 13 seconds in the profile. Compare with comments on deep structure in NRC 2A-1-2 above.

Earthquake foci: Three foci projected into the profile at about 3 to 4 sec. plot in the zone of east dipping faults.

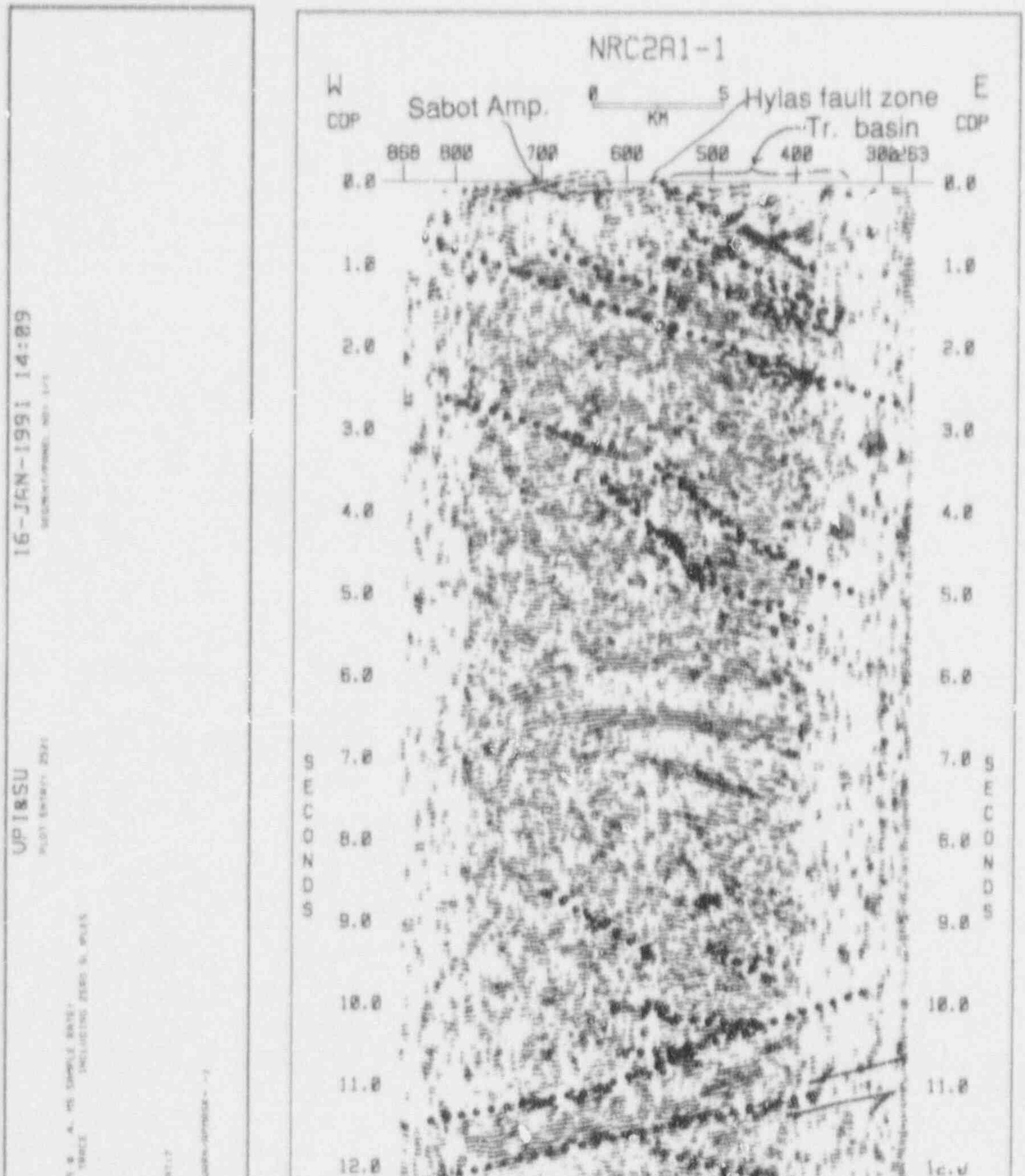

Figure 20. NRC 2A1-1 profile

NRC 2A-1-3 profile (Figure 21)

This is an oblique dip section located in the west central Piedmont along the James River (Plate 1).

Geological interpretation: Surface data shows that this profile crosses the Shores tectonic melange, Hardware olistrostrome (sedimentary melange), and ends on the west edge of the Arvonia Fm (Plate 1). An east dipping fault mapped between the Shores and Hardware has some expression in the seismic profile as a discordance at about 1 sec. East dipping mafic rocks mapped at the surface also are expressed as reflectors down dip in the profile From about 1 to 4 sec a number of subhorizontal to gently arched reflectors are broken by east dipping discordant surfaces that appear to be faults. Probable formations/lithologies are shown on the profile by comparison with the I-64 profile (Glover, 1989).

In this profile reflectors continue down into the middle crust to about 7 or 8 sec. There is no sharp break where one would expect to pass from layered supracrustal rocks into Grenville basement circa 3 sec. as occurs just west of here under the Blue Ridge (and throughout its length in the central and southern Appalachians). Comparison with the I-64 profile of Pratt and others (1988) and Glover (1989) indicates that NRC 2A-1-3 is located in the zone of transition between more highly reflective middle and lower crust east of the Blue Ridge and poorly reflective crust under the Blue Ridge. In progressing westward from the eastern Piedmont, reflectivity of the crust diminishes downward and westward as shown in the I-64 profile. This probably represents diminishing effects of Paleozoic and Mesozoic deformation on reletively homogeneous Grenville crust.

Although the lower crust is poorly reflective, reflections at 15 to 16 sec. suggest the depth of the M discontinuity in this profile.

Earthquake foci: Three foci projected from near the profile plot between about 2 and 3.5 sec in association with the Catoctin metabasalts and within the zone of faulting.

NRC 2A-1-4 profile (Figure 22)

This profile has the north half crossing structure in dip section and the south half nearly north-south and oblique to the dip.

Geological interpretation: From north to south the profile crosses the Hardware, Shores and Diana Mills Gabbro. A fault is possibly indicated in the profile between the Hardware and Shores as shown. The exact position of this fault at the surface is poorly controlled, but on the basis of regional information it must be nearby. The Diana Mills appears to be a thin tabular body because reflections in the range of 0.2 to 1 sec. or more pass unbroken below its surface contacts. Correlations with the Evington Group and Catoctin are made with the I-64 profile as in NRC 2A-1-3 above.

As in 2A-1-3 above, the upper crust is reflective down to about 8 sec.. Below

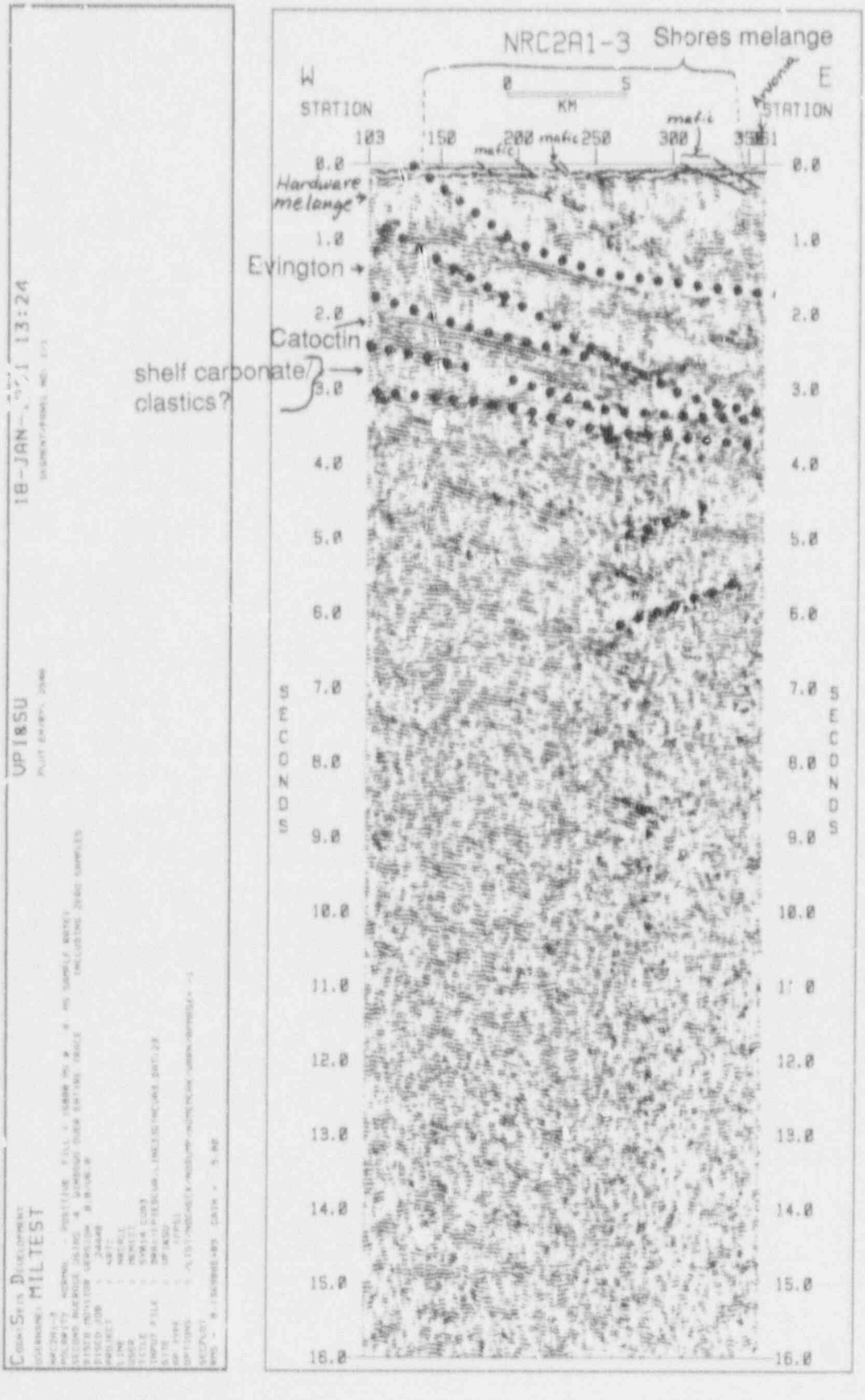

Figure 21. NRC 2A1-3 profile

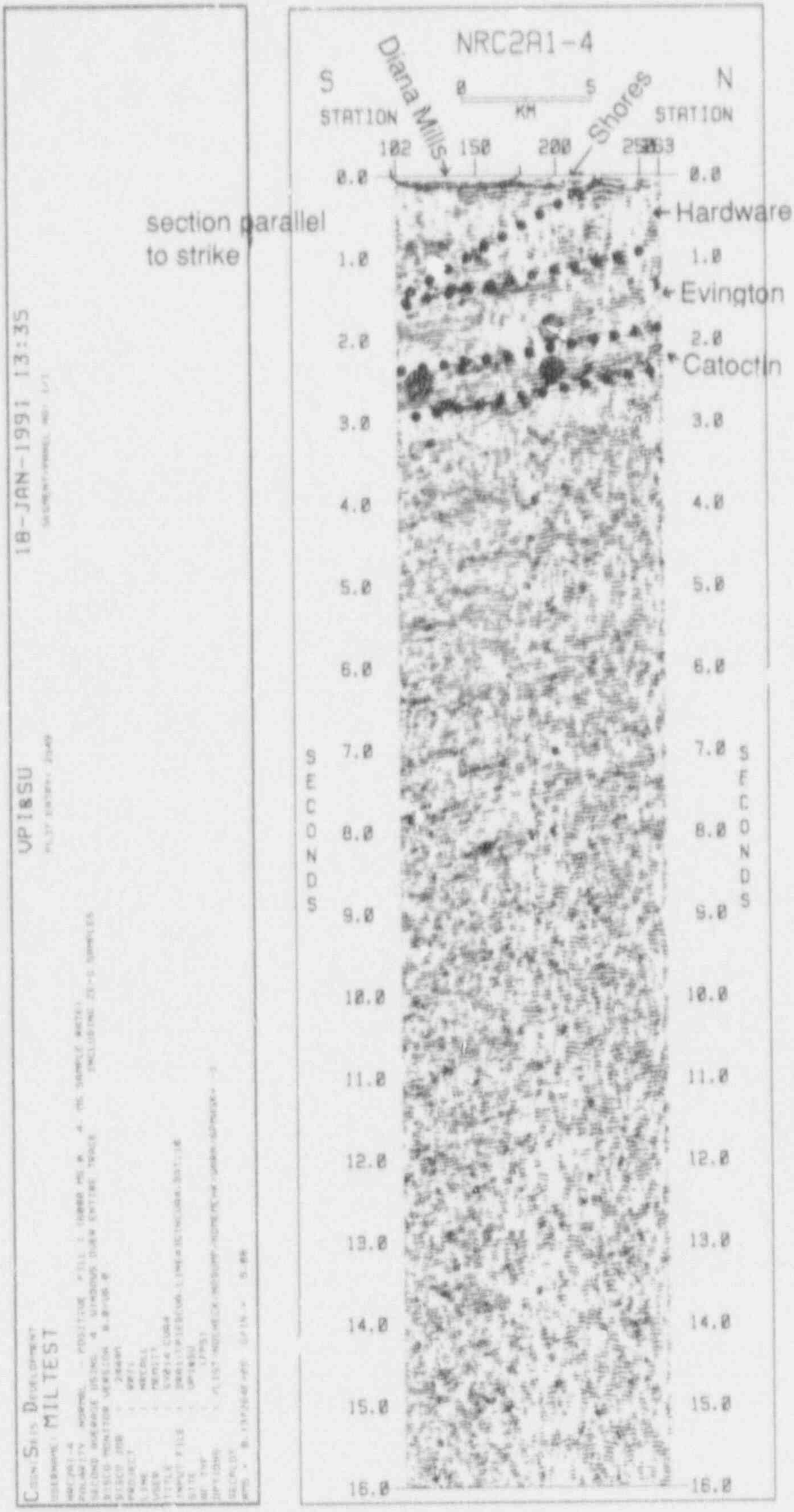

Figure 22. NRC 2A1-4 profile

that it is poorly reflective and the M discontinuity is not apparent in this profile.

Earthquake foci: The three foci projected into this profile plot within or adjacent to the Catoctin metabasalt.

References

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Çoruh, Cahit, Bollinger, G.A., Costain, J.K., 1988, Seismogenic structures in the central Virginia seismic zone: Geology, v. 16, p. 748-751.Gates, A.E., Glover, L., III, 1989, Alleghanian tectono-thermal evolution of the dextral transcurrent Hylas zone, Virginia Piedmont, U.S.A.: J. of Structural Geology, v. 11, p. 407-419.

Glover, Lynn, III, 1989, Tectonics of the Virginia Blue Ridge and Piedmont; *in* Glover, L., III, Evans, N.H., Patterson, J.G.,and Brown, W.R.(eds), "Tectonics of the Virginia Blue Ridge and Piedmont: Field Trip T363, 28th International Geological Congress.

Pratt, T.L., Costain, J.K., Coruh, C., and Glover, Lynn, III, (1988), A geophysical study of the Earth's crust in central Virginia with implications for lower crustal reflections and Appalachian structure: Journal of Geophysical Research, v. 93B, p. 6649-6674.

Part D. - Seismic reflection data acquisition and interpretation in the Roanoke River traverse in aseismic south-central Virginia (Task 2, Subtask 2A-3): Comparison with the seismically active James River geologic framework.

By Lynn Glover, III, Çahit Coruh, Alexander E. Gates, Wang, Ping, Judith Patterson, J.K. Costain and Gilbert A. Bollinger.

Introduction

The automatic line drawing process was used to process these profiles, and the introductory remarks to part C also apply to this part.

Data Acquisition and Processing

Roanoke River traverse 2A-3 (Figure 23a, b)

The Roanoke River traverse (Plate 1) is a dip section through the Blue Ridge and Piedmont in south central Virginia. The purpose of the traverse is to compare an aseismic corridor (the Roanoke River) with a seismically active corridor (the James River) to see whether any differences that exist could be related to the localization of seismicity. Only the central segment of the Roanoke River traverse is controlled by seismic reflection data (Plate 1, Figure 23a). Seventy percent of the geology of the Roanoke River Traverse shown in Figure 23a was mapped for this project. The seismic data was also acquired for this project.

The western boundary of the Evington Group (Plate 1; Figure 22a) separates the Blue Ridge Province on the west from the Piedmont Province on the east. The traverse thus crosses the Blue Ridge in its entirety and about 60% of the Piedmont (Plate 1). Figures 23a and 23b show the surface geology and rock type along the corridor.

Geological interpretation: In the northwestern part of the corridor (Figure 23a) metamorphosed Precambrian Grenville basement has been thrust over the unmetamorphosed Cambrian Rome and Shady formations of the Valley and Ridge. The latter formations are exposed in the Goose Creek window. It is obvious from the sinuous trace of the Blue Ridge thrust framing the window that the fault surface dips very gently toward the southeast.

Volcanics and sandstone of the Lynchburg Group (Wang, in progress) overlie the basement on the east and dip monoclinally steeply to the southeast (Figure 23a). The seismic profile (Figures 24 and 25) show subhorizontal to gently east-dipping reflections between stations 900* and about 650 at about 0.3 sec. two-way time below the Lynchburg. These are strongly discordant to the steeply dipping Lynchburg strata and imply a subhorizontal thrust fault separating the two units as shown in Figures 24, 25. The rocks from .5 to 2.5 sec. below the Blue Ridge show strong impedance contrasts and appear to be an imbricate stack of the carbonates and clastics that crop out in the Valley

Figure 23 a . Geology of the Roanoke River traverse

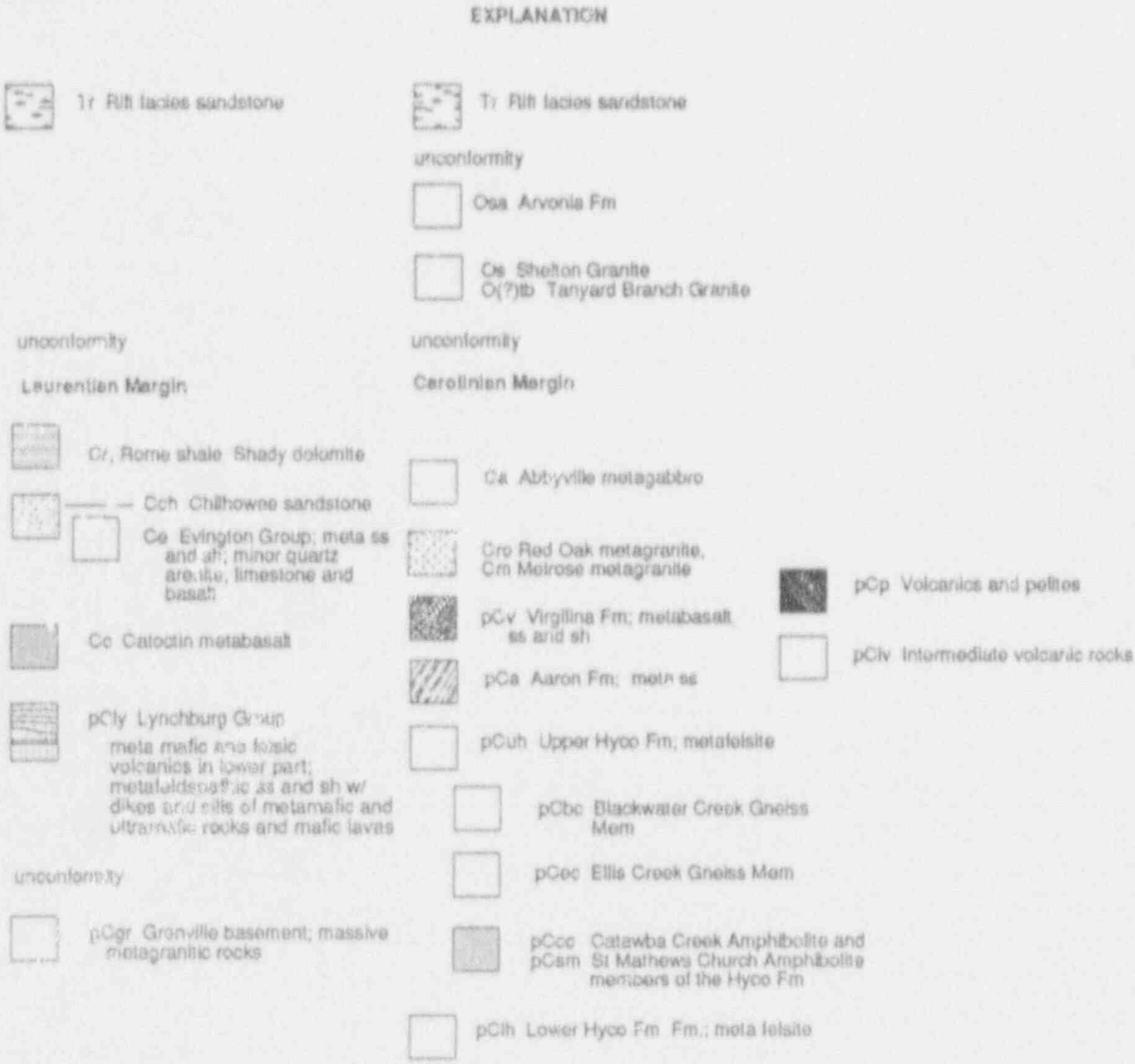


Figure 23b. - Explanation of geology of the Roanoke River traverse

and Ridge A zone of continuous horizontal reflectors at about 2.3 to 2.5 sec is interpreted to be a basal thrust zone or, less likely, stratigraphic layering at the base of the Paleozoic sequence. In Grenville basement below the Blue Ridge rhomboid packages of reflectors appear between 3.5 and 7 sec. The gross geometry of these reflectors suggests that they are not stratigraphic. Rather, they resemble anastomosing deformational zones of a ductile nature similar to those that form in relatively massive granitoid rocks at all scales (Kligfield and Crespi, 1984). The M discontinuity is not obvious if present in the western part of the profile.

From station 650 to 1680 a metamorphic, domed and faulted sequence of Lynchburg, Catoctin basalt and Evington group clastic rocks crops out(Figures 24, 25). These are deep water rift-related rocks of late Precambrian and Cambrian age that formed near the rifted margin of the Laurentian continent. They traveled with the Blue Ridge as they were thrust westward over the Laurentian shallow platform carbonate/clastic sequence during Paleozoic orogenies. The seismic profile below stations 650 - 1680 down to about 3 sec. two-way time shows numerous concave upward packages of reflectors separated by SE-dipping discontinuities considered here to be faults (Figures 24, 25). Discontinuities that intersect the surface are coincident with faults mapped at the surface near stations 800, 970 and 1000. The fault at 650 has not been found at the surface. The fault at 1600 has been seen outside of the corridor but its position on the profile is a projection along strike. The basal thrust zone at about 3.3 sec. is well developed . Between 3.3 sec and approximately 1 sec two-way time ,fault-imbricated carbonate and clastic platform rocks probably occur. This is supported by relatively slower interval velocities computed from stacking velocities near station 900 (Li and others, 1990). The interval velocity determined is about 4.2 - 4.6 km/s for the interval between 1.2 and 3 s. This interval velocity is 1 to 2 s slower than velocities expected for crystalline rocks. Grenville basement below 3.3 sec may be obscurely layered in a subhorizontal orientation or it may be massive. Minor SE-dipping discordances occur. The M discontinuity appears at 12.5 sec below stations 1300-1400 and persists eastward to the end of the Profile.

From station 1680 to the end of the profile at station 2393, surface outcrops include the sandstones of the Danville Triassic basin and the Cambrian Melrose granite (Figure 24). The Triassic basin between stations 1680 and ca.1870 is not well imaged in the profile. The road network dictated that the line turn northward near station 2100 before going east again where it just crosses the Arvonia Formation and some Carolinian volcanics at the end of the traverse. In this segment of the profile there is little information above 2 sec. perhaps in part because it is mostly massive granite. Below 2 sec. it is clear that the carbonate/clastic rocks continue down to as far a 3.7 sec. In the Grenville basement, below 4 sec horizontal reflecting packages become common. The geometry of these reflections suggests that pure shear, in extending the crust during early Mesozoic rifting, has aligned inhomogeneities in the crust so that they are mostly parallel and subhorizontal. An alternative explanation for some of these horizontal reflectors is that they are gabbroic sills injected during Mesozoic extension.

East of the seismic profile, from point "D" to point "E" (Figure 24), the traverse was completed with surface data gathered, and mapping conducted, during this project (Baird, 1989). The rocks comprise part of the Piedmont Charlotte belt and are mostly

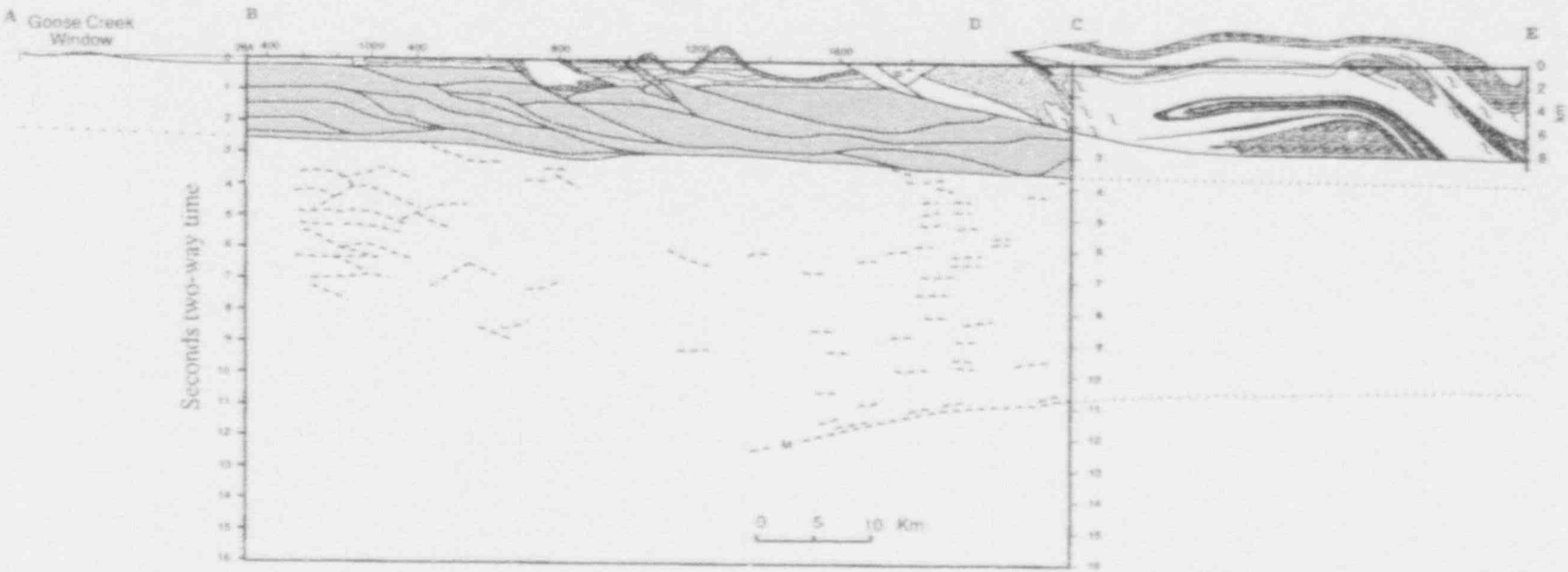


Figure 24 . Geological interpretation of Roanoke River traverse Geologic symbols same as Figure 23.

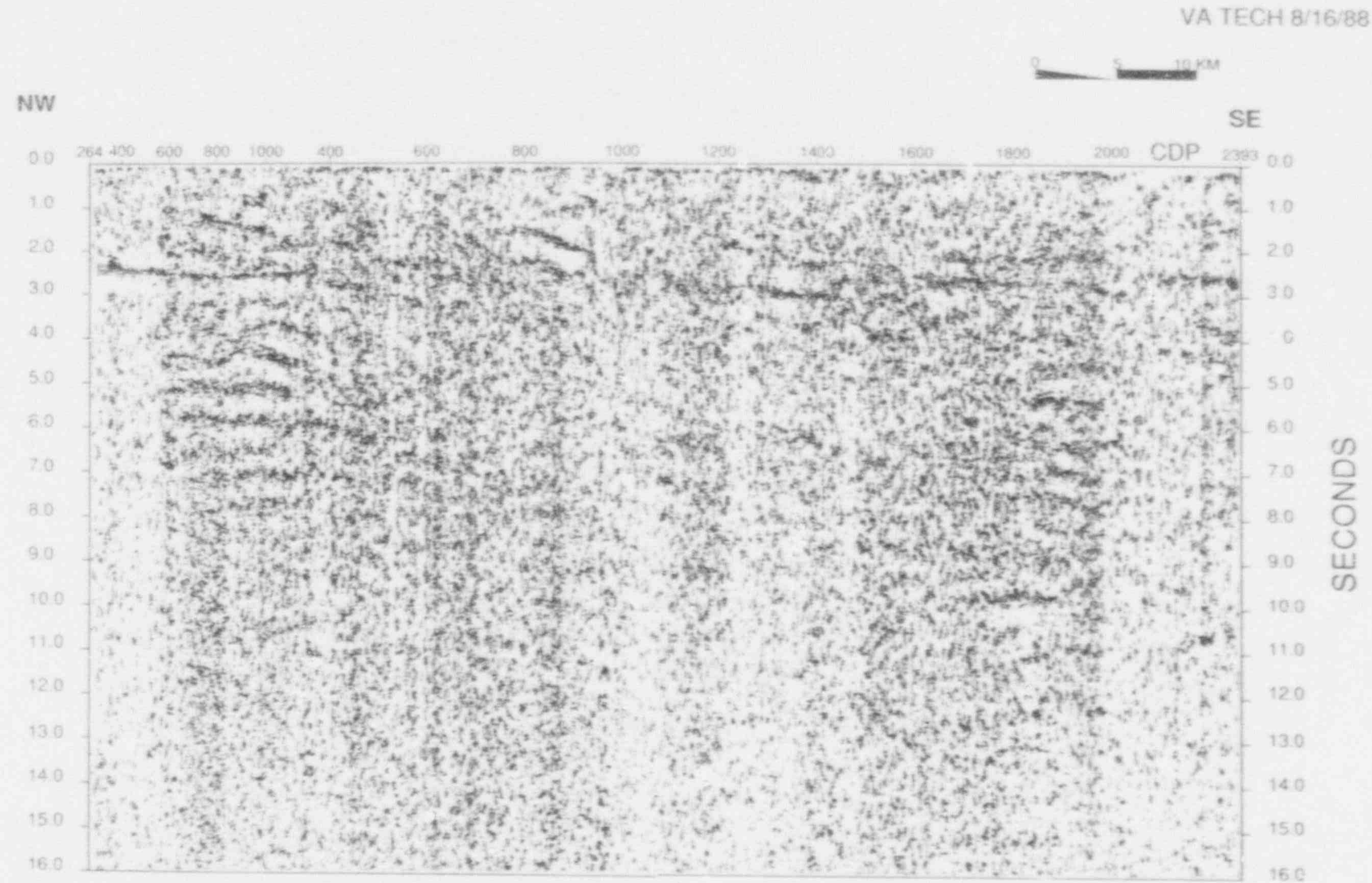


Figure 25 a . Roanoke River traverse vibroseis profile

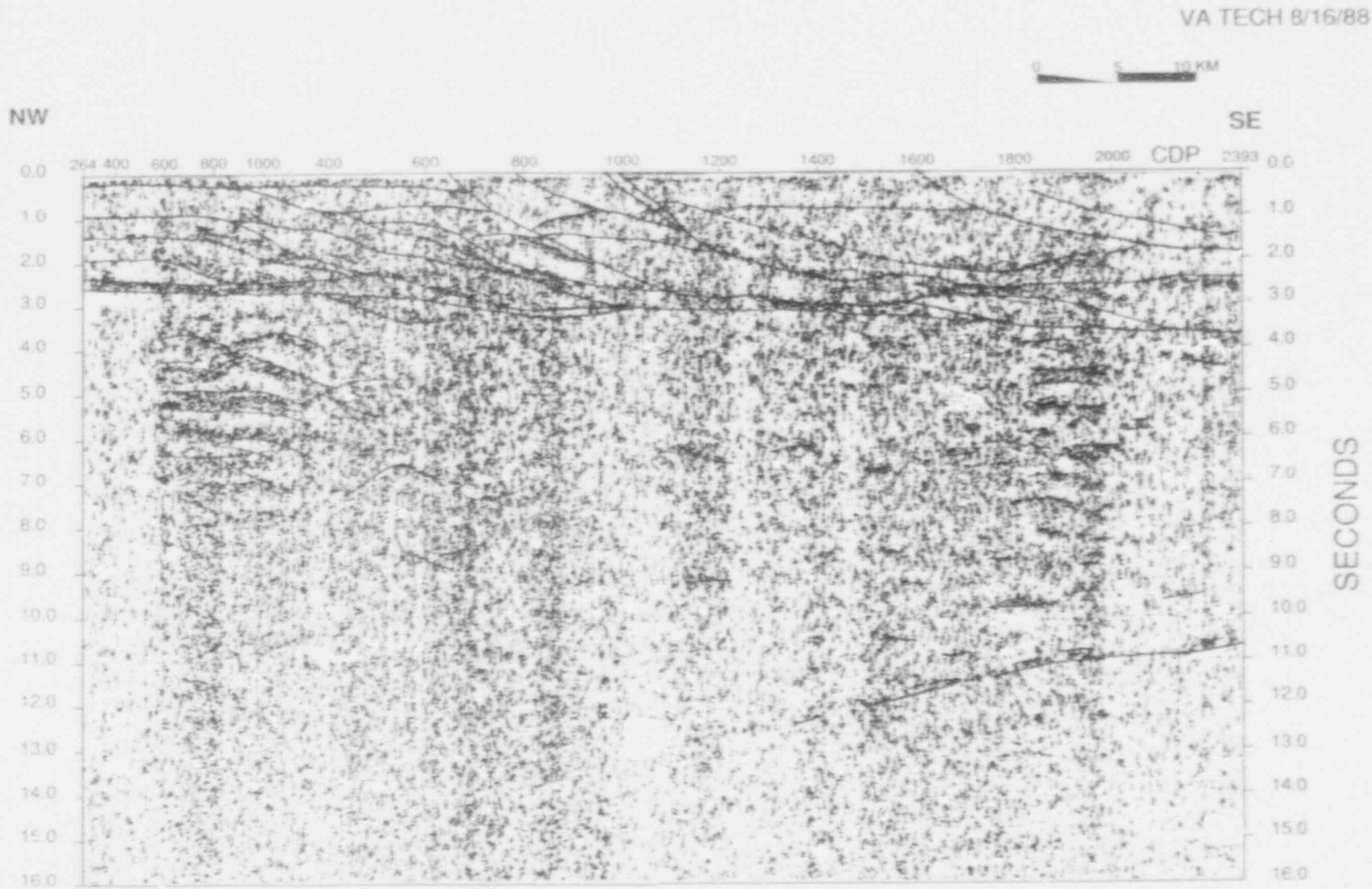


Figure 25b. Roanoke River traverse vibroseis profile; faults and deformation zones

amphibolite facies felsic and mafic metavolcanics of magmatic arc affinity. Most of the sequence is late Precambrian, and a minor amount may be Cambrian. Structurally the sequence is folded into a large recumbent fold nappe that formed mostly at the time of collision with the Laurentian margin during the Late Cambrian (Part A above). Therefore the Taconic suture occurs between these magmatic arc rocks and the rift stage continental margin rocks of Laurentia to the west (Figures 23, 24). The Melrose granite intruded the suture during Late Cambrian time, but the suture was reactivated as a mylonite zone that cut the Melrose during the late Paleozoic (Gates and others, 1986).

Comparison of Roanoke River and James River traverses

James River: Several interpretations of the I-64 profile along the James River are now in existence (Figures 26, 27, 28, 29). All recognize an arch-like structure in the Piedmont culminating under the Goochland nappe, but give it differing interpretations. Glover and others (1987, 1988) favored Mesozoic crustal extension as an explanation for the arch-like geometry of the Piedmont crust and subjacent M discontinuity. Pratt and others (1988) considered the arch to be related to Alleghanian dextral transpression. This theme was elaborated on by Gates and others (1988). Coruh and others (1988) and Costain and others (in Press) describe the arch-like structure as an antiform with a roof (limbs B and E in Figure28) and a floor (C in Figure 28), and attribute it to a combination of compressional and extensional tectonics. Their compressional stage was envisioned as dextral transpression producing a strike slip megaduplex between the Brevard zone on the west and the eastern Piedmont fault zone on the east. The eastern fault boundary was thought to be vertical. Mesozoic extension then allowed the western flank to slump and dip westward. Mesozoic dike swarms were thought to have invaded these vertical fault zones.

The Glover (1989) interpretation of I-64 shows extension, relatively minor dip slip offset on the east dipping fault, below the Hardware melange (Figure 29). This version is based on a manually produced line drawing of the seismic profile and this line drawing is reproduced with the geologic interpretation in Figure 29. The Çoruh and others' (1988) automatic line drawing version of I-64 produces a more detailed and objective drawing of the reflectors and using this in conjunction with the surface geology a new interpretation is given in Figure 30.

The interpretation in Figure 30 shows backslipping on Paleozoic thrusts during Mesozoic extension. Many mylonites in the area are known to have been reactivated in extension (Glover and others, 1980; Gates and Glover, 1989), not only those bordering Mesozoic basins, but many others as well. Reflectors within the backslipped block below the Goochland nappe and Chopawamsic volcanics appear to be rotated counter-clockwise to lower dip angles than one finds on either side. This counterclockwise rotation is also expressed in the westward dip of the melange (suture) below the Chopawamsics. Similar structure is known off the NE coast of Scotland where SE dipping Paleozoic thrusts in metamorphic terrane were reactivated during the Devonian forming inversion structures, graben filled with sandstone (Coward, M.P. and others, 1989).

In Figure 30 the suture as well as the reflectors within the block appear to have

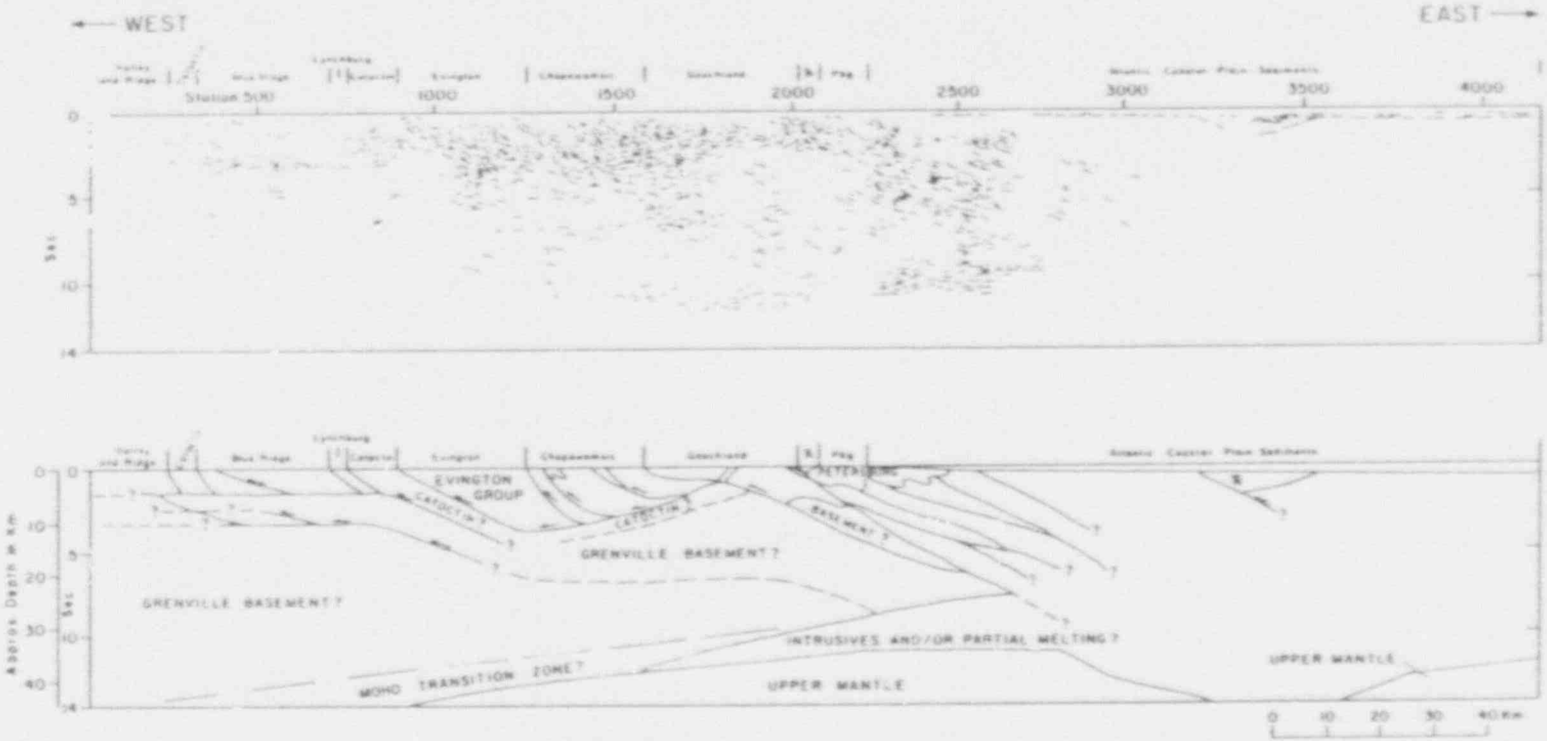
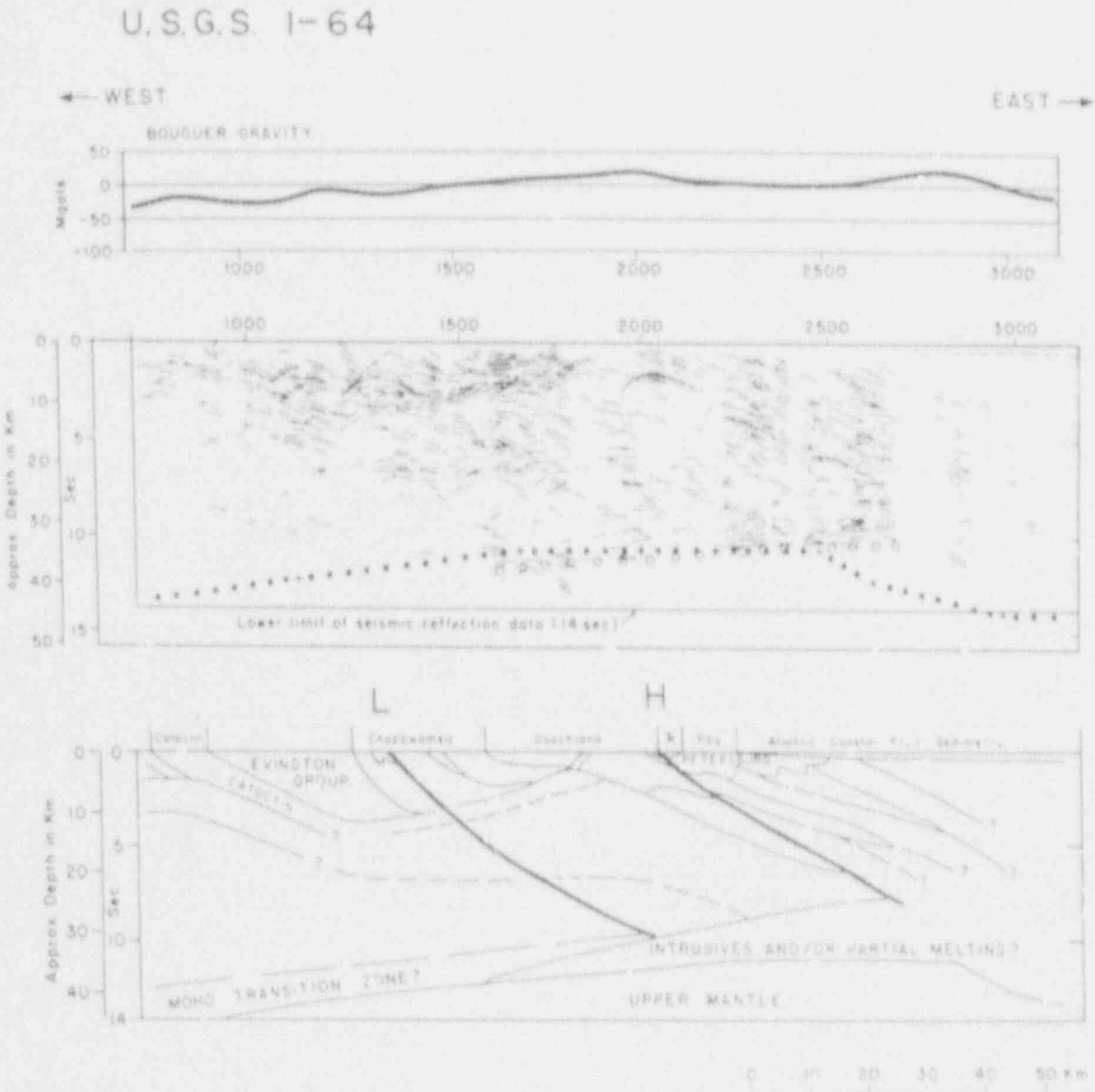


Figure 26. Pratt and others (1988) interpretation of U.S.G.S. Line I-64



Fgure 27. Gates and others (1988) interpretation of U.S.G.S. line I-64

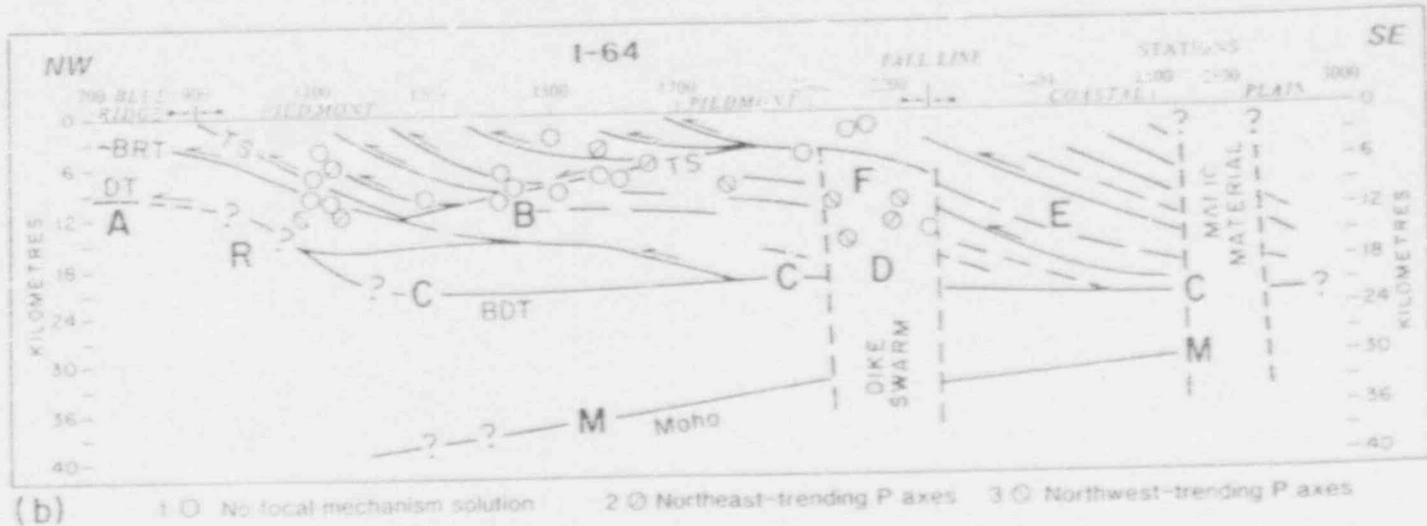


Figure 28. Coruh and others (1988) U.S.G.S. line I-64 interpretation

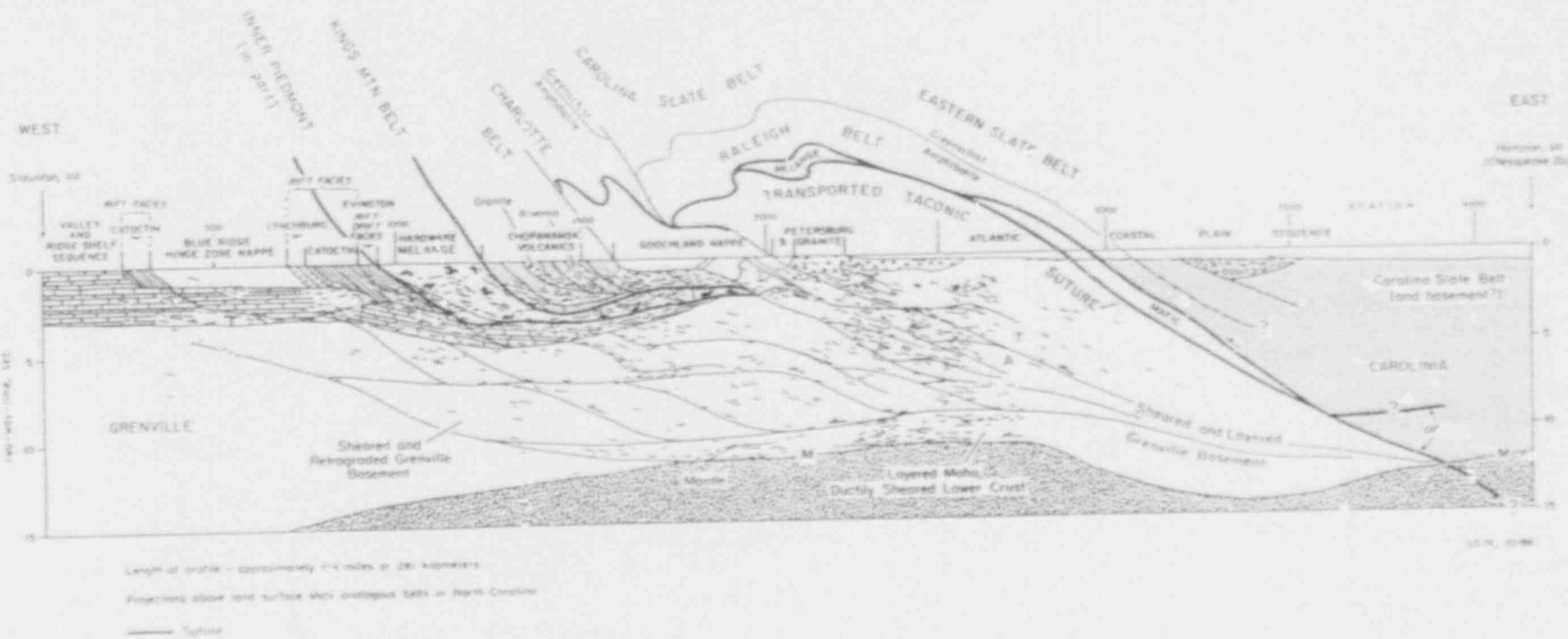


Figure 29, Glover (1989) U.S.G.S. I-64 interpretation

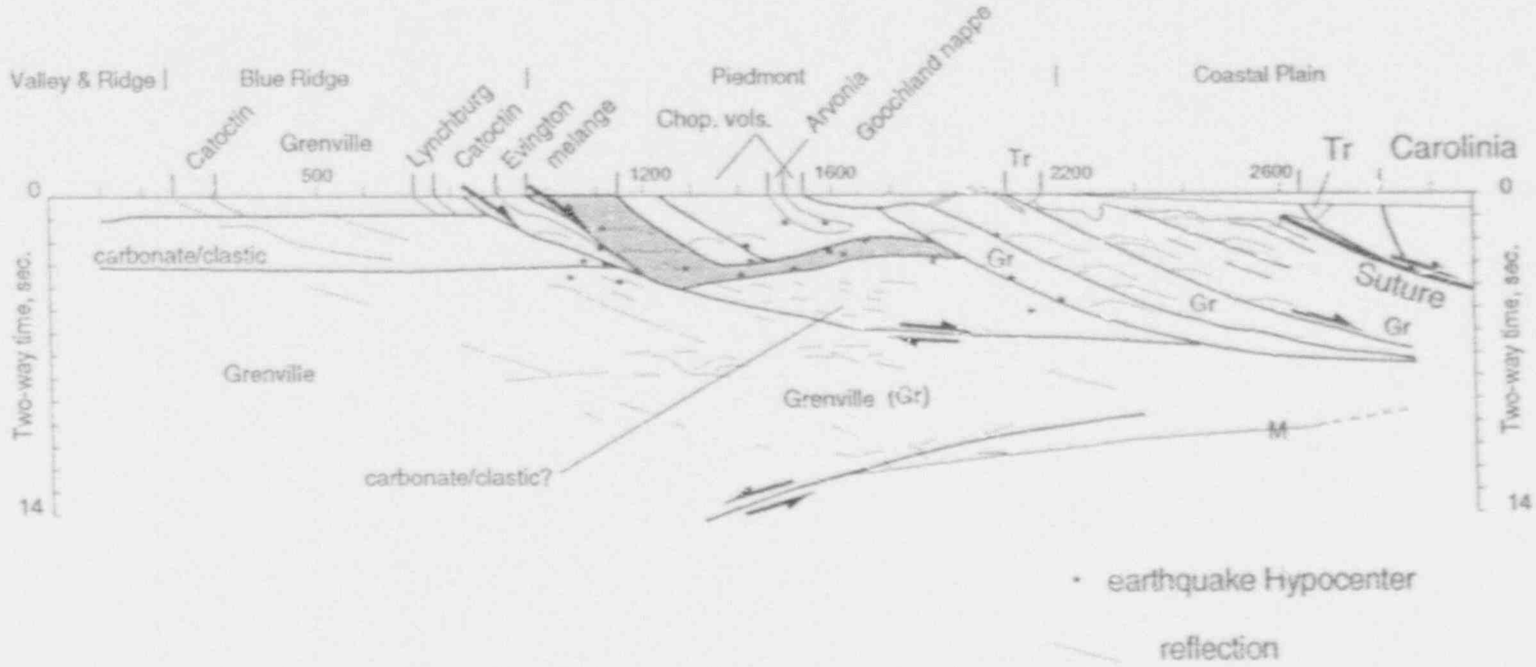


Figure 30. Glover (this volume) alternate interpretation of I-64 profile based on automatic line drawing version of Çoruh and others, 1988). All faults shown in the profile moved as thrust and/or transpressional faults during the Paleozoic. Mesozoic extensional movement is shown by arrows in cases where old contractional faults were reactivated.

been rotated counterclockwise by 20 °- 30°. This kind of rotation (really slumping) under extension is commonly called "rollover" and requires internal deformation to accommodate the change in shape. This deformation would be expressed as dominantly west-dipping high angle normal faults. Such faults are not obvious in the profile but they may be small and widely distributed, and because of their high angle orientation would not produce reflectors anyway.

Profiles I-64, 2A-1-1 and 2A-1-2 give geometric evidence of extensional faulting at the Moho. This unusual feature is another confirmation of the role of Mesozoic extension faulting on the development of the crustal structures and thinning of the curst under the Piedmont.

If this new interpretation of the I-64 profile is correct it suggests that the carbonate clastic sequence is also present in the upper midcrustal region below the melang having been offset downward from the Blue Ridge block during Mesozoic extension.

Comparison of Figures 30 and 23 shows clearly that extension has affected both profiles. It is also clear from the geometry of the profiles that the amount of extension is much greater in the central Virginia seismic zone where the rollover is greater than any other structure presently known in this part of the Piedmont.

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Part E. - Cause and localization of seismicity

by Lynn Glover, III, with addenda by Cahit Çoruh, J.K. Costain, and G.A. Bollinger; and by John Costain and G.A. Bollinger.

Cause

The stress-inducing regional cause of seismicity in the eastern U.S. is broadly conceived to be in the plate tectonic mechanism of ridge push, perhaps influenced by topographic and geologic loading (summarized in: White and Long,1989. p. 112 - 129). There may also be a drag or push exerted at the base of the lithosphere by mantle convection (Zoback and Zoback, 1980). Average stress orientation in the eastern U.S. is northeasterly (White and Long, 1989) with some variation in the Appalachian orogenic system. Causal mechanisms at this level were not addressed in the work reported on here.

Localization of earthquakes in the central Virginia region: Previous work

Granted that the lithosphere is constantly in a state of stress, this project addressed the problem of identifying any elements of the geologic framework that might be responsible for localizing seismicity.

First order structural features: Wheeler and Bollinger (1984) proposed that seismicity of the southeastern U.S. might tentatively be attributed to "characteristics and differences between various suspect terranes and the Iapetan passive margin". Quoting the Williams and Hatcher (1982) terrane map which shows a marked narrowing of the Avalon terrane where it overlaps much of the central Virginia seismic zone, Wheeler and Bollinger suggested that this part of the terrane is "most likely to have been broken by faults and other fractures, and to have had fractures reactivated,. during the growth, transport, and accretion of the terrane. Thus, the narrow parts of the terrane might have remained comparatively weak, with their fractures unhealed, so that they could be preferred areas for seismic release of strain energy.".

Glover (1989, and Part A of this report) has shown that The Piedmont terrane does not exist, because there is no suture in the Blue Ridge and "Piedmont terrane" rocks are actually deformed Laurentian margin rocks. Similarly Sheridan and others (1991), suggest that the eastern boundary of the Avalon terrane may be far east of its position, as shown by Williams and Hatcher (1982), in Virginia.. Similar arguments (Glover, Part A of this report) conflict with terrane boundaries as presented in more recent papers ([illegible] and others, 1989; Hatcher and others, 1989; Rankin and others, 1989).

Additionally, the idea presented by Wheeler and Bollinger (1984) that the narrow parts of the terrane might have remained comparatively weak, with their fractures unhealed, does not comport with a history of metamorphism that recrystallized ("healed") these rocks three times during the Paleozoic (Glover,1989). Open fractures in this part of the Piedmont may be younger than Paleozoic. The mechanical differences and anisotropy that [illegible] at the boundary between terranes even after metamorphism would make them candidates as loci of strain accumulation. However, as shown below, the Taconic suture in the central Virginia seismic zone is not oriented conformably with the slip and plane of any possible focal mechanism solution to date.

The tectonic model presented by Glover in Part A of this report differs from existing models in four important respects; 1) there is a large uplift of 1Ga Grenville basement in the eastern Piedmont of VA. 2) Only one suture (Taconic) is recognized in the exposed Appalachians, and that separates the Carolina (Avalon) magmatic terrane from the Laurentian passive margin. 3) The Chopawamsic/ James Run volcanic belt is recognized as a part of Carolinia/Avalonia, and is not a different island arc. 4) The eastern margin of Laurentia (and its upper bounding surface, the Taconic suture) extends in the subsurface below the coastal plain at least 50 kilometers east of Richmond. The impact of this model for seismicity is that seismicity is not related to terrane boundaries in any simple way because seismic zones exist within terranes as well as across terrane boundaries.

Hydroseismicity: This hypothesis, "suggests that in crustal volumes with fracture permeability, natural increases in hydraulic head caused by transient increases in the elevation of the water table in recharge areas of groundwater basins can be transmitted to depths of 10-20 km and thereby trigger earthquakes." Costain and others (1987, reproduced in Append). Costain and others' (1987) application of the hypothesis is as though one could transport the James River system anywhere on the Atlantic seaboard and where it crossed rifted crust a seismic zone would be induced. The absence of extensive seismicity in the Roanoke River groundwater basin (investigated by the Roanoke River traverse in this report) is attributed by them primarily to the lower elevations of the headwaters of the Roanoke River and consequently to a lower potential for pore-pressure fluctuations in the upper crust. Although this may be true of the Roanoke River, it is not true of the Potomac River where the seismicity under rifted areas crossed by the river is minimal at best (Costain and others, 1987).

The hypothesis is very attractive with regard to its implications for structural weakening of the rock volume, and the possibility of increasing pore pressure within a fault that is stressed to near failure. The presence of water in the upper brittle crust is a factor in the rate of release of seismic energy whether or not pore pressure fluctuations are important. It is not clear that the "hydroseismicity" hypothesis is the primary cause of the localization of seismicity. For example, if the relatively open fractures produced by Mesozoic extension were not there water would probably not penetrate deeply enough in the crust to impact the rate of seismic release.

Complex thrust and vertical shear reactivation: Another view of the causes of localization of seismicity in the central Virginia seismic zone is given by Çoruh and others (1988) The paper is reproduced in the Appendix and is updated by Çoruh here below.

An alternative interpretation; by Cahit Çoruh:

An alternative interpretation is given in Çoruh et al. (1988) using the automatic line drawing of I-64 reflection seismic data. This interpretation and interpretation of other reflection seismic data in the southeastern U.S. combined into the following alternative interpretation by Çoruh and Costain. Over much of its extent, especially between stations 1100 and 2700, the seismic reflection response in the ALD display of the I-64 data set in the central Virginia seismic zone exhibits excellent detail from the upper crust to the Moho discontinuity and suggests constraints for the geologic interpretation of the distribution of earthquake hypocenters (Figures 31, 32). On the basis of reflection data leading into the Blue Ridge from the northwest, and results of reflection profiling in other areas (Çoruh et al., 1987; Çoruh et al., 1988; Hubbard et al., 1991; and references therein), a zone of subhorizontal reflections (A) at about 3 s two-way traveltime near station 700 (Figure 32) on the western part of the line (west of Charlottesville) is interpreted to originate from parautochthonous lower Paleozoic shelf strata. Poorly reflective Grenville basement is below the deepest detachment(s) (DT in Figure 32) and shelf strata. The Blue Ridge master decollement at 1 s (BRT in Figure 32) lies at the base of the overlying allochthonous crystalline thrust sheet(s), as imaged beneath the Blue Ridge on other southern Appalachian seismic reflection data. The thickness of this metamorphic allochthon remains relatively constant over an on-strike distance of at least 400 km (Costain et al., 1987a). The Moho (M) reflections appear to be missing west of Charlottesville and east of Richmond, suggesting that the M discontinuity is more prominent in areas where the crust has been stretched.

In the middle part of the line in central Virginia, a distinctive difference in the reflectivity of the crust is apparent with respect to other parts of the line (Figure 32). The reflectors in this part are as follows (Figures 31, 32):

1. Lower crustal reflectors, including the west-dipping Moho discontinuity (M) at about 9-12 s.

2. Subhorizontal mid-crustal reflector zone (C) at 6-8 s, interpreted to represent early Proterozoic detachment zone. The east-dipping reflectors (E) above the reflectors (C) project to near surface and might be correlated with surface exposures of eastward-dipping mylonites (Gates et al., 1986). At depth, these reflectors asymptotically appear to join with C on the east, possibly because of increased shearing near the brittle-ductile transition (BDT in Figure 32). The number of east-dipping reflectors above the mid-crustal 6-8 s reflection zone C is considerably higher than below, suggesting that zone C is real and critical to any interpretation.

3. A dominant reflection package B (TS in Figure 32) undulates between 0.5 and 7 s and truncates seismic signatures that can be followed from the surface. This package defines the east flank of a large antiform about 100 km wide between stations 1100 and 2600 (Figure 32). The mid-crustal reflections (C) are interpreted to define the floor of this antiform. The antiform has a maximum vertical relief of about 17 km. The depth to the roof of the antiform varies between 3 and 18 km, where the eastward- (E)

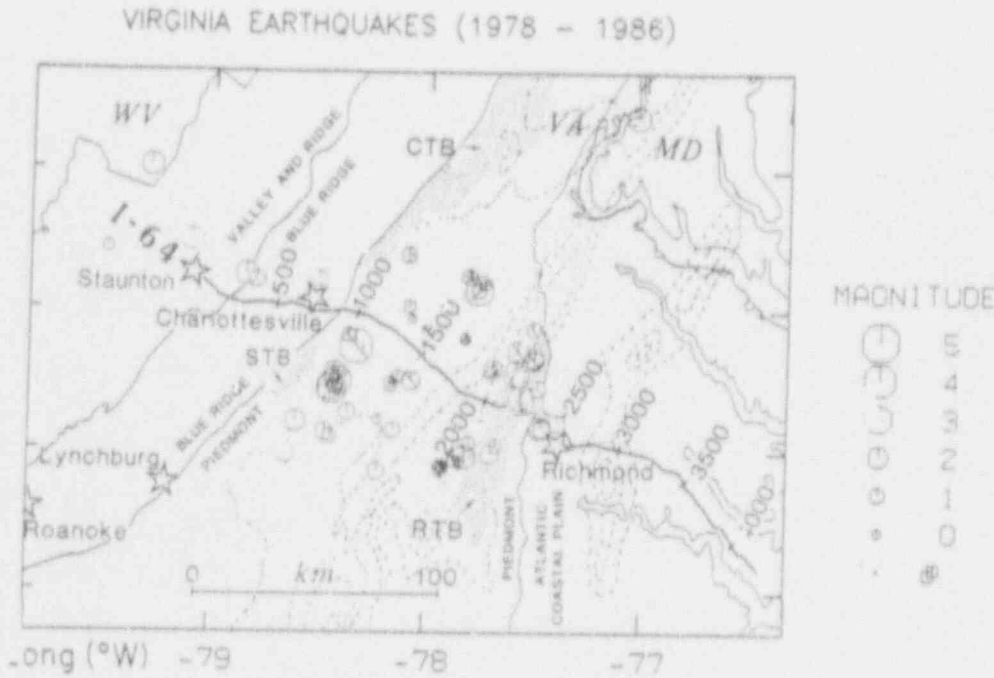

Figure 31. Earthquake epicenters along I-64 seismic reflection profile. Triassic basins: RTB, Richmond; STB, Scottsville; CTB, Culpeper. For correlation with Figure 32, hypocenters were projected into vertical plane of I-64. Shaded epicenters have a ± 5 km error ellipse. Note high density of epicenters between Scottsville and Richmond Triassic basins. Contours with dashed lines are distinct Bouguer gravity anomalies in the area. Matching aeromagnetic anomaly coincides with gravity anomaly of 20 mgal east of Richmond. Geologic boundaries from Williams, 1978; gravity anomalies from Haworth et al., 1980. From Çoruh and others, 1988.

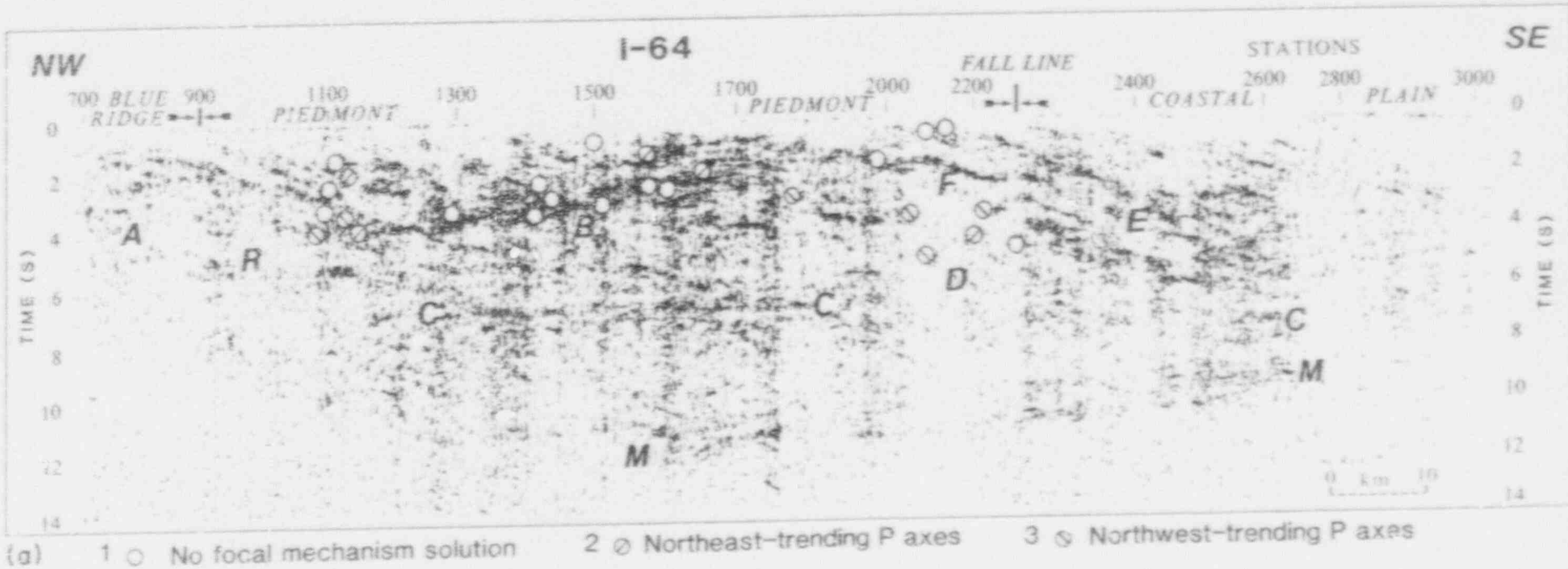

Figure 32a. Central part of automatic line drawing of I-64 seismic reflection data.

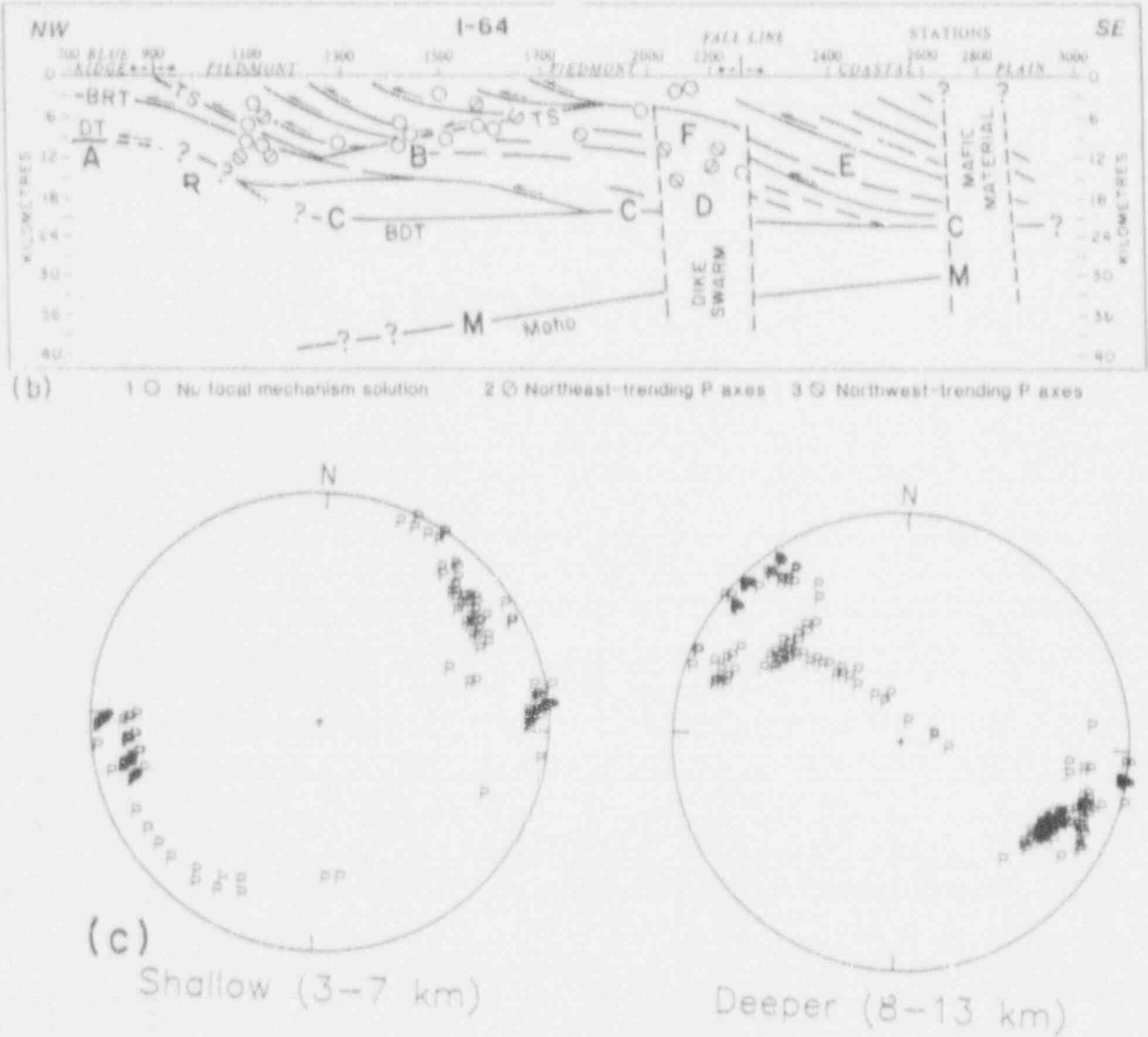

Figure 32b,c. b: Simplified cross section. A represents parautochthonous lower Paleozoic shelf strata. Below shelf strata is poorly reflective Grenville basement. Distinctive difference in reflectivity of crust is apparent with respect to western and eastern parts of profile. Large antiform is defined by reflections B, F, and E at roof and C at floor. Ramp R is interpreted to east. D is believed to be Mesozoic dike swarm; mafic material is interpreted below station 2800. Note that slope of Moho (M) reflectors at east and west of dike swarm D is different. BRT is Blue Ridge master decollement; DT is deeper detachment; TS is transported Taconic suture; BDT is brittle-ductile transition zone; and east dipping reflectors E are Alleghanian and earlier shear zones and thrusts. Circles and diagonal bars indicate projected hypocenters and orientation of P-axes, respectively. c: Orientation of P-axes from focal mechanisms for 11 events. From Çoruh and others 1988.

and westward-dipping (B) events correspond to the eastern and western flank of the antiform, respectively.

4. The antiform is bounded on the east by the east-dipping reflectors E. The change in gross reflectivity in the west is interpreted as a ramp (R) extending from the mid-crustal level to the upper crustal reflectors. West of Charlottesville (station 700), the crustal reflections disappear, except for those from lower Paleozoic shelf strata at about 2 to 3 s.

   It is suggested that imbrication by westward thrusting, crustal thinning, and a possible westward tilting (Mesozoic) are all responsible for the gross geometry of the antiform, a composite compressional-extensional feature. The imbricate structures, as well as thinning, are evident from the geometry of the reflectors of the upper crust and Moho, respectively. The westward-dipping west flank of the antiform may, in part, be related to the exposed Mesozoic basins in Virginia and might therefore be the result of westward tilting of a block of crust that slumped during Mesozoic extension along a reactivated decollement(s).

5. Between stations 2050 and 2250 the roof of the antiform is represented by a high-amplitude and narrow zone of reflections (F), below which a zone (D) shows considerably less reflectivity relative to the surrounding region. This change to less reflectivity is also apparent in the mid-crustal and Moho reflections and is interpreted to be the seismic signature of a dike swarm. Furthermore, most of the high amplitude reflections in the deep crust are attributed to injected sills (Hubbard et al., 1990). The dike swarm (D) can be correlated with the positive Bouguer gravity anomaly (Haworth et al., 1980) that extends about 80 km to the northeast (Figure 31). There is no distinct aeromagnetic anomaly (Zietz et al., 1980) related to this dike swarm.

Even with extreme processing parameters of the ALD it was not possible to decrease the difference in the apparent reflectivity of the interpreted dike swarm and other parts of the reflective crust in the central Virginia seismic zone. A similar pattern of a poorly reflective zone is interpreted below station 2800 on the east, where both Bouguer gravity and aeromagnetic anomalies are present. Those anomalies extend about 100 km to the northeast and about 50 km to the south. The fact that no distinct aeromagnetic anomaly occurs for the interpreted dike swarm below station 2100 may be attributed to its relatively great depth (6-8 km), defined by the F reflector; however, magnetic modeling suggests that the poorly reflective zones below station 2100 and 2800 do not represent the same mafic material. The lack of apparent earthquake activity related to the poorly reflective zone below station 2800 supports our interpretation that the origin and nature of these poorly reflective zones are different. Costain et al. (1987b) proposed a tectonic setting for the latest Alleghanian at which time a large strike-slip duplex was hypothesized to form in the southeast United States. Dominantly vertical structures were thus formed by a transpressional Alleghanian orogenic event, and these later became zones of weakness that were reactivated and opened during Mesozoic extension (Costain and Çoruh, 1990). We interpret the zone below station 2100 to be related to a dike swarm that was passively intruded in the weakened, reactivated crust during Mesozoic extension. The zone below station 2800 may be related to mafic material (slate belt volcanics) that was vertically aligned by transpression during formation of the Alleghanian strike-slip duplex. Late Proterozoic extensional features imaged in reflection seismic data from South Carolina by Hubbard et al. (1991) suggest that the extensional feature "D" might be an older feature to correlate with the similar features imaged in South Carolina.

Reflections that can be followed downward from exposed surface units between stations 900 and 1700 in Figure 32 are truncated by the reflections that outline the roof of the antiform on the west. The layered Catoctin metavolcanics are recognizable because of their high reflectivity (Pratt et al., 1988). The Evington and Chopawamsic are also highly reflective and appear to lie above the roof of the antiform and beneath their surface outcrops. The reflections that define the roof of the antiform on the west probably represent reactivated decollements along which the overlying rocks were transported (Pratt et al., 1988). The relatively thick zone of roof reflectors (B) and complex structures above may be due, in part, to reactivation. The geometry of the reflections from within the antiform suggests imbrication where the east-dipping events within the antiform between 3 and 7 s were interpreted by Pratt et al. (1987) to be deformation zones (mylonites), indicative either of nappe structures or major Alleghanian strike-slip deformation. To the west and east of central Virginia, the reflection data do not image the Moho on the I-64 profile. We interpret these changes in gross reflectivity to be real and due to lithologic-structural causes. Costain et al. (1987b) suggested that the no-reflection area east of station 3000 is due to the onset of a large strike-slip duplex that extends in a strike direction from central Virginia to Georgia and in a dip direction from the Brevard fault zone to the eastern Piedmont fault system. In this interpretation the most stretched crust in Virginia is between the onset (station 3000) and the offset (station 1100) of the hypothesized strike-slip duplex, indicating that maximum stretching took place here because of the wider zone of crust weakened by transpression and the development of vertical shear zones.

Earthquake hypocenters and discussion

In spite of the relative sparseness of the epicenters, the ALD display of the reprocessed I-64 reflection data suggests a spatial correlation between seismic reflectors and hypocenters. To examine the correlation, only hypocenters (shaded in Figure 31) with a ±5 km vertical error were considered. A velocity of 6 km/s was used for the conversion of vertical reflection traveltime to depth. Focal mechanisms from 11 of these earthquakes exhibit northeast-trending P (maximum compressive stress) axes for shallow sources (<8 km) and northwest-trending axes for deeper foci (>8 km;

Figure 32), and a mixture of reverse and strike slip faulting on planes that exhibit an average dip of $62 \pm 16°$. There are, however, two deeper foci with northeast-trending P axes that are exceptions to the above grouping; however, the vertical errors can easily locate these hypocenters below the given depths. Nelson and Talwani (1985) concluded that all the central Virginia focal mechanisms exhibit a stress field oriented northeast by using only P-wave polarities and a graphical analytical procedure. The results from Bollinger et al. (1986) are favored herein because they used a quantitative computer algorithm search routine that evaluates P/S wave amplitude ratios as well as P-wave polarities to obtain the required focal mechanisms.

The correlation between the hypocenters, the reflectors, and the poorly reflective zone may indicate that different seismogenic structures are associated with two different groups of hypocenters. The hypocenters in group 1 are related to the structures at the roof (B) of the antiform and above. They have shallower depths (3-7 km) and northeast-trending P axes that coincide with the general tectonic strike in the area. The events in group 2 are related to the structures within the antiform. They are deeper (8-13 km) and have northwest-trending P axes. The patterns of reflection truncations by the dike swarm suggest that the dike swarm postdates what we interpret as an older thrust zone coincident with the brittle-ductile transition zone at level C. It is suggested, therefore, that the earthquake activity in the central Virginia seismic zone may be detachment-related only on the west flank of the roof of the antiform (TS in Figure 32, the transported Taconic suture zone, probably reactivated during the Alleghanian) The hypocenters do not penetrate below the mid-crustal reflectors (C) and show no direct relation to the lower crustal reflectivity bounded by the top of the lower crust (C) and the Moho zone (M). There is no earthquake activity east of the Fall line (Figure 32), although the imaged lower crustal reflectors continue eastward along with the east-dipping reflectors. Indirect correlations between the reflectivity and the distribution of the hypocenters also suggest that the earthquake activity is limited to the parts of intensely sheared and stretched crust.

4. This is supported by seismic interval velocities of 4.5 km/s determined from the stacking velocities beneath station 900 for the interval interpreted as Paleozoic shelf strata (Li et al., 1990). This relatively low interval velocity suggests unmetamorphosed rocks and constrains the thickness of the metamorphic plate.

Comments, by Lynn Glover, III, on the above Çoruh and others interpretation.

Çoruh and others note an arch-like "antiform" in the upper crust with a crest at about station 2000 (Figure 32). They further propose that the structure was formed by both compressional and extensional means. Although they offer little justification for this speculation, the suggested origin has merit and the rationale is here discussed more fully in a later section.

Çoruh and others also call upon the Costain and others (1987, 1990) strike slip duplex model for the development of the central and southern Appalachian Piedmont to explain the vertical panels of low reflectivity in a 12 km-wide panel below station 2100 and a much wider panel below station 2800. Reference to models by Woodcock and Fischer (1986) shows that the geometry of the Piedmont structure in the I-64 traverse and Plate 1 (Geologic map of the VA Piedmont and Blue Ridge) does not resemble a strike-slip duplex. This is because there are no vertical faults bounding vertical horses that are known in the region (Glover, Part A of this report). Late Paleozoic dextral transpression occurred along moderately eastward dipping ductile faults (Glover, Part A) such as those mapped at the surface in the Goochland nappes and traced into the lower crust on the I-64 profile between stations 2000 to 2400. Therefore there were no vertical faults to be injected by Mesozoic diabase dikes which Çoruh and others (1988) call upon to produce the vertical panels of low to no reflectance in the I-64 profile. Lateral extension of the crust to create a 12 km wide panel of low reflectivity implies a composite width of vertical dikes measured in kilometers. Yet if present, it seems likely that these dikes would reach the surface of this the most concentrated accumulation of Mesozoic dikes ever postulated in the Piedmont. If, as Çoruh and others believe, the dikes rose no higher than the bright reflector "F" (Çoruh and others, 1988, Figure 2) at about two seconds two-way time below station 2100, then the extensive lateral separation below that level would place the crust above it in tension so that rifts would occur there. However, the surface area in question has been well studied (Bobyarchick and Glover, 1979; Poland, 1976; Reilly, 1980; Glover, unpublished) and does not contain vertical strike slip faults of ductile or brittle nature. The Triassic graben located over the western side of the low reflectance panel under stations 2000 - 2250 formed as a result of brittle reactivation of the ductile Hylas fault zone which dips moderately eastward. Movement on this fault during the Mesozoic would obviously not relieve the extension below it during a postulated dike injection episode.

Çoruh and others suggest that the postulated Mesozoic dike swarm under station 2100 is, ..."correlated with the positive Bouguer gravity anomaly that extends about 80 km to the *northeast* (Figure 31)". This is not supported by the *N20°-30°W* trend of the field of narrow, elongate anomalies shown on the Aeromagnetic Map of Virginia (Zietz and others, 1977) which correlate well with the Mesozoic dikes shown on the Geologic Map of Virginia (Calver and others (1963), published before the magnetic data was available.

One can also see dipping reflectors passing through these low reflectance panels. Whatever the cause of the low reflectance panels, and assuming that they are real, they are superimposed on the Paleozoic structure of the I-64 profile without deforming it. They are also only seen in records using the unpublished Automatic Line Drawing display program of Cahit Çoruh.

Coruh and others suggest that, "... the earthquake activity in the central Virginia seismic zone may be detachment-related only on the west flank of the roof of the antiform (TS in Figure 32), the transported Taconic suture zone, probably reactivated during the Alleghanian)." The spatial correlation is there but the focal mechanisms indicate reverse and strike slip faulting on planes oriented at high angles to the gentle west-dipping structure of Çoruh and others. The attitude of nearly all of the preferred focal mechanism planes is northwesterly and the average attitude of five of the six preferred planes below 9 km in the *eastern and western* parts of the central Virginia seismic zone (CVSZ) is N20 (-20°,+10°)W 50°(+19°-15°) NE.

Comparison of the Roanoke River (RRT) and James River (JRT) traverses with respect to seismicity: The Roanoke River profile shows somewhat less westward slumping of the structure during Mesozoic extension. Clearly, under station 800, backslipping has occured, but the amount of rollover is less as measured by the more gentle westward dip of the crystalline plate between stations 800 and 1600. The westward extensional fault along which the rollover took place is at the same stratigraphic and structural position in both profiles. The thickness of the crystalline plate is as much as 9 km thick in the I-64 profile, while it is only about 3 km thick under the Roanoke River traverse. In both cases the crystalline Piedmont is believed to be underlain by relatively unmetamorphosed Cambrian - Ordovician carbonates, sandstone and shale. Greater ductility of carbonates and clastics may allow aseismic deformation in these rock volumes in both profiles, as most of the seismicity of the I-64 profile plots within the upper plate crystallines. The magnetic map (Zietz, 1977) indicates that three or four large Mesozoic dikes cross the Roanoke River traverse but they trend more northerly and are not as abundant as along the James River.

It may be significant that the earthquake hypocenters of the central Virginia seismic zone cluster around and in the inversion structure (westward slump of the central Piedmont block during the Mesozoic, shown in Figure 30) in I-64. The epicenter map shows a very diffuse zone, a shape compatible with the volume expected to be affected by slump-generated normal faults. This would facilitate deep penetration of groundwater, which in turn could reduce the yield point of the rock volume under stress and increase the frequency of seismic events. In this case the central Virginia seismic zone is conspicuous because of the frequency of small events. A corollary might be that the aseismic regions have fewer but larger seismic events with periodicities longer than the historical record (ie. > 500 years), and probably longer than about 5000 years, the length of the record in the eastern United States examined by Amick and Gelinas (1991).

Correlations of hypocenters and focal plane orientations with structures on seismic reflection profiles in the central Virginia seismic zone: On profiles I-64, NRC-10, and 2A-1-1 through 4, hypocenters have been projected along NE-SW structural strike between 1 - 5 km. into the planes of section. This introduces some error into the position shown on the profiles, an error that is in addition to the 1-5 km vertical error in position of the hypocenters related to uncertainties in location. Therefore, the apparently very good correlation between hypocenters and postulated faults as shown on these profiles needs to be addressed with caution.

In Figure 33 it can be seen that the preferred focal planes generally have attitudes at high angles to the gently west-dipping suture between 1.5 and 3.5 sec. two-way time below stations 1200 to 1700. (Figure 29, 30). Above this west-dipping structure are moderately east dipping (30 - 40°) fault planes whose strike, from surface mapping, is about N20°E (thus an average attitude would be about N20°E 35°SE). This contrasts with the preferred focal plane attitudes of about N20-[illegible] W, 42 - 79°E and N20-30° E 40-60°W. Choosing the alternate focal plane does not remedy the lack of concordance with known structure. It seems probable that the structures seen by seismic reflection have

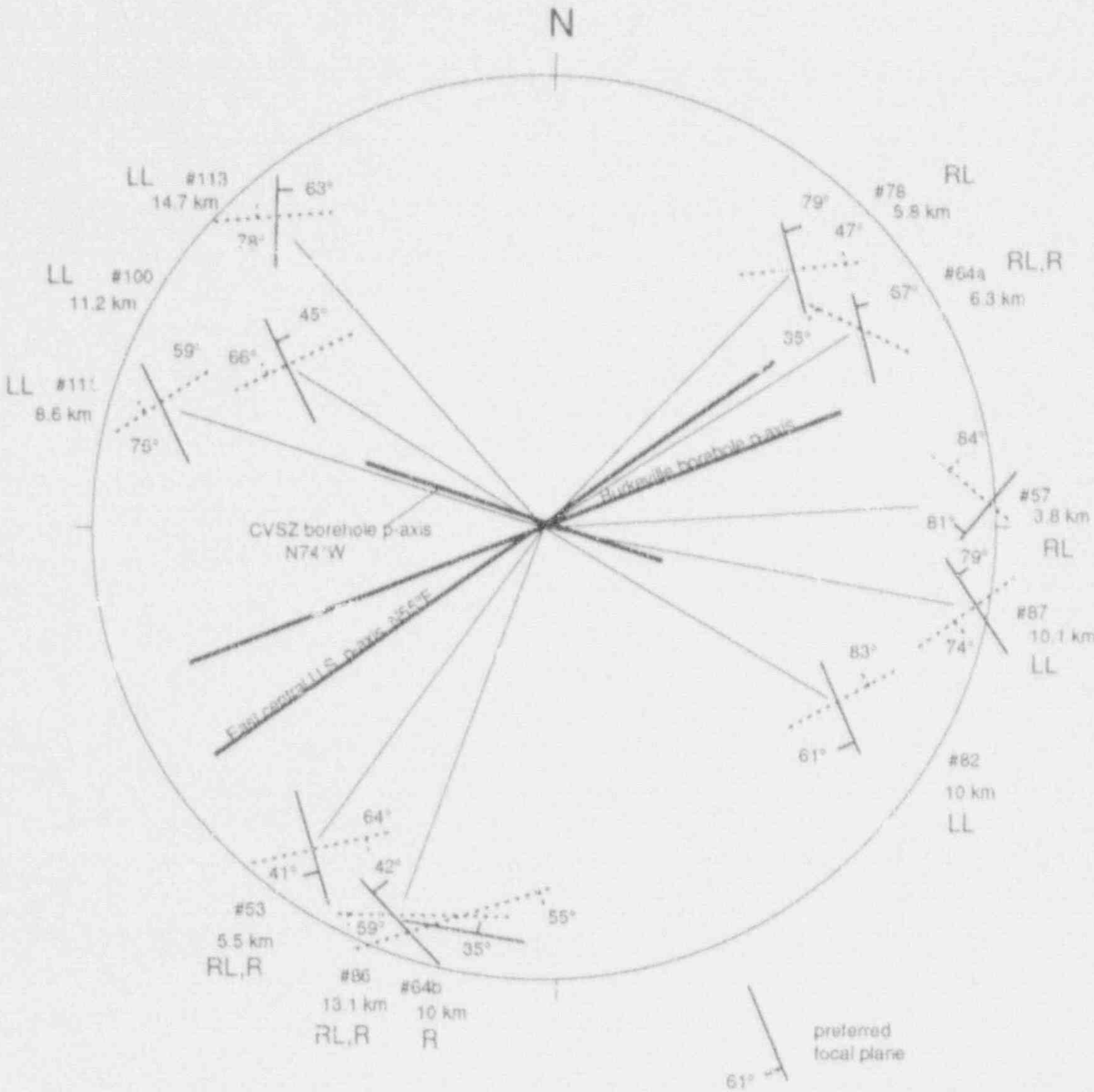

Figure 33. Stereo plot of p-axes in the central Virginia seismic zone with map orientation of focal planes. Earthquake numbers are shown with depth to hypocenters. RL, R = right lateral with strong reverse component, etc. Data from Bollinger and others(1985) and Munsey and Bollinger (1985). Borehole p-axes from Rundle and others (1987).

little relation to the seismogenic structures implied by the focal plane orientations. The same conclusion applies to all of the seismic reflection profiles taken in the James River corridor. The only geologic structure known in the CVSZ with an orientation close to that of the focal plane orientation are dikes of Mesozoic age (Munsey and Bollinger 1985).

Mesozoic dike contacts emerge as a possible seismicity-localizing anisotropy. Although there is little data from field measurements, Mesozoic dikes are usually observed to dip near 90° whereas the NW striking focal planes dip 40° to 80°.

Seismic reflection and surface geologic mapping therefore provide evidence that simple models of fault reactivation from Paleozoic fault structures are inadequate to explain seismicity in central Virginia, but Mesozoic dikes may be associated with the seismicity.

Relationship of regional and local p-axes to the orientation and slip on focal planes: A stereo plot (Figure 33) of p-axes for the central Virginia seismic zone, borehole p-axes within and outside of the zone, and the dominant east central U.S. p- axis are shown with relation to the orientations of possible focal planes for 11 events. Single-event p-axes trend NE and NW, with no well defined partitioning between shallower and deeper crust. A 300 meter borehole p-axis measurement in the CVSZ of N74°W ± 13° (Rundle and others, 1987) is consistent with the group of NW-trending single-event p-axes. They also recorded a N74°E ± 10° p-axis approximately 40 miles SW of the first hole and outside of the central Virginia seismic zone. This p-axis from an aseismic region in the Atlantic Seaboard is conformable with the p-axes in the east central U.S. west of the Appalachians. Near the Ramapo, N.Y fault zone they measured p-axes near the seismic zone boundary and within it and both axes were N69°E and N72°E.

The borehole p-axis variation inside and outside of the CVSZ is repeated in the Moodus zone of New England where Rundle and others (1987) measured trends. nearly identical to those in Virginia. Therefore, it appears that the p-axes as measured in boreholes within some of these seismic zones are different from those outside of the zone.

As measured by focal plane mechanism studies within the CVSZ, both NE and NW trending p-axes can be inferred. This is puzzling because it seems physically impossible for two different stress vectors to exist simultaneously in the same volume of rock. It is well known that p-axes determined from focal mechanism solutions do not represent unique solutions because of the effect that anisotropies in the crust can have on the orientation of the plane of failure (McKenzie, 1969). Therefore, the focal planes and slip vectors can be considered much closer to real*ity than the stress vectors derived from them* (Gephart and Forsyth, 1985).

From the above, it would seem that if a single regional p-axis can be found that will satisfy the focal plane and slip vector data then that should be the real stress vector we are looking for. A N55°E p-axis generally concordant with the east central U.S. field west of the Appalachians appears to satisfy the data. In most cases the alternate focal plane of Munsey and Bollinger (1985) is the one that is concordant with the required orientation and slip, exceptions are events # 78 and 64a which are the preferred orientations of Munsey and Bollinger. The northeast set of focal planes strikes about 30° east of common Appalachian structural strike in the area and the dip is mostly NW, opposite to that of Appalachian structure.

Attempts to graphically find a NW trending regional p-axis that would be concordant with the data have failed. A computer oriented approach to testing various models will be initiated.

The problem of the NW borehole p-axis within the seismic zone might be explained as a refraction of the regional field as a result of local stress release (see Zoback, 1987). This idea will require future testing.

Conclusions:

1. A new tectonic model of the Appalachian orogen indicates that one, not two or more, terrane boundaries is present in the Piedmont and Blue Ridge of the central and southern Appalachians.
2. This terrane boundary is the Taconic suture, it has been transported in the allochthonous Blue Ridge/Piedmont crystalline thrust nappe, and it is repeated at the surface by faulting and folding associated with later Paleozoic orogenies.
3. The suture passes through the lower crust and lithosphere somewhere east of Richmond.
4. The suture is spatially associated with seismicity in the central Virginia seismic zone, but is not conformable with earthquake focal planes and appears to have little causal relation to their localization.
5. A velocity and Q study in central Virginia implies that the gross mineralogy at depth in the upper crust is free of hydrous phases.
6. Subsurface structure in the central Virginia seismic zone differs in several ways from that along strike in the aseismic Roanoke River traverse. The metamorphic Blue Ridge/Piedmont plate probably overlies carbonates and clastics in both areas, but the metamorphic plate is 9 km thick in the central Virginia seismic zone but only 3 km thick in the Roanoke River traverse. As estimated by the amount of rollover (westward slumping during the Mesozoic), the central Virginia seismic zone may be more pervasively broken by distributed high angle normal faults than is the Roanoke River area. This implies greater access to deep upper crustal crystalline rocks by groundwater. Deeper penetration by groundwater may reduce the yield point of rock under stress and shorten the period of seismicity. This implies that the central Virginia seismic zone is localized by groundwater access. A corollary may be that the aseismic areas have very long period (>500 to 5000 ? years) seismicity and earthquakes of greater magnitude.
7. Focal mechanism planes of Munsey and Bollinger (1985) have attitudes of, 1) NW to NNW strike and steep NE or SW dips, or 2) ENE to NE strike and steep NW or SE dips. These planes are all at rather high angles to Paleozoic structure and would seem unrelated to it. The NNW set is somewhat concordant with the strike of Mesozoic dikes in the area but not with their dip.
8. Focal plane solutions in the Appalachians commonly give both northwesterly and northeasterly striking p-axes. Because it is unlikely that the same rock volume could transmit two distinct p-axes, one or both of them may be wrong.

9. Single seismic event p-axes are dependent only on the orientations of the focal planes which may be strongly influenced by crustal anisotropies (McKenzie, 1969). The focal planes and slip axes are the more likely to be real. Preliminary attempts to fit a single regional p-axis to all of the planes of Munsey and Bollinger (1985) gives an apparently good fit for a N55°E trending p-axis. This is approximately parallel with the dominant NE regional p-axis west of the Appalachians.
10. The best fit focal planes are oriented generally ENE, dip NW and SE steeply and are not concordant with any geologic structure in the area.

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Appendix-

Related Publications

Seismogenic structures in the central Virginia seismic zone

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ABSTRACT

A correlation between earthquake hypocenters and seismic reflection data in central Virginia has become apparent on an "automatic line drawing" (ALD) display of seismic reflection data. With the reprocessed Virginia I-64 reflection-Vibroseis data extended to 14 s, reflectors are imaged from the lower crust as well as from the upper crust. Specifically, the improved resolution and data quality of ALDs have produced an image of an antiformal structure bounded by mid-crustal reflections on the bottom and by major thrusts at the top. The reflections that define the roof of the antiform are most prominent from about 6 s (18 km) on the east near Richmond under the Coastal Plain sediments, to 1–1.3 s (3–4 km) between Richmond and Charlottesville, and to 3.5 s (10.5 km) on the west. Seismic signatures that can be followed downward from the surface between Charlottesville and Richmond appear to be truncated at the roof of the antiform. The dominant reflections that define the roof correlate with the seismic signature of the transported Taconic suture on the west flank and mylonites on the east flank.

The distribution of hypocenters in the area shows an excellent correlation with the westward-dipping reflections that form the roof of the antiform on its western flank. Earthquake activity in this locale may be related to reactivation of the thrusts defining the roof and/or faults above the antiformal structure; however, distribution of the easternmost and deepest set of hypocenters appears to be related to an extensive near-vertical diabase dike swarm of Mesozoic age.

INTRODUCTION

We suggest a correlation between earthquake hypocenters and subsurface structures interpreted from seismic reflection data in the central Virginia seismic zone, an area of persistent, generally low level seismicity in the Virginia Piedmont (Fig. 1). The results of instrumental monitoring by the Virginia Regional Seismic Network showed a spatial distribution of hypocenters that was both vertically and horizontally diffuse. Such a pattern was attributed by Bollinger et al. (1986) to multiple, rather than single, seismogenic structures. Herein we suggest a specific correlation between the distribution of the hypocenters and seismic reflector patterns as imaged on the Virginia I-64 seismic profile (Fig. 1).

The I-64 Vibroseis profile begins in the Virginia Valley and Ridge and crosses the adjacent Blue Ridge anticlinorium (Wehr and Glover, 1985), as shown in Figure 2. Grenville basement, interpreted to underlie the deepest detachment, has a relatively low reflectivity. Grenville basement in other areas of the southeast (Costain et al., 1986; Çoruh et al., 1987) also has a relatively low reflectivity due, in part, to lack of laterally continuous and relatively undeformed reflectors. On the surface east of the Blue Ridge are exposures that include Lynchburg metasedimentary rocks, Catoctin metavolcanic rocks, Evington Group metasedimentary rocks, and Chopawamsic metavolcanic rocks (Conley, 1978; Wehr and Glover, 1985, and references therein). All those units have undergone metamorphism and deformation in either the Taconic, Acadian, or Alleghanian events (Glover et al., 1983).

SEISMIC REFLECTORS

The central Virginia seismic zone provides an exceptional opportunity to correlate earthquake studies and seismic reflection data. The original I-64 seismic line (Harris et al., 1982) along In-

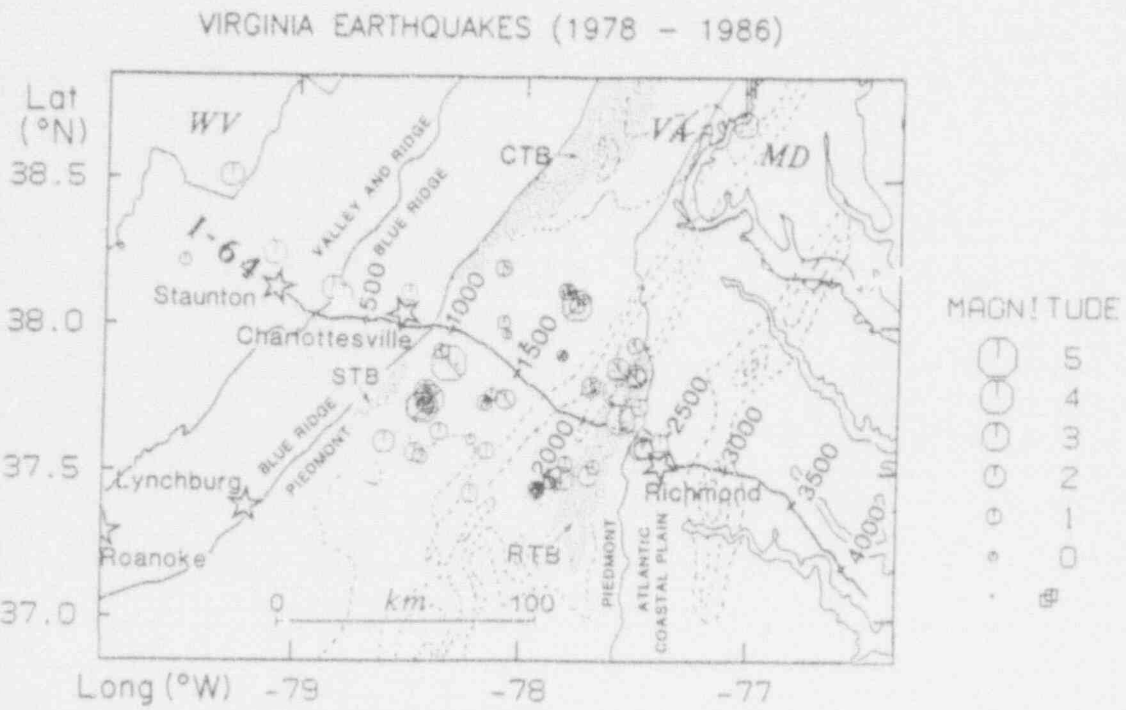
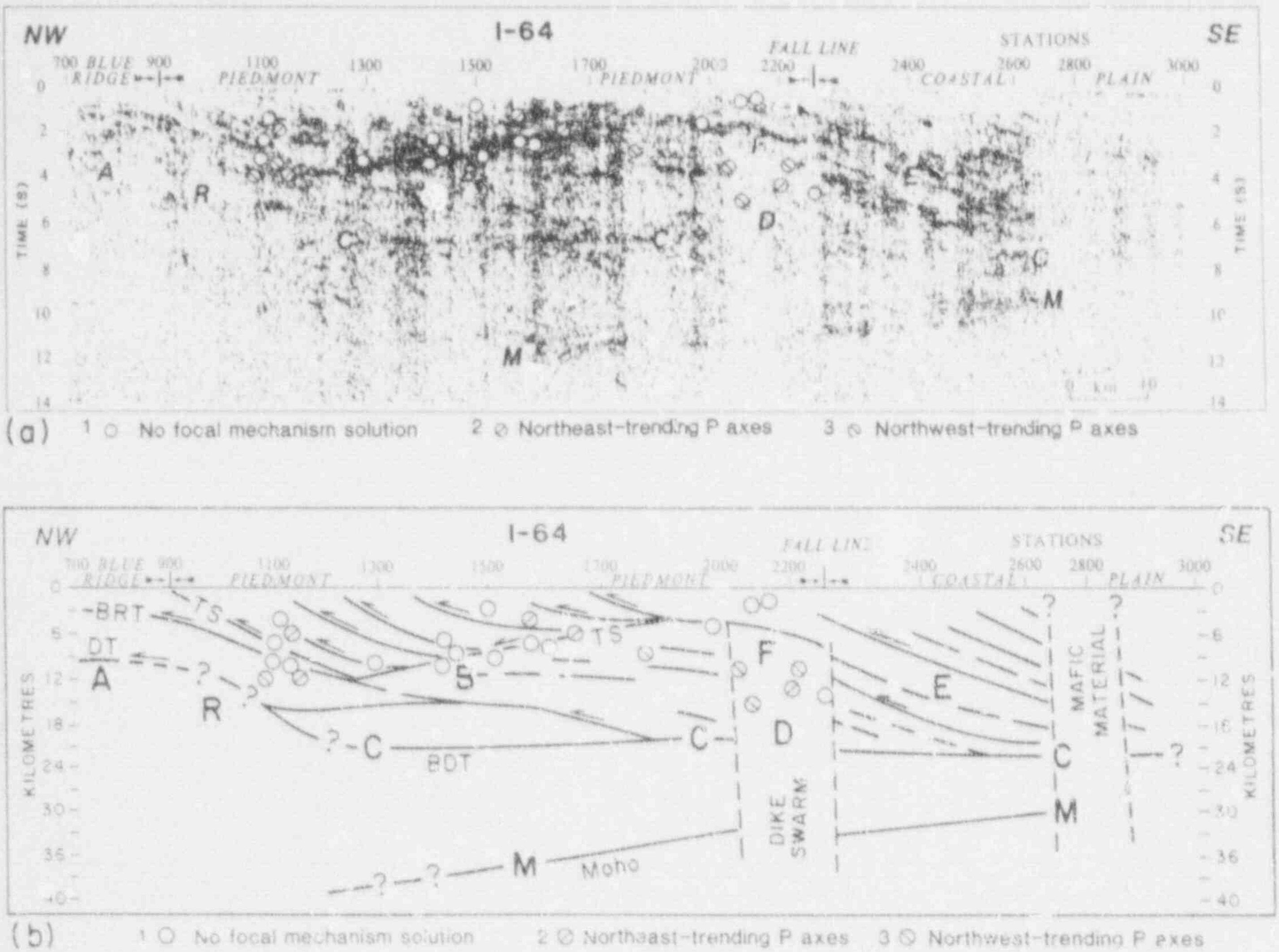


Figure 1. Index map of Virginia with earthquake epicenters showing I-64 seismic reflection profile, boundary of geologic provinces, and location of exposed Richmond (RTB), Scottsville (STB), and Culpeper Triassic basins (CTB). For correlation shown in Figure 2, hypocenters are projected into vertical plane defined by northwest-southeast line that averages I-64 profile. Shaded epicenters have ±5 km error ellipse. Note high density of epicenters between Scottsville and Richmond Triassic basins. Contours with dashed lines are distinct Bouguer gravity anomalies in area. Matching aeromagnetic anomaly coincides with gravity anomaly of 20 mgal east of Richmond. (Geologic boundaries from Williams, 1978; gravity anomalies from Haworth et al., 1980.)



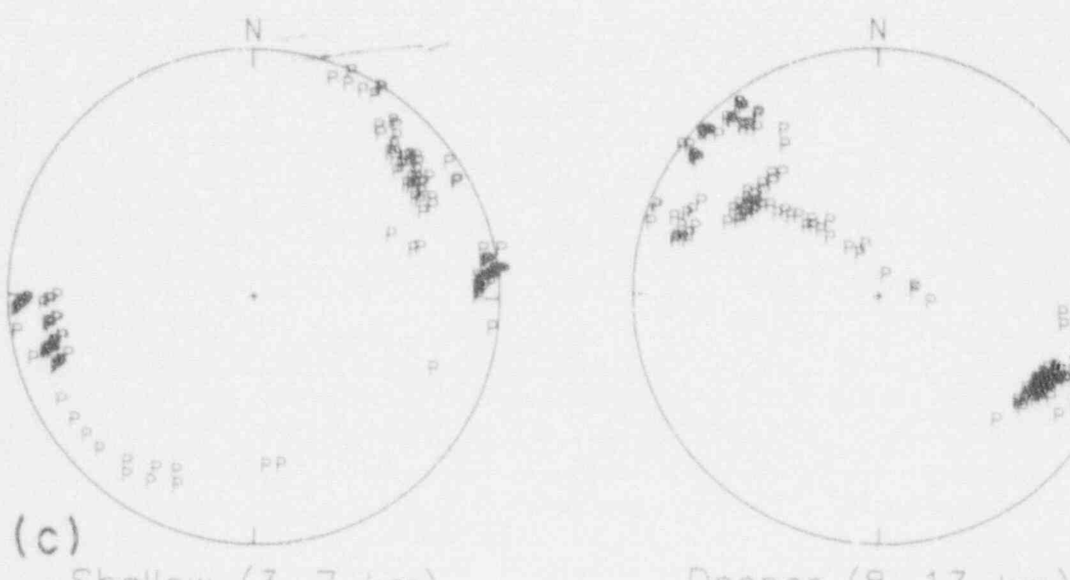


Figure 2. a: Central part of automatic line drawing of I-64 seismic reflection data. b: Simplified cross section. A represents parautochthonous lower Paleozoic shelf strata. Below shelf strata is poorly reflective Grenville basement. Distinctive difference in reflectivity of crust is apparent with respect to western and eastern parts of profile. Large antiform is defined by reflections B, F, and E at roof and C at floor. Ramp R is interpreted to east. D is believed to be Mesozoic dike swarm; mafic material is interpreted below station 2800. Note that slope of Moho (M) reflectors at east and west of dike swarm D is different. BRT is Blue Ridge master decollement; DT is deeper detachment; TS is transported Taconic suture; BDT is brittle-ductile transition zone; and east-dipping reflectors E are Alleghanian and earlier shear zones and thrusts. Circles and diagonal bars indicate projected hypocenters and orientation of *P*-axes, respectively. c: Orientation of *P*-axes from focal mechanism solutions for 11 events.

terstate 64 (I-64) between Staunton, Virginia and the Atlantic coast (Fig. 1) was reprocessed at Virginia Tech by Pratt (1986) to produce 14 s records by using extended Vibroseis correlation with the application of Vibroseis whitening (Çoruh and Costain, 1983). Pratt produced the key seismic reflection section in Virginia to image crustal reflectors, including the Moho discontinuity in the 9–12 s range. The I-64 data were subjected to the process of automatic line drawing, an objective technique, to produce an unconventional display in which the *original waveforms are preserved*. The processing is based on the idea that the reflections from a subsurface reflection point are on several traces in a time-space window. The ALD processing converts the traces into coherency estimations from a moving window in the time-space domain. The resulting form of the seismic data is considered to be a section of relative seismic reflectivity (Fig. 2a).

Over much of its extent, especially between stations 1100 and 2700, the seismic reflection response in the ALD display of the I-64 data set in the central Virginia seismic zone exhibits excellent detail from the upper crust to the Moho discontinuity and suggests constraints for the geologic interpretation of the distribution of earthquake hypocenters (Fig. 2b). On the basis of reflection data leading into the Blue Ridge from the northwest, and results of reflection profiling in other areas (Çoruh et al., 1987, and references therein), a zone of subhorizontal reflections (A) at about 3 s two-way traveltime near station 700 (Fig. 2a) on the western part of the line (west of Charlottesville) is interpreted to originate from parautochthonous lower Paleozoic shelf strata. Poorly reflective Grenville basement is below the deepest detachment(s) (DT in Fig. 2b) and shelf strata. The Blue Ridge master decollement at 1 s (BRT in Fig. 2b) lies at the base of the overlying allochthonous crystalline thrust sheet(s), as imaged beneath the Blue Ridge on other southern Appalachian seismic reflection data. The thickness of this metamorphic allochthon remains relatively constant over an on-strike distance of at least 400 km (Costain et al., 1987a). The Moho (M) reflections appear to be missing west of Charlottesville and east of Richmond, suggesting that the M discontinuity is more prominent in areas where the crust has been stretched.

In the middle part of the line in central Virginia, a distinctive difference in the reflectivity of the crust is apparent with respect to other parts of the line (Fig. 2b). The reflectors in this part are as follows (Fig. 2a and 2b).

1. Lower crustal reflectors, including the west-dipping Moho discontinuity (M) at about 9–12 s.

2. Subhorizontal mid-crustal reflector zone (C) at 6–8 s, interpreted to represent the present brittle-ductile transition zone. The east-dipping reflectors (E) above the reflectors (C) project to surface exposures of eastward-dipping mylonites (Gates et al., 1986). At depth, these reflectors asymptotically appear to join with C on the east, possibly because of increased shearing near the brittle-ductile transition (BDT in Fig. 2b). The number of east-dipping reflectors above the mid-crustal 6–8 s reflection zone C is considerably higher than below, suggesting that zone C is real and critical to any interpretation.

3. A dominant reflection package B (TS in Fig. 2b) undulates between 0.5 and 7 s and truncates seismic signatures that can be followed from the surface. This package defines the east flank of a large antiform about 100 km wide between stations 1100 and 2600 (Fig. 2). The mid-crustal reflections (C) are interpreted to define the floor of this antiform. The antiform has a maximum vertical relief of about 17 km. The depth to the roof of the antiform varies between 3 and 18 km, where the eastward- (E) and westward-dipping (B) events correspond to the eastern and western flank of the antiform, respectively.

4. The antiform is bounded on the east by the east-dipping reflectors E. The change in gross reflectivity in the west is interpreted as a ramp (R) extending from the mid-crustal level to the upper crustal reflectors. West of Charlottesville (station 700), the crustal reflections disappear, except for those from lower Paleozoic shelf strata at about 2 to 3 s.

We suggest that imbrication by westward thrusting, crustal thinning, and a possible westward tilting (Mesozoic) are all responsible for the gross geometry of the antiform, a composite compressional-extensional feature. The imbricate structures, as well as thinning, are evident from the geometry of the reflectors of the upper crust and Moho, respectively. The westward-dipping west flank of the antiform may be in part related to the exposed Mesozoic basins in Virginia and might therefore be the result of westward tilting of a block of crust that slumped during Mesozoic extension along a reactivated decollement(s).

5. Between stations 2050 and 2250 the roof of the antiform is represented by a high-amplitude and narrow zone of reflections (F), below which a zone (D) shows considerably less reflectivity relative to the surrounding region. This change to less reflectivity is also apparent in the mid-crustal and Moho reflections and is interpreted to be the seismic signature of a dike swarm. The dike swarm (D) can be correlated with the positive Bouguer gravity anomaly (Haworth et al., 1980) that extends about 80 km to the northeast (Fig. 1). There is no distinct aeromagnetic anomaly (Zietz et al., 1980) related to this dike swarm.

Even with extreme processing parameters of the ALD it was not possible to decrease the difference in the reflectivity of the interpreted dike swarm and other parts of the reflective crust in the central Virginia seismic zone. The apparent sharpness of boundaries of the dike swarm that mark the change in reflectivity in Figure 2a is partly due to the scale of the plot, and thus the boundaries are interpreted to be real. Zone D spatially correlates with the Richmond Triassic rift basin (Fig. 1). A similar pattern of a poorly reflective zone is interpreted below station 2800 on the east, where both Bouguer gravity and aeromagnetic anomalies are present. Those anomalies extend about 100 km to the northeast and about 50 km to the south. Figure 1 includes the Bouguer anomaly contours for this area where the gravity and magnetic anomalies coincide. The fact that no distinct aeromagnetic anomaly occurs for the interpreted dike swarm below station 2100 may be attributed to its relatively great depth (6–8 km), defined by the F reflector; however, magnetic modeling suggests that the poorly reflective zones below station 2100 and 2800 do not represent the same mafic material. If they did, an appreciable magnetic anomaly should also be observed around station 2100. The lack of apparent earthquake activity related to the poorly reflective zone below station 2800 supports our interpretation that the origin and nature of these poorly reflective zones are different. Costain et al. (1987b) proposed a tectonic setting for the latest Alleghanian, at which time a large strike-slip duplex was hypothesized to form in the southeast United States. Dominantly vertical structures were thus formed by a transpressional Alleghanian orogenic event, and these later became zones of weakness that were reactivated and opened during Mesozoic extension (Costain and Çoruh, 1988). We interpret the zone below station 2100 to be related to a dike swarm that was passively intruded in the weakened, reactivated crust during Mesozoic extension. The zone below station 2800 may be related to mafic material (slate belt volcanics) that was vertically aligned by transpression during formation of the Alleghanian strike-slip duplex.

Reflections that can be followed downward from exposed surface units between stations 900 and 1700 in Figure 2 are truncated by the reflections that outline the roof of the antiform on the west. The layered Catoctin metavolcanics are recognizable because of their high reflectivity (Pratt et al., 1988). The Evington and Chopawamsic are also highly reflective and appear to lie above the roof of the antiform and beneath their surface outcrops. The reflections that define the roof of the antiform on the west probably represent reactivated decollements along which the overlying rocks were transported (Pratt et al., 1988). The relatively thick zone of roof reflectors (B) and complex structures above may be due, in part, to reactivation. The geometry of the reflections from within the antiform suggests imbrication where the east-dipping events within the antiform between 3 and 7 s were interpreted by Pratt et al. (1987) to be deformation zones (mylonites), indicative either of nappe structures or major Alleghanian strike-slip deformation. To the west and east of central Virginia, the reflection data do not image the Moho on the I-64 profile. We interpret these changes in gross reflectivity to be real and due to lithologic-structural causes. Costain et al. (1987b) suggested that the no-reflection area east of station 3000 is due to the onset of a large strike-slip duplex that extends in a strike direction from central Virginia to Georgia and in a dip direction from the Brevard fault zone to the eastern Piedmont fault system. In this interpretation the most stretched crust in Virginia is between the onset (station 3000) and the offset (station 1100) of the hypothesized strike-slip duplex, indicating that maximum stretching took place here because of the wider zone of crust weakened by transpression and the development of vertical shear zones.

EARTHQUAKE HYPOCENTERS AND DISCUSSION

In spite of the relative sparseness of the epicenters, the ALD display of the reprocessed I-64 reflection data suggests a spatial correlation between seismic reflectors and hypocenters. To examine the correlation, only 26 hypocenters (shaded in Fig. 1) with a ±5 km vertical error were considered. A velocity of 6 km/s was used for the conversion of vertical reflection traveltime to depth. Focal mechanisms from 11 of these earthquakes exhibit northeast-trending *P* (maximum compressive stress) axes for shallow sources (<8 km) and northwest-trending axes for deeper foci (>8 km; Fig. 2c), and a mixture of reverse and strike-slip faulting on planes that exhibit an average dip of $62° \pm 16°$. There are, however, two deeper foci with northeast-trending *P* axes that are exceptions to the above grouping. Nelson and Talwani (1985) concluded, by using only *P*-wave polarities and a graphical analytical procedure, that all the central Virginia focal mechanisms exhibit a stress field oriented northeast. We favor the results from Bollinger et al. (1986), because they used a quantitative computer algorithm that evaluates *P/S* wave amplitude ratios as well as *P*-wave polarities to obtain the required focal mechanisms.

The correlation between the hypocenters, the reflectors, and the poorly reflective zone may indicate that different seismogenic structures are associated with two different groups of hypocenters. The hypocenters in group 1 are related to the structures at the roof (B) of the antiform and above. They have shallower depths (3–7 km) and northeast-trending *P* axes that coincide with the general tectonic strike in the area. The events in group 2 are related to the dike swarm (D) within the antiform. They are deeper (8–13 km) and have northwest-trending *P* axes. The patterns of reflection truncations by the dike swarm suggest that the dike swarm postdates what we interpret as the brittle-ductile transition zone at level C. We suggest, therefore, that the earthquake activity in the central Virginia seismic zone may be detachment-related only on the west flank of the roof of the antiform (TS in Fig. 2b, the transported Taconic suture zone, probably reactivated during the Alleghanian). The hypocenters do not penetrate below the mid-crustal reflectors (C) and show no direct relation to the lower crustal reflectivity bounded by the top of the lower crust (C) and the Moho zone (M). There is no earthquake activity east of the Fall Line (Fig. 2), although the imaged lower crustal reflectors continue eastward along with the east-dipping reflectors. Indirect correlations between the reflectivity and the distribution of the hypocenters also suggest that the earthquake activity is limited to the parts of intensely sheared and stretched crust. Similar ALD displays in this zone and in other areas where seismic activity is also present, e.g., the Piedmont in South Carolina, are necessary for more general conclusions to be reached.

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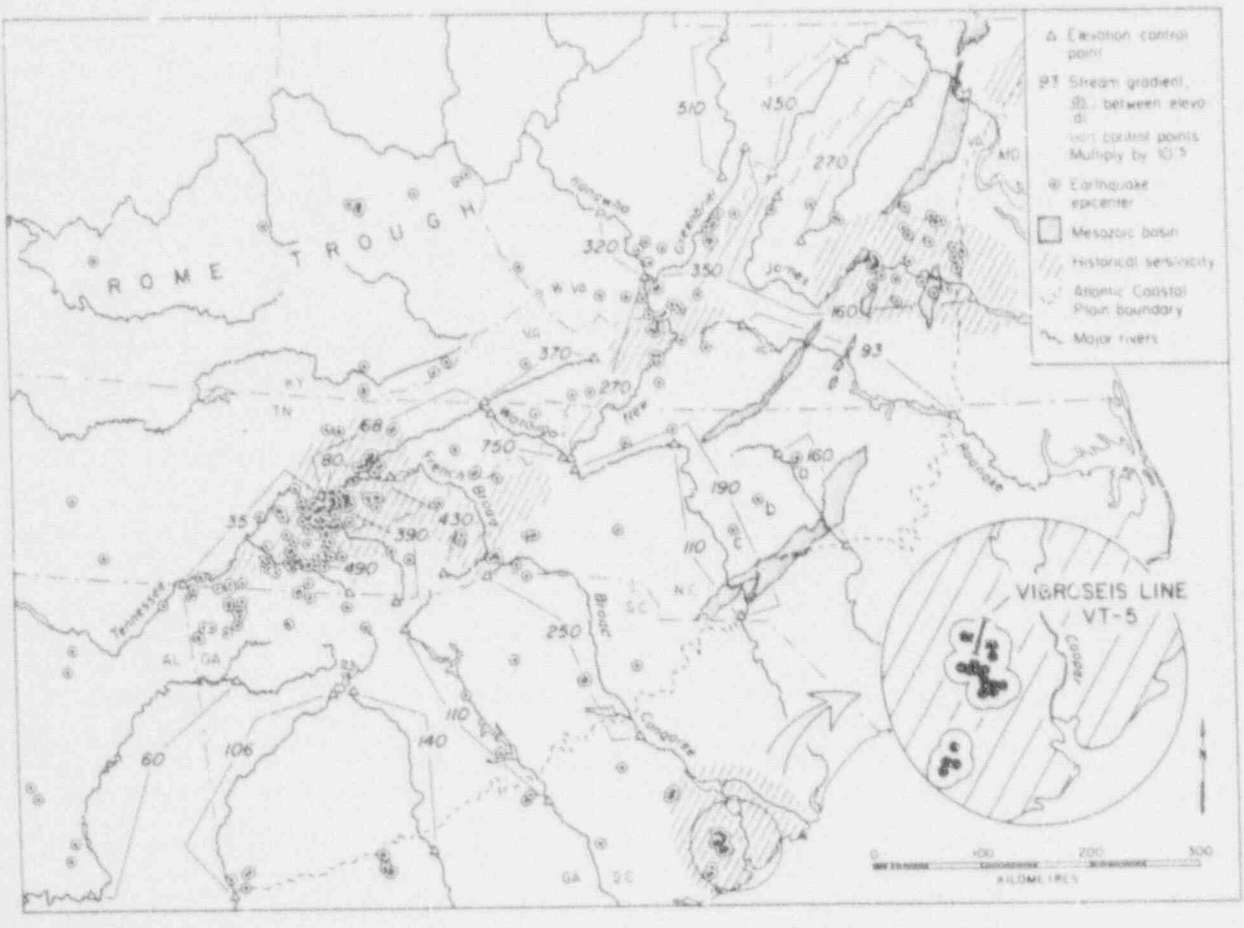

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HYDROSEISMICITY: A HYPOTHESIS FOR THE ROLE OF WATER IN THE GENERATION OF INTRAPLATE SEISMICITY

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ABSTRACT

A new hypothesis termed Hydroseismicity that has hydrologic (diffusion of pore pressure transients from recharge areas of groundwater basins), geologic (rifted, fractured crust), and chemical (solubility of minerals) elements is proposed to explain the role of water in the generation of intraplate seismicity. Its basis is a spatial correlation in the southeastern U. S. between 1) seismogenic crustal volumes of high seismicity, 2) large gravity-driven river basins that can provide an adequate supply of water to the upper- and mid-crust, and 3) a permeable crust that is tectonically stressed close to failure. It is suggested that in crustal volumes with a combination of connected fractures and adequate groundwater, natural transient increases in hydraulic head in recharge areas of groundwater basins can be transmitted to depths of 10-20 km, and thereby trigger earthquakes, via a flow-path geometry that resembles except for scale the model familiar to groundwater hydrologists for near-surface flow. Possible trigger mechanisms for Hydroseismicity include small increases in fluid pressure at hypocentral depths caused by such transient increases, and hydrolytic weakening of minerals that leads to structural weakening. Implicit in the model is a *diffuse* distribution of epicenters (as is observed in the region) rather than concentrations along discrete geologic (faults) or geomorphic (rivers) elements. Open fractures imply fracture roughness, i.e., asperities under a higher stress that keep fractures open even in an ambient tectonic stress field. Intraplate earthquakes in a fractured crust prestressed to near-failure are thus postulated to be triggered by small transient increases in fluid pressure transmitted along preexisting fractures in a rock fabric weakened by stress corrosion of asperities. Abundant petrologic evidence is available to justify an assumption of fracture permeability to depths of 20 km near passive rifted margins. All four principal seismogenic volumes in the southeastern U. S. are within gravity-driven groundwater basins that can provide an abundant supply of water to the crust, and that intersect known or suspected Eocambrian or Mesozoic rifted crust. The host basins have the largest surface recharge areas and contain rivers with the highest average stream gradients as measured from their headwaters to the Fall Line. Seismicity in the region is characterized by steeply dipping focal mechanism nodal planes and diffuse alignments and/or clusters of epicenters. These characteristics are compatible with a steep to vertical fracture fabric currently being reactivated by pore pressure diffusion from surface recharge of groundwater basins.

> *"As happens sometimes in science, the appearance of a new concept that synthesizes a vast volume of experimental material gets a cool reception from a considerable fraction of the scientific community. Yet, eventually, despite all doubts (and frequently, violent criticism), the new concept may win general recognition if it proves to be helpful and time saving in explaining both old and new experimental data. Under such circumstances the criticism gives way to proclamations that the ideas underlying the new concept were set forth years earlier and that the concept proper is no more than a mere repetition of the forgotten past. The elements of old ideas inevitably coexist in any new concept, for new ideas do not appear in a void. Therefore, by selecting individual statements made in previous epochs and by subjecting these concepts to appropriate interpretation in the light of present-day knowledge, one may produce the impression that the new concept coincides completely with erroneously forgotten older ideas."*
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INTRODUCTION

Many references to possible empirical correlations between seismicity, rainfall, and rivers have been made. Part of the reluctance to introduce surface meteorological boundary conditions into a theoretical model that includes the flow of meteoric water to mid-crustal depths has been:

1. No general recognition of an apparent correlation between various aspects of basin geometry and recharge, groundwater flow, and seismicity,

2. Uncertainty about the magnitude of a fluid pressure transient that could be propagated from the surface recharge area of a groundwater basin to mid-crustal depths,

3. Lack of data about the nature of the in situ stress field at hypocentral depths as well as the level of stress required to trigger earthquakes,

4. Lack of evidence to justify the required values of intrinsic permeability to depths of 20 km,

5. No proposed trigger mechanism that incorporates groundwater recharge areas and the downward propagation of fluid pressure transients.

We address each of the above and suggest a model for intraplate seismicity. Examining a few of the published suggestions for a relationship between meteorological surface conditions and seismicity, we agree in general with the possibility of such correlations. The observations of these early workers can be interpreted to suggest that somehow transient increases in fluid pressure are transmitted downward into a fractured upper crust where they can trigger earthquakes. What was missing in previous correlations was a conceptual physical model to transmit these transients from surface recharge areas down to depths of 20 km. Also, justification is required to demonstrate that hydraulic conductivity can occur to such depths and to identify the mechanism(s) by which fluids can trigger earthquakes.

Our hypothesis suggests that natural increases in hydraulic head can indeed be transmitted to depths of 10-20 km in the crystalline upper crust via groundwater flow-paths that resemble the geometry of the model familiar to groundwater hydrologists for near-surface flow, and thereby trigger earthquakes by transient mechanical, in combination with chemical, means. Although we believe that our conceptual model may have global application for eventually leading to an understanding of the role of water in the generation of intraplate seismicity, i.e., the New Madrid area and the Basin and Range province, we focus herein primarily on the southeastern United States.

PREVIOUS CORRELATIONS BETWEEN FLUID PRESSURE TRANSIENTS AND SEISMICITY

Many references have been made by others about apparent correlations between earthquakes and groundwater pressure gradients resulting from water injection in deep wells, increased rainfall and stream discharge, thermal springs (Gastil and Bertine, 1986), and the filling of large reservoirs (Simpson, 1986); however, few attempts have been made to correlate regional changes in the height of the water table with deep seismicity. The importance of the role of fluid pressure in decreasing the shear strength of rocks is generally accepted both theoretically and experimentally. On a large scale, the classic paper by Hubbert and Rubey (1959) addresses the influence of groundwater fluid pressures on the movement of thrust faults by invoking the Mohr-Coulomb failure theory in its effective stress formulation.

Water injection-induced seismicity

Confirmation of the effects of elevated fluid pressures on earthquake generation was made by Evans (1966) in connection with a disposal well near Denver, Colorado. During the period April, 1962 to September, 1965, 710 small earthquakes were recorded in the Denver area where the only previously recorded earthquake had occurred in 1882. Evans noted that the first earthquake occurred just one month after the first injection of liquid wastewater at a disposal well at the U. S. Army's Rocky Mountain Arsenal. The well penetrated sedimentary strata, bottoming at a depth of 3.67 km in fractured Precambrian schist and granite gneiss. Evans also noted that the earthquake frequency over the period 1962-1965 correlated with the volume of wastewater injected. Epicenters of almost all the shocks were located within a circular area 16 km in diameter centered at the Rocky Mountain Arsenal. The increases in fluid pressures caused by wastewater injection triggered small fault movements. Subsequently, Healy *et al.* (1968) concluded that the Hubbert-Rubey mechanism provided a complete and satisfactory explanation for the triggering of the Denver earthquakes.

In a follow-up study, Healy (1975) reported on earthquakes associated with the Rangley, Colorado oil field from 1969 through 1974. This site was chosen because of seismic activity that occurred during the latter stages of exploitation of the oil reservoir using injection of water as part of a secondary recovery program. When the fluid pressure in the seismically active zone was reduced under controlled conditions, the seismic activity was greatly decreased, especially in the region within 1 km of the control wells. In November 1972, the fluid pressure was increased and a new series of earthquakes was triggered. In March 1973, pumping was reversed, with an attendant decrease in the earthquake activity. As part of the same study, Raleigh *et al.* (1972) determined the frictional properties of rock core taken from the oil field. Those data, together with in situ stress determinations, allowed an independent calculation of the values of fluid pressure, p, at which seismicity would be expected to occur. The predicted critical level was $p = 2.57 \times 10^7$ N/m^2. Values of fluid pressure in the seismically active part of the reservoir at the time of frequent earthquakes were measured at 2.75×10^7 N/m^2, establishing beyond doubt the importance of fluid pressure as a critical parameter in the triggering of earthquakes.

Talebi and Cornet (1987) monitored microseismicity induced by fluid injection into a granitic rock mass. They concluded that the microseismicity occurred because of shear failure along preexisting fissures through which water percolated from a main fracture and not along the main fracture itself.

The increases in pore pressure from fluid injection (a few hundred bars) are much higher than those to be expected from the filling of a reservoir or from increases in the height of a water table (about 0.1 bar/meter rise in the reservoir or water table); however, the stress required to trigger an event must depend on the ambient stress level as well as a complex of factors that include the orientation of preexisting open fractures, fracture permeability, the relative importance of hydraulic weakening, and the amount of water available.

Increased rainfall and stream discharge-induced seismicity

Drake (1912) and Sayles (1913) published early empirical correlations between earthquakes and rainfall. Drake concluded for China that "the rapid and strong atmospheric variations, assisted to some extent by rain, are the forces most effective in the final stage of earthquake activity.".

Taber (1914) noted that, for the seismic activity in the epicentral region of the historic Charleston earthquake of 1886, "A close relationship between rainfall and earthquake frequency in this district is clearly indicated by every method of comparison that has proved possible of application.". Taber noted that the ". . . period of high seismic activity beginning in 1886 followed two years of unusually heavy precipitation which must have resulted in an important elevation of the water table. While the rainfall for 1886 was much lower than for the previous years, 86 per cent of it occurred prior to the earthquake in August, and during June 10.78 inches were recorded. The years of comparatively low rainfall between 1886 and 1891 are marked by a period of lower seismicity, while the year of heavy precipitation in 1893 coincides with a year of very high earthquake frequency. The period of great seismic activity occurring in 1893 began at 11:05 p.m. June 20th with a shock having an intensity of about VII R.-F., and during that month 16.50 inches of rain fell at Charleston, this being next to the highest record for any single month since the establishment of the Weather Bureau Station."

"The yearly rainfall curve suggests no explanation for the relatively high seismicity of 1896, but when the monthly records are examined it is seen that the rainfall was above the average during the first few months of the year, while in June and July the precipitation was 7.57 and 10.58 inches respectively. Most of the earthquakes of that year occurred during August and September."

"The heaviest precipitation for the past eighteen years was recorded in 1912 and there was a noticeable increase in the earthquake frequency for that year. Moreover, most of this rain fell during the first half of the year instead of in the last half as is customary, and on June 12th there occurred the severest earthquake experienced since 1893 and possibly since 1887."

"During the period of high seismicity extending from 1886 to 1897, the average annual rainfall at Charleston was 52.17 inches, while during the period of relatively low seismicity from 1898 to date the average annual rainfall has been only 39 inches."

Although the above quotations from Taber (1914) document only *local* correlations near Charleston between seismicity and rainfall, they do support a relationship between earthquake frequency and precipitation.

Wong and Simon (1981) suggested that the annual tenfold increase in the discharge of the Colorado River in the Paradox Basin, Utah and corresponding change in the river stage may influence the occurrence of microseismicity along the river. 95 per cent of that seismicity was observed along a 35-km long section of the Colorado River, whereas the remaining 5% microseismicity was *diffusely* distributed throughout other portions of the Paradox Basin. Hypocentral depths in this region were generally less than 15 km.

Nava (1983) suggested that naturally occurring variations in the water load of the Mississippi River, and fluctuations in barometric pressure and rainfall may be linked to the unusual frequency patterns observed in the New Madrid seismic zone. She found correlations between variations in Mississippi River stage and New Madrid seismicity. On the basis of a large and accurate data set, Nava concluded that a causal relationship does exist between New Madrid seismicity and the level of the Mississippi River. McGinnis (1963) also related earthquake frequency to water load in southeastern Missouri.

McClellan (1984) noted that the observed frequency of earthquakes occurring in the spring over a 25-year interval before the San Francisco earthquake of 1906 significantly exceeded seasonal frequencies expected from random variations in the earthquake rate. McClellan proposed mechanisms for the springtime seismicity observed there that include an increase in pore-fluid pressure by a downward-propagating pulse of above average crustal pore pressure due to increased infiltration of water from surface and subsurface reservoirs refilled by winter rainfall.

Thus, rainfall-induced seismicity has been proposed as a triggering mechanism for earthquake swarms for several years. We acknowledge that statistics relating major earthquake occurrences to streamflow and high-water stages are not well established and that the evidence for a connection between rainfall and seismicity is tenuous. We feel, however, that the conceptual groundwater basin model proposed herein as the vehicle for Hydroseismicity has neither been proposed nor investigated previously.

Aftershocks

Nur and Booker (1972) suggested that large shallow earthquakes can induce changes in the fluid pore pressure that are comparable to stress drops on faults, and that a redistribution of pore pressure by fluid flow subsequent to the main event slowly decreases the strength of the rock possibly resulting in delayed fracture. They concluded that the flow of groundwater can provide the viscous element necessary to produce aftershock sequences, and that the number of aftershocks per unit time decays initially as $t^{-1/2}$. Sibson *et al.* (1975) suggested that considerable volumes of groundwater are rapidly redistributed by fluid diffusion after shallow earthquakes.

Reservoir-induced seismicity

The association of a significant increase in seismic activity during and after the filling of some large reservoirs has been well documented (Bell and Nur, 1978 and references therein). More recently, O'Reilly and Rastogi (1986) describe case histories of reservoir-induced seismicity from India, China, Japan, the U.S.S.R. and the U.S.A. The deepest reservoirs add stress loads of only about 20 bars; . more likely cause of the seismicity is the result of the small increase in stress triggering the release of a large prestress, i.e., an increase in pore pressure stress acting to trigger failure along a preexisting fault already tectonically stressed close to failure.

As Bell and Nur (1978) pointed out for the earthquake sequence at Oroville, California, the most important implication of a pore pressure trigger mechanism is the presence of water to depths of 10 km or more in the crust. Simpson (1976) summarized characteristics of reservoir-induced seismicity in various parts of the world and concluded that the potential for reservoir-induced seismicity appears to be highest in areas of strike-slip or normal faulting, and that induced seismicity is most common in areas of high to moderate strain accumulation. Except for the Kremasta reservoir in Greece, the time delay between the first presence of water in the reservoir and the largest earthquakes is at least as long at that predicted by the curve of Scholz *et al.* (1973).

The Aswan High Dam (110 m) on the Nile River impounds the second largest reservoir in the world. The reservoir began to fill in 1964 and the water level rose slowly until 1975 when a water depth of approximately 93 m was reached (Simpson *et al.*, 1982a). Since then the water level has varied seasonally between 89 and 94 m. No earthquake activity (M > 5) was known within 100 km of the area prior to 1975. On November 14, 1981 a magnitude 5.5 earthquake occurred near the western edge of the reservoir, 50 km upstream from the dam, and was followed by a long sequence of aftershocks (Simpson *et al.*, 1982b). The main shock occurred four days after the seasonal maximum in water level in the reservoir. The largest aftershock was of magnitude 4.5 on August 20, 1982, and occurred 8 days after the water level began to increase at a rate of 3-4 cm/day following the seasonal minimum (Simpson *et al.*, 1982a). The hypocenters of the aftershocks are at depths of 5-25 km (Toppozada *et al.*, 1982; Simpson *et al.*, 1982b). Most of the activity was confined to a narrow zone of right-lateral strike-slip faulting at the western edge of the reservoir at those depths. Simpson *et al.* (1982b) noted that the stress at hypocentral depths is influenced by the load and by the pore pressure of the reservoir. Simpson *et al.* (1982b) and Toppozada *et al.* (1982) concluded that the seismicity was triggered by the filling of the Aswan Reservoir.

The best and most recent documented case of reservoir-induced seismicity in the southeast U.S. is the filling of the Monticello Reservoir, South Carolina (Talwani, 1979; Talwani et al., 1980). The frequency of earthquakes increased about three weeks after filling began, with the largest events (M_L up to 2.6-2.9) occurring up to two years later. The seismicity has declined with time, but remains greater than before impounding. The earthquakes are interpreted to occur along numerous, small, pre-existing fractures throughout the volume of the rocks (Duc, 1980; Talwani, 1981; Talwani et al., 1980) and are believed to be triggered in response to the small added stress and pore pressure changes related to reservoir filling (Zoback and Hickman, 1982; Fletcher, 1982). By monitoring the spatial growth of epicenters associated with filled reservoirs, and assuming it to be associated with the diffusion of pore pressure, Talwani and Acree (1985) computed a "seismic hydraulic diffusivity", equal within an order of magnitude to D, the hydraulic diffusivity. They concluded that diffusion of pore pressure in the upper one km of the crust accounts for the time lag between the onset of reservoir-induced seismicity and the impoundment of a reservoir 100 m or greater in depth.

We propose that seismicity occurring at focal depths of 10-20 km shares some common feature with reservoir-induced seismicity. Fluid pressure is assumed to trigger shallow reservoir-induced seismicity by decreasing the shear strength of rocks. For the deep seismicity, perhaps the weakening of the rock along preexisting microfractures by the presence of fluid causing stress corrosion is also required.

EVIDENCE FOR FRACTURE PERMEABILITY TO DEPTHS OF 20 KM

The continental lithosphere is weaker than the oceanic lithosphere by about a factor of three (Vink *et al.*, 1984). Because of this, lithospheric rifting, while prevalent in the continents, rarely occurs in oceanic regions. Any rifting close to an ocean-continent boundary will prefer a continental pathway (Vink *et al.*, 1984). Evidence for large-scale fracture anisotropy in the southeast U. S. can be construed to include fault mechanisms that typically indicate strike-slip motion along vertical to steeply-dipping surfaces. In addition, the occurrence of swarms of Jurassic dikes and widespread zeolite-lined/filled steep-to-vertical fractures is suggestive that any intrinsic permeability that has persisted since active extension is strongly anisotropic and is greater in the steep-to-vertical directions.

Zoback (1983) noted that the density of fractures in holes entered for in situ stress determinations decreases only moderately with depth. The Russians' discovery of flowing water in fractures at depths of 11 km (Kozlovsky, 1982) is direct and spectacular evidence for fracture permeability at depth. Indirect, but equally impressive, evidence for in-place fracture permeability at mid-crustal depths is the aforementioned seismicity associated with the filling of the Aswan Reservoir in Egypt.

There is extensive petrologic and experimental evidence for significant and widespread permeability and circulating water in crystalline rocks at great depths. In the past, it was often assumed that water in most crystalline rocks at depth was present in small amounts and, because of the low permeability of the rocks, did not move far or rapidly. The effects of fluid flow on metamorphic reactions, mass transport, and deformation processes have been explored recently (Bruton and Helgeson, 1983; Etheridge *et al.*, 1983), but the implications of the permeability of the rocks to change the rigidity properties of large volumes of the crust have received much less attention. This section reviews the petrologic evidence for permeability in crystalline rocks and the probable nature of the crustal permeability in the southeastern U.S.

Isotopic

Study of stable oxygen and hydrogen isotopes has indicated that the crust is permeable to flow of fluid in convection cells set up as a result of emplacement of hot plutons. This is true for the rather shallow Skaergaard Intrusion (Taylor and Forester, 1979) and the Idaho batholith (Taylor, 1977, 1978; Criss *et al.*, 1982; Criss and Taylor, 1983). The Skaergaard produced flow through fractures to depths of 6-10 km in the country rocks with water volumes 0.34 to 1.5 times that of the rock. The inferred bulk intrinsic permeability is 10^{-11} cm² (1 darcy ≅ 10^{-8}cm²) for the surface basalt (0-7 km depth), 10^{-13}cm² for the gabbro (4-8 km), and 10^{-16}cm² for the gneissic country rocks (7-10 km). Younger plutons emplaced in the Idaho batholith produced convection cells within the batholith over an area of 5,000 km² to depths comparable to the Skaergaard. Depending on the locality, the volume of fluid was variable with rock/fluid volume ratios of 0.01 to 28. On the basis of geothermal gradients and the temperatures in hot springs systems, Blackwell (1984) concluded that present groundwater circulation within the Idaho Batholith must extend to depths of over 6 km. For both the Skaergaard and Idaho batholith, the water was meteoric. While these two plutons are rather shallow, studies in the Coast Range batholith of British Columbia, emplaced at depths of about 10 km or greater, show comparable features (Margaritz and Taylor, 1976; Taylor, 1977; 1978).

In examining the tendency of regional metamorphic rocks to exhibit oxygen isotopic equilibration, Rumble and Spear (1983) found that equilibration is more likely in fractured rocks. Equilibration took place by means of flow of an aqueous fluid through intergranular pores. One such regional metamorphic area which contained reequilibrated oxygen isotopic compositions was the Franciscan of California where blueschist metamorphism occurred under conditions of 200-300°C and ≥ 8 kbar.

Wickham and Taylor (1985) concluded from oxygen and hydrogen isotopic data from the regional metamorphic terrane of the Trois Seigneurs Massif, France that there had been massive infiltration of externally-derived pore fluids, perhaps sea water, to depths of 12 km during the time of active metamorphism. Wickham and Oxburgh (1985) suggested that the most reasonable tectonic setting for the area is a rifted continental crust. They do not have evidence for the actual fluid pathways but noted that many similar rifted settings have extensive flow of fluids at depth.

Kerrich *et al.* (1984) used oxygen isotopic compositions to identify the source, temperature, and amount of fluids transported through rocks in fault zones and their deeper counterparts, brittle-ductile shear zones. They found that some fault systems may be closed systems, but that many develop large-scale permeability as they develop. The fluids evolve from small quantities of high temperature and pressure, locally derived fluids to large quantities of meteoric or cognate fluids at depths of up to 15 km. They concluded that flow of deep (metamorphic) and shallow (meteoric) liquids can be coeval and noted that such flow has been implicated in contemporary seismic activity on the San Andreas fault (Irwin and Barnes, 1975; Sibson, 1982).

Mineral assemblages

Abundant fluid flow can also be documented by mineral equilibria and changes in rock compositions. In the metamorphic rocks of south-central Maine, the occurrence of the pyrite → pyrrhotite transformation by desulfidation (Ferry, 1981), appearance of biotite by decarbonation of calcareous metapelites (Ferry, 1984), and changes in rock compositions (Ferry, 1982, 1983) suggested that volumes of fluid, equal to 0.9 to 2.2 times the rock volumes, flushed through the rocks during metamorphism. Metamorphic conditions were 380-520°C and 3.5 kbar. A granitoid pluton emplaced in the same terrain (Ferry, 1978; 1979) produced a flow of fluid through the pluton equal to 0.1-1.0 rock volumes. The source of these fluids cannot be identified by the techniques used, although Ferry (1979) points out that the amount of fluid evolved from the metamorphic rocks equals that needed to change the isotopic composition of the granite.

Stone and Kamineni (1982), in studying the fractures of the Eye-Dashwa Lakes pluton, concluded that the fracturing began before the pluton completely solidified (650-600°C) and continued to temperatures below 100°C. The fracture fillings are believed to have crystallized from fluids passing through the fractures. Unfortunately, pressures could not be estimated in this instance but it does illustrate the almost continual formation of fluid-filled fractures in a crystalline terrane from plutonic depths to the near-surface.

Rock Structures

Intrinsic permeability of a rock mass can be inferred from the measurement of spacing and aperture of the fractures. This technique yielded permeabilities of 10^{-7}cm² to 10^{-9}cm² in the Mayflower granitoid stock and greater than 10^{-10}cm² in the Pike's Peak Granite (Brace, 1980).

The widespread occurrence of spaced cleavage in low-grade arenaceous rocks, suggested to Etheridge *et al.* (1983, 1984) that an amount of fluid equal to a fluid/rock volume ratio of 10^3 was required to flow through greenschist facies metamorphic rocks. This assumes that the spaced cleavage originates by loss of silica to a migrating fluid which is able to dissolve only a small amount of silica.

Planar arrays of fluid inclusions in minerals are presumed to form by the healing of fluid-filled fractures, trapping some of the fluid. This would suggest that the rocks fractured in the presence of a fluid phase during some time in its metamorphic or igneous history. Based on the volume of fluid trapped, a minimum fracture width of 0.02 μm in a variety of metamorphic terrains was suggested by Walther and Orville (1982). While it is possible that these fluid inclusions represent what has been described as exsolution from the host mineral by Spear and Selverstone (1983), the slow diffusion of OH, similar densities of all fluid inclusions in such arrays, and similar fracture widths calculated for CO_2 filled inclusions suggest exsolution is not important (Walther and Orville, 1982).

Experimental

In-situ measurements of permeability are about 10^3 times greater than laboratory measurements on rock samples, probably as a result of fractured rocks not being sampled. Values of permeability of crystalline rocks determined from both laboratory and in situ measurements can vary by 4 to 6 orders of magnitude with no systematic variation with depth to several km. Freeze and Cherry (1979, p. 29) listed the range of values of intrinsic permeability for fractured igneous and metamorphic rocks as 10^{-11} to 10^{-7}cm^2. Values of permeability of crystalline rocks determined from both laboratory and *in-situ* measurements have been summarized by Brace (1980, 1984) who concluded that:

1. Values of permeability determined in drill holes (≅2-3 km) in crystalline rocks range from about 10^{-14} to 10^{-9}cm^2,
2. Over some interval in nearly all of the boreholes, permeability was 10^{-14} to 10^{-9}cm^2,
3. Permeabilities inferred from earthquake migration and other large-scale crustal phenomena range, for crystalline rocks, from 10^{-12} to 10^{-10}cm^2, *and are thus about the same as the more permeable intervals in boreholes.*

Brace concluded that, in areas where crystalline rocks extend to the surface, *the pore pressure will be equal to the hydrostatic pressure to depths of at least 10 km.*

Less direct, geophysical methods have also lead to postulates of deep, fluid-filled fractures in the Earth's crust. Nekut *et al.* (1977) suggested that in the crystalline rocks of the Adirondacks a water-filled pore space to a depth of 20 km is required to explain the relatively high electrical conductivity of the crust. Permeabilities inferred from earthquake migration and other large-scale crustal phenomena ranges (Brace, 1980, 1984), for crystalline rocks, from 10^{-12} to 10^{-10}cm^2 which are about the same as the more permeable intervals in boreholes. Crustal permeability in the area of the wastewater injection-induced Denver earthquakes is estimated to be about 10^{-11}cm^2 (Hsieh and Bredehoeft, 1981; Brace, 1984).

Numerical modeling

In modeling of the hydrothermal-meteoric convection cells that result from emplacement of igneous plutons in the crust, Norton and Knight (1977) and Norton and Knapp (1977) used permeabilities of between 10^{-12}cm^2 and 10^{-16} cm^2. These authors concluded that such fluid convection around a pluton can occur at depths up to 20 km. While the factor that causes the fluid flow is the heat from the pluton, their work would imply that the country rocks are already permeable, having connected hydraulic channels over large areas, requiring only a perturbation such as an igneous intrusion to initiate flow.

Walther and Orville (1982) modeled the production and transport of volatiles in the metamorphism of an average pelite and found that grain boundary diffusion is not an effective method of fluid transport. Rather, they concluded that the rocks must fracture because the permeability of the unfractured rock is insufficient to accommodate the flux of volatiles produced during metamorphism. The volatiles escape along fractures between 0.1 and 10 μm in width, but a discrete fluid phase will be present only during a devolatilization reaction.

Walder and Nur (1984) assumed values of intrinsic permeability on the order of 5 X 10^{-16}cm^2 (50 ndarcy) for crystalline rocks with no megascopic fractures at upper to midcrustal depths. This is a lower limit; fractured crust can have permeabilities orders of magnitude higher than this. Walder and Nur (1984) suggested that episodic fracturing and crack healing may be common processes throughout much of the crust, and that this suggestion is supported by detailed studies of the morphology of cracks in exhumed crustal rocks and of crack healing in synthetic laboratory materials.

HYDROSEISMICITY

We propose that in Hydroseismicity, fluids trigger seismicity at focal depths of 10-25 km and that the mechanism(s) share some common features with deeper, as well as shallower, reservoir-induced seismicity. Diffusion of pore pressure is the primary factor that triggers shallow reservoir-induced seismicity by decreasing the shear strength of the rocks by either transient mechanical or longer term chemical means.

Fluid diffusion and transient changes in pore pressure to depths of tens of km will be most effectual in gravity-driven river basins with adequate recharge volumes. Hydrolytic weakening is a general name for various forms of mineral weakening by the addition of water. Presumably, a supply of water for hydrolytic weakening can also be most effectively maintained in the deep crust in the more efficient gravity-driven basins; i. e., in basins with relatively high average stream gradients. The result can be a "hydraulically induced" seismicity trigger that acts somewhere along flow paths such as those shown in Figure 1, a process we term *Hydroseismicity*.

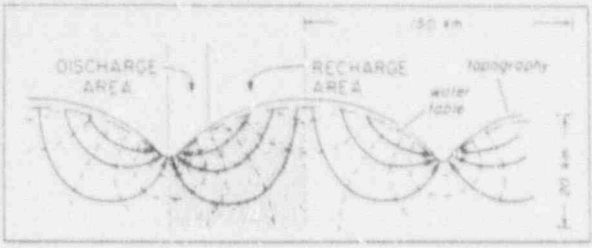

Figure 1. Conceptual groundwater model. Stippled area is one-half of a transverse cross sectional area through a groundwater basin. There will also be a component of flow out of the plane of the diagram in a "down-basin" direction. Note the upward direction of groundwater flow toward the discharge area in both figures, and the greater groundwater flow beneath rivers (after Freeze and Cherry, 1979). Maximum depth of penetration of regional flowlines is hypothesized to be 10-20 km. Figures not to scale.

Conceptual model

A groundwater basin is a hydrogeologic volume of rock with connected and interrelated flow paths (Figure 1). Such a volume may or may not coincide with a physiographic unit. In areas of low physiographic relief the locations of groundwater basin divides cannot yet be extended to great depths with any confidence. Gravity-driven groundwater basins in the eastern United States occupy large (100-200 km wide by several hundred km in length by ≥20 km in depth) volumes of the crust. Groundwater flow within these volumes is hypothesized to be 3-dimensional and anisotropic but conceptually simple. Highlands are recharge areas and lowlands are discharge areas. Discharge areas usually constitute only 5-30% of the surface area of a watershed (Freeze and Cherry, 1979). The direction of deeply penetrating groundwater flow is constrained by the geometry of the boundary conditions of the groundwater basin to be *steep* and *down* at boundaries of recharge areas and steep and up beneath rivers (Figure 1). Although the flow paths shown in the figure are schematic and do not reflect the tensor properties of hydraulic diffusivity, the flowlines indicate the general geometry of the Hydroseismicity conceptual model, and are hypothesized to extend to depths of about 20 km. Focal depth distributions (Figure 2a) suggest that 90% of the earthquakes occur above depths of about 11, 13, 17, and 21 km, in the Charleston, central Virginia, Giles County, and southeastern Tennessee areas, respectively.

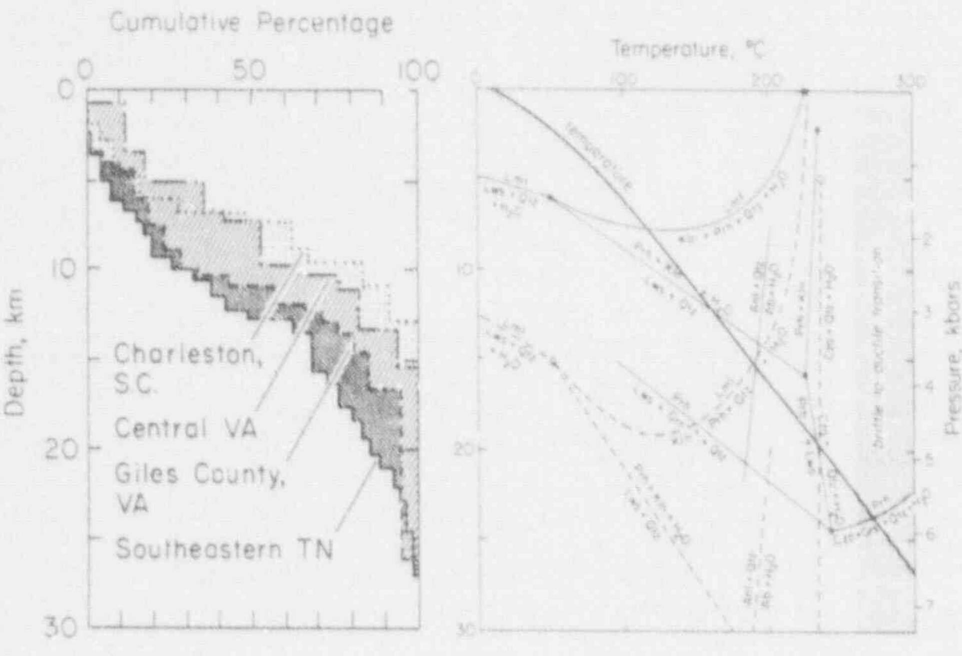

Figure 2. Earthquake hypocenters and mineral phase relations. Left: Depth distribution of hypocenters in southeastern U. S. seismic zones. Right: Depth vs. temperature (heavy curve) and the brittle-ductile transition from Bollinger and others (1985b) with phase relations under lithostatic (solid line, $\rho = 2.5$ gm/cm^3) and hydrostatic (dashed line, $\rho = 1$ gm/cm^3) from Liou (1971c), Perkins and others (1980), and Bruton and Helgeson (1983). Mineral abbreviations are those of Kretz (1983).

Maxima in seismic activity occur a few km shallower than these depths. In the Hydroseismicity model, increased rainfall results in an elevation of the water table and fluid pressure transients that are propagated downward from basin recharge areas. The depth of penetration of flowlines depends, in part, upon the location of the recharge with respect to the geometry of the basin. A high spatial correlation between seismicity and rivers is therefore not necessarily to be expected in our model because the geometry of flow trajectories is from recharge areas (highlands) to discharge areas (rivers), distances of tens of km. The geometry of fluid and pore pressure diffusion in our proposed model implies a *diffuse* distribution of epicenters rather than concentrations along named and mappable, discrete geologic (faults) or geomorphic (rivers) elements. Fracture permeability may be considerably enhanced, however, along major structural elements such as reactivated ductile deformation zones. Thus, the major rivers should not necessarily be expected to be the locus of higher intraplate seismicity.

Nur (1972), Scholz *et al.* (1973), and Whitcomb *et al.* (1973) discussed the dilatancy model of earthquake prediction, and described the role played by the interaction of the tectonic stress field and the fluid pressure field just prior to the actual triggering of movement on a fault. The dilatancy model uses changes in the traveltime ratios of seismic waves to infer changes in the physical properties of the rock in focal regions prior to an earthquake. The changes in the physical properties of the rock are assumed to be caused by local progressive fracture of the rock associated with the opening of crack-pore space (dilation), a decrease of seismic velocities, and a dramatic increase in rock permeability (Rummel *et al.*, 1978). The assumptions are based in part on dilatancy experiments (Crouch, 1970; Friedman, 1974). Lockner and Byerlee (1978) suggested that the velocity anomalies might be biased by the difficulty in picking arrival times of emergent events. In any case, lack of changes in traveltimes does not mean that dilatancy is non-existent. The permeability and porosity of fractured rocks is a function of fluid pressure (Snow, 1968a). Reviews and practical aspects of flow in a fractured medium can be found in Snow (1965; 1968a,b; 1969), Bear and Braester (1972), and Streltsova-Adams (1978).

Howells (1974) using a simple uncoupled diffusion equation estimated that instantaneous pressure changes at the surface would have a significant effect on fluid pressures at depths of 2.5-7.5 km after 100 days, and at depths of 10 km or more after several hundreds of days. We have repeated and extended the calculations of Howells (1974) to investigate the magnitudes of pressure transients for an assumed permeability range for different crustal depths (or, equivalently, along flowlines of different lengths if the flowline is not vertical). Use of the specific storage coefficient in the one-dimensional calculations of Figure 3 and Figure 4 accounts for the elastic behavior of the framework surrounding the fluid as well as the compressibility of the fluid without requiring details of deformation. Details of individual fracture geometries and permeabilities will always be imperfectly known and may not justify use of discrete fracture models (Huyakorn and Pinder, 1983). Each curve on Figures 3 and 4 was generated from the sum of step functions of the form (Howells, 1974):

$$\frac{p(z,\ t)}{p(0,\ t)} = 1 - \operatorname{erf}\left[\frac{z}{2\sqrt{ct}}\right] \qquad (1)$$

where $c = k\rho g/\mu S_s$, erf is the error function, z is depth (or path length along a flowline), t is time, k is intrinsic permeability, ρ is density, g is the acceleration of gravity, μ is dynamic viscosity, and S_s is specific storage. Isothermal conditions and a constant viscosity were assumed. The effect of temperature on viscosity is shown in Figure 4.

In addition to one-dimensional calculations, we have investigated the behavior of fluid pressure transients using a two-dimensional Galerkin finite element model that can simulate fluid transport in deformable fractured-porous isothermal media where the coupled governing equations of deformation and fluid flow for an unfractured porous medium are (Biot, 1941; Huyakorn and Pinder, 1983):

$$G\frac{\partial^2 u_i}{\partial x_j \partial x_j} + (\lambda + G)\frac{\partial^2 u_j}{\partial x_i \partial x_j} - \frac{\partial p}{\partial x_i} = -F_i - \frac{\partial \sigma_{ij}^0}{\partial x_j} \qquad (2)$$

The incremental displacement of the solid skeleton is u_i and the governing equation for fluid flow is

$$\frac{\partial}{\partial x_i}\left[\frac{k_{ij}}{\mu}\left(\frac{\partial p}{\partial x_j} + \rho_f g_j\right)\right] = \varphi\beta\frac{\partial p}{\partial t} + \frac{\partial}{\partial t}\left(\frac{\partial u_i}{\partial x_i}\right) \qquad (3)$$

where λ and G are Lame's constants, F_i is the body force per unit volume of the medium, σ_{ij}^0 is the initial effective stress, and other quantities are as defined for equation (1). This formulation assumes that the solid material making up the aquifer framework is incompressible, but that the fluid is slightly compressible (Huyakorn and Pinder, 1983). We have not yet

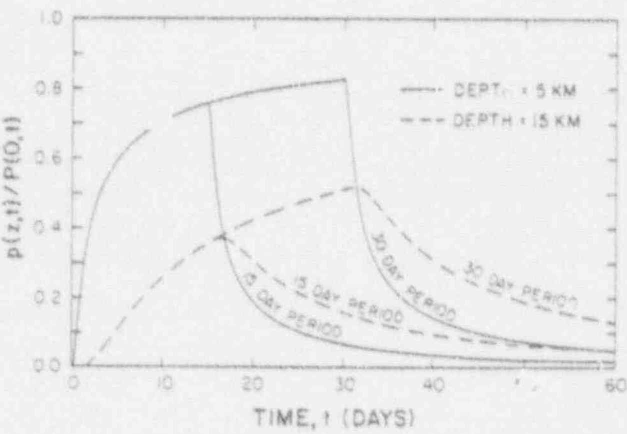

Figure 3. Ratios, $p(z,t)/p(o,t)$, of increase in fluid pressure at surface, $p(o,t)$, to that at depth, z, after a time, t. $k = 10^{-9}$ cgs, $S_s = 10^{-10}$ cgs.

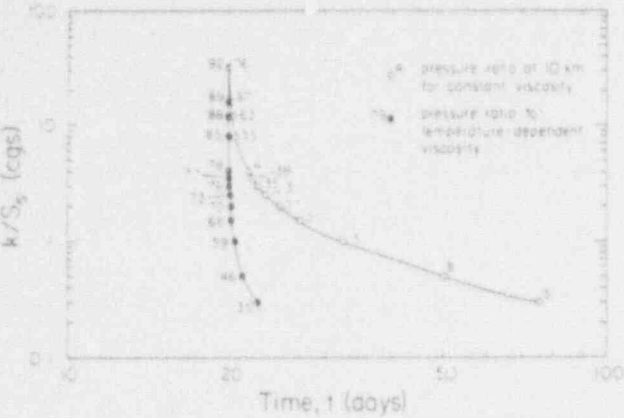

Figure 4. Effect of temperature-dependence of viscosity on magnitude of maximum pressure ratios for various values of k/S_s plotted against arrival time of the pressure maximum at a depth of 10 km. Assumed 20-day elevation of water table and a depth and temperature such that viscosity, μ, is lowered one order of magnitude. Right-hand curve: $\mu = 9.8 \times 10^{-3}$ cgs. Left-hand curve: $\mu = 9.8 \times 10^{-4}$ cgs. Values of maximum fluid pressure ratios in percent are shown near plotted points on each curve. Note increases in fluid pressure ratios for $\mu = 9.8 \times 10^{-4}$ cgs. Calculations from diffusion equation of Howells (1974).

applied the diffusion solutions of Rice and Cleary (1976) that take into full account the compressibility of the fluid and solid constituents.

Several conclusions can be drawn from the one- and two-dimensional analyses. In hindsight some of these are obvious, but we feel that others are not.

1. Intrinsic permeabilities inferred from earthquake migration and other large-scale crustal phenomena range, for crystalline rocks, from 10^{-12} to 10^{-10}cm² (Brace, 1984). Using any of those values for the entire crust, it is clear from the modeling results that such permeabilities are adequate to allow diffusion of pressure transients from recharge areas to depths of 20 km over time intervals appropriate for our hypothesis; values of 10^{-14} to 10^{-8}cm² are in the range of those reported by Brace for nearly all permeable fracture zones encountered by the drill in crystalline rocks. We conclude that a crust with fracture permeability in the range of 10^{-14} to 10^{-8}cm² can diffuse meteoric water and pore pressure transients to the upper and middle crust.

2. Pressure ratios obtained from 2-dimensional modeling of coupled flow and deformation compare favorably with those obtained from the 1-dimensional analyses. If a fraction of a bar is sufficient to trigger seismicity in some regions, as suggested by Nava and Johnson (1984), then such magnitudes can be transmitted along fracture zones and, combined with hydrolytic weakening, result in rock failure.

3. For coupled flow and deformation, as the value of Young's modulus is increased above 10^9 mks, fluid pressure ratios increase markedly.

4. Anisotropy in the intrinsic permeability tensor significantly affects the distribution and magnitude of deep crustal fluid pressure transients. Major plate collisions in the southeast U.S. resulted in the development of anisotropic fracture permeability. The most obvious reasons for anisotropic crustal volumes in the Southeast are the Taconic plate collision and the Late Alleghanian oblique collision, the latter resulting in major dextral strike-slip motion along inclined faults that were later reactivated and that controlled the geometry of the Mesozoic rift basins. Major portions of the crust in the southeast U.S. have been affected by extensional rifting. Intrinsic permeabilities would be expected to be higher in a vertical direction than in a horizontal direction.

5. Gravity-driven groundwater basins with higher average surface stream gradients supply more water per unit time per unit volume to a fractured crust.

6. The spatial and temporal convolution of pressure transients at a given crustal depth can result in significant increases in fluid pressure above what would result from a rise in the water table at a single recharge location. The spatial and temporal basin recharge function can thus have an important effect on the fluid pressure at depth because of the opportunities for linear superposition of pressure transients having different origin times at different recharge areas. For example, at a given depth, z, transient increases in pressure can arrive at the same time via completely different flow paths from different recharge areas in the same basin as a result of recharge functions with different locations, intensities, durations, and origin times.

7. The magnitude of a pressure transient at crustal depths depends upon k/S_s, the ratio of intrinsic permeability, k, to specific storage, S_s (Figure 3). An increase in the ratio results in an increase in the magnitude of the pressure transient.

8. Pressure ratios increase with depth as the surface area of transient elevation of the water table increases.

9. The duration of a rainy period is important only insofar as it results in an increased elevation of the water table.

10. The maximum fluid pressure at depth increases with increasing duration of elevation of the water table (Figure 3).

11. If the permeability decreases with depth, the magnitude of a pressure transient can be sustained by simply increasing the duration of the period of elevation of the water table (Figure 3).

12. The pressure maximum is delayed with respect to the onset of the elevation of the water table as depth increases (Figure 3).

13. Intense rainy periods of short duration may be less important as seismicity triggers than long rainy periods of less intensity because of the opportunity for the subsurface convolution of pressure transients, and because the maximum fluid pressure increases with increasing duration of the elevation of the water table.

14. Since the dynamic viscosity of water decreases with increasing temperature (Haar *et al.*, 1984; Wahl, 1977), for a given value of intrinsic permeability, pressure ratios are markedly increased at depth by a decrease in viscosity (Figure 4). For one data set in Figure 4 the viscosity is held constant for all assumed values of k/S_s. Times of pressure maxima are plotted versus k/S_s with the values of the maximum pressure ratios shown beside the plotted points. For a second data set the viscosity is assumed to be order of magnitude lower. The lower viscosity results in higher maximum pressure ratios as shown beside the left-hand curve. Higher geothermal gradients thus tend to increase the magnitudes of fluid pressure transients in the upper crust for a given depth and intrinsic permeability.

15. For ratios of k/S_s greater than about 0.5×10^{-3} (Figure 4) maxima in pressure transients are not delayed significantly after the end of a rainy period; for ratios less than about 10^{-3} maxima can be delayed significantly with respect to the end of a rainy period (with respect to the end of a period of elevation of the water table), although values of maximum pressure ratios are considerably decreased. Comparison of values of pressure ratios for the two curves indicates that the temperature dependence of viscosity has more effect on pressure ratios for smaller values of k/S_s.

SEISMICITY, FRACTURE PERMEABILITY, AND MAJOR RIVER BASINS IN THE SOUTHEASTERN U.S.

Seismicity

The regional spatial distribution of seismicity in the southeastern U. S. is nonuniform and nonrandom. Hadley and Devine (1974) contoured historical earthquake frequency for the southeastern U. S. Their results are generally consistent spatially with the modern, instrumentally determined earthquake epicenters of Sibol and Bollinger (1984) shown in Figure 5. Four active areas have been identified: the South Carolina-Georgia seismic zone, the Central Virginia Seismic Zone, the Giles County, Virginia, Seismic Zone, and the Eastern Tennessee Seismic Zone. In each of these areas, focal mechanisms indicate steep to vertical fault planes (Talwani, 1982; Bollinger and Wheeler, 1982; Bollinger and Wheeler, 1985; Johnston *et al.*, 1985; Bollinger *et al.*, 1985a), and, commonly, strike slip motion. The spatial distribution of both historical and modern seismicity in the southeastern U. S. is characterized by diffuse alignments and clusters (Bollinger, 1973; Talwani, 1982; Sibol and Bollinger, 1984). Locally, earthquakes in the Piedmont province tend to be *scattered*, i.e., *diffuse*; those in the Coastal Plain tend to *cluster*. Epicenters in the Valley and Blue Ridge tend to be *aligned*.

Ninety percent of all hypocenters ($M > 0$, depth error estimates < 5 km) in the southeastern U. S. are at depths of less than 20 km. Focal mechanisms for southeastern U. S. earthquakes exhibit steep to vertical fault planes and, commonly, strike-slip motion. Seismicity in the southeast U. S. is characterized spatially by steeply-dipping nodal planes and *diffuse* alignments and clusters of epicenters. These are all characteristics that are consistent with a steep to vertical fracture fabric that is currently being reactivated by pore pressure diffusion from the surface recharge of groundwater basins.

Major rivers and rifted crust

Major rivers, average river gradients, epicenters, and buried and exposed rift basins in the southeastern U. S. are shown in Figure 5. Epicenters in the *South Carolina-Georgia Seismic Zone* tend to parallel the Broad, Congaree, Cooper river system (Figure 5). The major Charleston, S.C., earthquake of August 31, 1886 was located near the center of the southeastern part of the Santee ACE (Ashley-Cooper-Edisto) River Basins. Together these basins comprise the largest groundwater basin in the State, approximately 18,700 mi² (Snyder, 1983). Groundwater basin boundaries are defined primarily on the basis of surface physiography; basin boundaries for deep regional groundwater flow in the Piedmont may approximate these surface definitions. In the Coastal Plain sediments, however, boundaries are not well known below a depth of about 100 m. Coruh *et al.* (1981) imaged a buried Mesozoic rift basin (Vibroseis Line VT-5 on Figure 5) beneath the Atlantic Coastal Plain near the location of the major Charleston, S. C. earthquake of 1886. In addition, the density of dikes in the Georgia-South Carolina seismic zone is noticeably greater than that observed elsewhere (Ragland *et al.*, 1983); the dominant direction of the dikes is northwest, parallel to the seismogenic zone, suggesting that deep crustal intrinsic permeability could be considerably greater in the northwest direction, parallel to the axis of the Santee-ACE basin.

The *Central Virginia Seismic Zone* is located in the James River groundwater basin. The zone is bisected by the James River and is approximately coincident with the "Triassic Lowlands" formed by the Scottsville and Richmond Mesozoic rift basins. The spatial distribution of the epicenters in this zone is diffuse. The east-west extent of the instrumentally de-

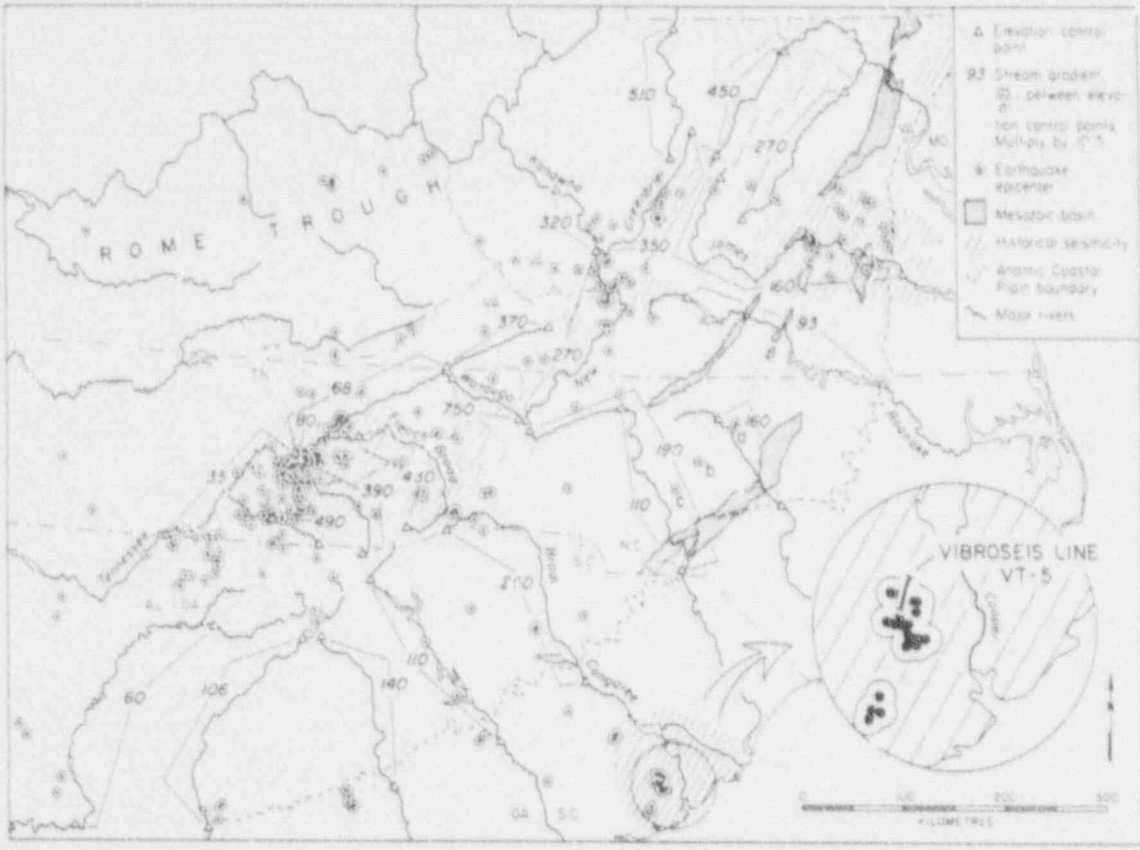

Figure 5. Major rivers, historical seismicity, and instrumentally located epicenters. Hatchuring shows distribution of Hadley and Devine (1974) of > 16 historical earthquakes per 10^4 km^2 with intensity > MM III during the period 1800-1972. A buried Mesozoic basin is clearly imaged on Vibroseis line VT-5 in the area of the major Charleston, SC, earthquake of 1886. Epicenters labeled a, b, and c in North Carolina parallel exposed early Mesozoic basins (EMb) and lie on an extension of the trend of EMb in central Virginia.

termined epicenters is approximately the same as that of the exposed Early Mesozoic basins.

The Culpeper rift basin in northern Virginia is traversed by the Potomac River. Although recent seismicity is absent, the area has been historically active (Bollinger, 1973).

The absence of extensive seismicity in the Roanoke River groundwater basin which drains parts of Virginia and North Carolina is attributed herein primarily to the lower elevations of the headwaters of the Roanoke River, and consequently to a lower potential for pore pressure fluctuations in the upper crust. Southwest of the Roanoke River basin, the epicenters marked a, b, and c in North Carolina (Figure 5) are aligned with exposed Mesozoic rift basins to the north and south.

The New River in Virginia and the Greenbriar and Kanawha Rivers in West Virginia approximately bisect the *Giles County Seismic Zone* (Figure 5). The seismicity is within the crystalline basement in approximately vertical zones. An early Paleozoic passive margin extended the length of the Appalachians and probably included Giles County, Virginia, and eastern Tennessee (Bollinger and Wheeler, 1982). Reflection seismic data of Gresko and Costain (1985) support the suggestion of Bollinger and Wheeler (1982) that the seismicity is associated with Eocambrian rifting within the crystalline basement beneath about 6 km of Paleozoic shelf strata. The rift geometry suggested by the reflection seismic data in Giles County may be similar to that developed in the Rome Trough, a late early Cambrian major rift structure involving crystalline basement that was reactivated during the Mississippian. The northwest border fault is in crystalline basement and is apparently still active (Figure 5).

Four rivers, including the French Broad, that discharge into the Tennessee River comprise the principal part of the drainage system over the hypocenters in the *Eastern Tennessee Seismic Zone* (Figure 5), the spatially largest of the southeast seismic zones. Few epicenters and no major rivers are located immediately northwest of the Tennessee River.

Fracture permeability in the southeast U. S.

The most important fracture permeability in the crystalline rocks of the southeastern U. S. is along zeolite-lined fractures developed primarily during the latest rifting event. These minerals occur as euhedral fracture fillings and in open vugs, indicating the fractures were indeed open. Such fractures are virtually certain to be associated with a substantial flow of groundwater (Secor *et al.*, 1982, Figure 3). If fluid pressure in a fractured crustal volume is more nearly hydrostatic (from a fluid of density $\rho = 1$ gm/cm^3) rather than lithostatic (rock density $\rho = 2.67$ gm/cm^3), the maximum depth at which zeolite mineral assemblages can form in the fractures is 15-20 km.

As much as one-third of southeastern North America may be underlain by extensive rifting of Eocambrian and Mesozoic age that developed near a passive margin (Wheeler, 1983). The zone of Eocambrian Iapetan rifting in the southern Appalachians was at least 400 km and probably more than 500 km wide. Significant Triassic faulting in the eastern U. S. mimics that Proterozoic rifting (Rankin, 1984), and produced rift basins that are the end products of extensional megascopic crustal strain. Steep to vertical extension fractures in rifted upper crust are expected to be pervasive. The border faults of rift basins are known to have been localized in some instances by reactivation along ancient mylonite zones (Ratcliffe and Burton, 1985; Ratcliffe *et al.*, 1986). It is presumed that this latest Mesozoic rifting event is primarily responsible for the development of the new and reactivated listric to steep to vertical fracture fabric commonly encountered in the southeast U. S. in near-surface drilling. One manifestation of extensional strain at a passive margin is the injection of steep to vertical dikes that penetrate the crust from the upper mantle during the later stages of crustal extension (Ragland *et al.*, 1983, and references therein). Such macroscopic strain is observed in the southeastern U. S. from the coast to nearly the folded Appalachians (Ragland *et al.*, 1983).

The crystalline rocks of the southeast exhibit widespread fractures lined or filled with the zeolites laumontite and prehnite in association with calcite, epidote, K-feldspar, albite, quartz, siderite, fluorite, gypsum, thaumasite, clay minerals, and hematite. These minerals occur as euhedral fracture fillings and in open vugs, indicating the fractures were/are open. Such fractures occur in almost every rock type present in the southeast; metamorphic rocks in Virginia (Bobyarchick and Glover, 1979) and South Carolina (Gilbert *et al.*, 1982; Butler, 1984) in the postmetamorphic granitoids (Privett, 1973; Speer *et al.*, 1980, 1981; Secor *et al.*, 1982); the Mesozoic diabases in North Carolina (Furbish, 1965) and Virginia (Toewe, 1966); and the Mesozoic sedimentary rocks (Bobyarchick and Glover, 1979). Fractures spatially associated with the Mesozoic diabases are believed to be contemporaneous with the intrusions (Butler, 1977). Direct dating of fracture-lining laumontite is believed to yield minimum ages and the true ages may be not less than 150 Ma (Gilbert *et al.*, 1982; Secor *et al.*, 1982). Contemporaneous with this fracturing is the pervasive redistribution of U in the postmetamorphic granitoids (Sinha and Mertz, 1978, in Speer *et al.*, 1981) which also suggests fluid circulation.

Fracture mineralogy suggests that fracture permeability is not restricted to the near-surface, but is characteristic of the entire crust. The presence of laumontite and the absence of lawsonite, wairakite, and grossular in these fractures can be used to obtain an approximate maximum pressure for the fracture filling and provide an estimate of depths for significant fracture permeability in the southeastern U. S. Liou (1971a, b) showed that the stability range of laumontite is about 1.5 to 3.0 kbar and 150 °C to 275 °C. More pertinent for the fractures in the southeastern U. S. are some of the phase relations (Figure 2) for laumontite from Bruton and Helgeson (1983) and the CASH system by Perkins *et al.* (1980). The phase relations drawn for lithostatic pressure assume that the rock density is 2.5 gm/cm^3. Those for hydrostatic pressure assume that the water has a density of 1 gm/cm^3. Of most interest is the stability of laumontite on its own composition: 2 laumontite = prehnite + kaolinite + 3 quartz + 5 water. *If pore pressure is closer to the hydrostatic rather than the lithostatic pressure, the depth to which laumontite can form is increased.* Using the temperature gradient from Bollinger *et al.* (1985b), the maximum depth at which laumontite can form in fractures is 15 km. Prehnite-filled fractures could form at depths as great as 24 km, defined by the equilibria: prehnite = clinozoisite + grossular + quartz + water for lithostatic conditions. The phase relations shown in Figure 2 suggest that minerals in the veins may have formed over a range of conditions, with such minerals as epidote and albite forming earlier and at higher temperatures than the laumontite.

Thus, it is feasible that the zeolite-lined fractures that are commonly encountered in cores from the southeastern U.S. are comparable to those constituting fracture permeability of the crust at depths up to at least 20 km. At greater depths and temperatures, especially during high grade metamorphism, the physical conditions will be such that the rocks will flow plastically and thus lose their high permeabilities. This would be the lower limit of earthquake occurrences by the Hydroseismicity model. Using the depth for a cumulative 90% occurrence of earthquakes in a region as an estimate of the depth of the brittle-ductile transition, Bollinger *et al.* (1985b) report the brittle-ductile transition depth as 20 km in the Valley and Ridge/Blue Ridge and 13 km in the Piedmont/Coastal Plain. Based on heat flow models, these depths correspond to temperatures of 250°C and 160°C respectively.

Measurements of permeability in a borehole can vary by 4 to 6 orders of magnitude with no systematic variation with depth to several km. *In-situ* permeability measurements have been made at the Savannah River Laboratory, South Carolina, located in the vicinity of a Mesozoic fault basin. Marine (1966, 1967) reports permeabilities of $10^{-14}cm^2$ to $10^{-8}cm^2$ in the crystalline rocks of the basement at depths of 30-650 m. Permeability measurements in two 1.1 km deep holes drilled in granites and amphibolites at Monticello, South Carolina are about 10^{-11} cm^2 at three locations but as little as $10^{-13}cm^2$ at a fourth (Zoback and Hickman, 1982).

TRIGGER MECHANISMS

We have three choices for the trigger mechanisms for our hypothesis; seismicity can be induced by raising the shear stress, by reducing the shear strength of rocks, or both. We believe the following play an important role in the specifics of the hydroseismicity trigger mechanism:

1. Mechanical: diffusion of fluid pressure from recharge areas to a fractured crust, and

2. Chemical: the dissolution or reaction of minerals with the pore fluid, possibly enhanced by an existing stress, that results in weakening of the rocks along preexisting fractures.

The relative importance of these two is not well defined at the present; however, in the generally uniform stress field of the eastern U.S., it is more likely that failure would occur by first lowering the strength of the rock by chemical means followed by the mechanical effects of the fluid pressure transients.

Mechanical

Hydroseismicity assumes seismicity is triggered by transient increases in the height of the water table that result in the diffusion of pore pressure transients to hypocentral depths from recharge areas of groundwater basins. The crust is assumed to be in a prestressed state close to failure. Under this assumption, the magnitude of the required triggering stress can be a small fraction of a bar and such small transient increases in fluid pressure can be transmitted tens of km in a fractured crust. The magnitude of the fluid pressure transient from a single storm will depend not only on the amount of precipitation, but also on the already present moisture content of the unsaturated zone above the water table. During the warmer summer periods, the soil dries. Occasional rainstorms may result in short-term rises in soil-moisture content, but there is generally no significant groundwater recharge. Exceptionally heavy summer rains replenish the depleted soil moisture and raise the water content above field capacity, create a wave of infiltrated water that passes downward through the soil-moisture zone and past the roots of plants that would otherwise retain the water if the infiltration rate were lower, and thus recharge the groundwater reservoir causing the water table to rise (Fetter, 1980).

The moisture content above the water table is a time-varying function of infiltration rate, hydraulic conductivity, evapotranspiration, etc. such that the fluid pressure below the water table resulting from so many inches of rainfall will be greater than that simply computed from the weight of water before infiltration because the soil may already be, for example, 50% saturated at the time of infiltration, and more of the precipitation is able to reach and raise the height of the water table. Field determinations of fluid pressures at depth after adding water are often observed to be greater than that expected simply from the weight of water added (personal communication, George Pinder, 1986); thus flow equations in the unsaturated zone are nonlinear. Thus the unsaturated zone may play an important role with regard to both the magnitude and time rate of change of the pressure transient that is propagated downward from a recharge area.

Ambient stresses at hypocentral depths are on the order of kilobars; crustal earthquakes relieve only part of this stress and recent determinations indicate stress drops of tens of bars or less (Simpson, 198_). For an intraplate earthquake of m_b = 5.2, M_s = 5.0, the average stress drop is 10 bars; for M_s = 7, the average stress drop is only about 35 bars (Nuttli, 1983).

The possible origins of tectonic stress in the eastern U. S. have been reviewed by others. A stress direction of N 60° E - N 70° E appears to be the overwhelmingly dominant regional direction of the maximum horizontal stress in the eastern U. S., and is related to the tectonics of the North American Plate, not to local perturbations (Haimson, 1978; Zoback *et al.*, 1985a; Zoback and Zoback, 1985). Higher horizontal stress in the upper km of the crust is a common, but unexplained, characteristic of the regional stress field (Haimson, 1978). In the Michigan basin, the vertical and maximum horizontal stresses are close in magnitude over the depth interval 3-5 km so that both strike slip and normal faulting are favored (Haimson, 1978).

The role of fluid pressure transients as a seismicity trigger would be important in a gravity-loaded crust within which the horizontal lithostatic stress, σ_x, is just enough to prevent lateral strain, and σ_x is related to the vertical, lithostatic stress, σ_z, by Poisson's ratio, ν:

$$\sigma_x = \left[\frac{\nu}{(1-\nu)}\right]\sigma_z$$

If ν = 0.5, the stress is hydrostatic. If ν = 0, the rock is perfectly rigid and requires no lateral restraint, i.e., σ_x = 0. The hydrostatic fluid pressure to 20 km for a given geothermal gradient must be determined by integration using the equation of state of water (Haar *et al.*, 1984). If a fractured volume with vertical hydraulic continuity were in place, then for ν = 0.28 in a gravity-loaded crust, hydrostatic fluid pressure alone would provide most of the stress required to start crack dilation, which might lead to faulting. For ν = 0.23, the static hydrostatic fluid pressure would provide *more* than the required minimum stress for crack dilation, without requiring transient increases in fluid pressure. Stock *et al.* (1985) report in situ values of the least horizontal stress at Yucca Mountain in the rifted Basin and Range province that would be about equal to the fluid pressure as measured from a water table at the surface. Focal mechanisms indicate predominantly strike-slip motion on steeply-dipping nodal planes, consistent with the Yucca Mountain stress regime (Stock *et al.*, 1985), and favoring motion on preexisting vertical planes of weakness. The crust is able to relax, however, probably to different degrees at different depths, and a relation $\sigma_v \approx \sigma_h$ is more plausible.

At Monticello Reservoir, South Carolina, Haimson and Zoback (1984) found that the least horizontal stresses were less than the calculated vertical stress, but that the difference between the horizontal principal stresses was too small to be responsible for either reverse or strike-slip faulting. They suggested that, because of pervasive fracturing, in situ stress determinations over a depth range of 150-400 m may represent the vertical lithostatic stress rather than a horizontal stress. Thus, in situ determinations of horizontal stresses in deeper holes in the eastern United States remains an important objective that may help to correlate in situ stress determinations with the strike-slip motion along steep to vertical faults (permeable fracture conduits) indicated by earthquake fault mechanisms, and consistent with the Hydroseismicity hypothesis which requires such permeability to depths of 15-20 km.

As suggested by others, the crust in many places is stressed to near failure. We are suggesting that the introduction of pressure transients from recharge areas of groundwater basins and the longer-term effects of hydrolytic weakening are sufficient to trigger intraplate earthquakes in the southeastern U.S.

The rocks on each side of an approximately vertical fault plane are kept apart partly by a stress within asperities that separate the fracture walls, and partly by the fluid pressure that fills the fracture space (for example, Hubbert and Rubey, 1959; Freeze and Cherry, 1979). If the total stress across the vertical fracture is assumed to remain constant, then an increase in fluid pressure results in a decrease in intergranular stress. An increase in the volume of water (load) in rivers initially tends to increase intergranular stress at hypocentral depths; pore pressure diffusion upward into the river from below tends to decrease it at these same depths. Although changes in intergranular stress are probably small, the interplay of both factors may need to be considered in, for example, the New Madrid area. In addition, an increase in head in the discharge area (river) would propagate a pressure pulse back toward the recharge area. Rivers receive a fraction of their volume from below the river (Figure 1). The correlations between rivers, rainfall, and seismicity reported by others appear to be consistent with common groundwater flow models that incorporate variations of flow (diffusion of pore pressure, p) into rivers from below. Nava (1983) and Nava and Johnston (1984) noted that when the Mississippi River is at its maximum stage, earthquake frequency in the New Madrid area increases some 6 to 9 months later; they suggested that this delay reflects transient pore pressure changes at depth. The equivalent pressure change from maximum to minimum flood stage is 0.9 bar; however, when the river level changes corresponding to a change of 0.3 bars in a relatively short time, there is an almost instantaneous change (< 1 month) in earthquake frequency. Conversely, a sudden drop in the river results in a sudden decrease in seismicity and vice versa. Larger magnitude earthquakes (m > 2) are influenced by stage maxima and minima, with a greater number of larger magnitude earthquakes after (> 1 month) a high stage. Nava's (1983) results suggest that a fraction of a bar is sufficient to trigger seismicity in the rifted crust of the New Madrid area.

Simpson (1985) noted that episodes of accelerated stress increase may be responsible for triggering natural seismicity. The triggering of earthquake swarms by water level changes of less than one meter (equivalent to < 0.1 bar) at some sites of reservoir-induced seismicity in itself supports the idea that seismicity can be triggered on faults that otherwise remain stable even at stress levels close to failure (Simpson, 1985). His studies at Nurek Reservoir in central Asia demonstrated that rapid changes in water level as small as 20 cm could rapidly lead to increased seismicity. In the eastern U. S., we suggest that such small transient increases in fluid pressure are sufficient to trigger earthquakes in a crust already prestressed or stress-corroded to near-failure.

Chemical

In addition to the mechanical effects of the transient fluid pressures in fractures, changes in the physical properties of the rocks can be brought about by the slower introduction of new fluid into the fractures. These changes can be accomplished through purely chemical means: chemical dissolution or reaction of the minerals in response to changes in the pore fluid composition, static pressure, or temperature (corrosion). This corrosion can weaken the rock directly or produce secondary minerals (corrosion products) that can lubricate the faults. Talwani and Acree (1985) noted that the mechanical effects of pore pressure control the spatial and temporal pattern of reservoir-induced seismicity, but that the actual onset of seismicity may be influenced by the chemical effect of water reducing the coefficient of friction by developing clay minerals that fill preexisting fractures.

The changes in physical properties of the rocks can also be accomplished by chemical effects (corrosion) acting together with mechanical forces, causing strength failure in cases where mechanical influences acting alone or chemical influences acting alone would be ineffective. Combination of static stresses, either residual or applied, and mineral dissolution or reaction would produce an effect analogous to stress corrosion cracking. Combination of repeated stresses and mineral dissolution or reaction would result in a failure analogous to corrosion fatigue. The reduced strength of the rock in these cases is a result of the chemical attack on defects induced in the minerals by the stress. The defects are more chemically reactive than the more ordered areas of the mineral and are the sites of strength-reducing corrosion. The defects can also act as pathways for the reacting species. The scale of the defects can range from the atomic (i.e., dislocations, etc.) through the microscopic (microfractures) to the megascopic (fractures).

The presence of fluid, such as water with and without other dissolved species, has invariably been found to decrease remarkably the strength of rocks in many experiments (Griggs, 1967). The primary corrosive agent responsible for the strength reduction of silicate minerals is water (Scholz, 1968; Charles, 1959; le Roux, 1965). The presence of trace amounts of water dramatically lowers the transition temperature for the onset of dislocation creep in quartz (Voll, 1976; Tullis and Yund, 1980; Sibson, 1984). For the quartz/water system, the strong silicon-oxygen bonds are replaced with much weaker hydrogen bonds (Atkinson, 1982). Charles (1959) studied the static fatigue of a number of silicates and oxides and found them to be quite similar in behavior, and that static fatigue is due to hydration of the silicon-oxygen bond. Griggs (1967) discussed other types of weakening due to the presence of water, and le Roux (1965) and Stuart and Anderson (1953) have proposed different types of corrosion reactions. Given the potential variety of minerals, fluid compositions, and physical conditions involved, differing reactions can be expected.

Wintsch and Dunning (1985) noted that the density of dislocations in quartz is generally low, probably too small to affect chemical processes; however, dislocation densities may exceed 10^{11} line cm^{-2} in a small fraction of most quartz grains at tangles, pileups, and subgrain walls. They pointed out that this high density of dislocations may significantly increase the dissolution of quartz at those sites. This process they call strain solution and, unlike pressure solution, may occur under hydrostatic pressure (Wintsch and Dunning, 1985). As the high density of dislocations results from stress, past or present, the process shares features in common with stress corrosion and fatigue corrosion. The resulting high activity of aqueous silica would establish chemical potential gradients of SiO_2 that would drive the diffusive mass transfer of SiO_2 out of a fault zone (Wintsch and Dunning, 1985).

These observations on the influence of water and dislocations on the strength of rocks lead to the commonly used term hydrolytic weakening, which combines specific elements, in varying degrees, from what we term here corrosion, stress corrosion, and corrosion fatigue. Hydrolytic weakening depends largely on temperature, pressure, and the presence of microfractures (Paterson, 1986 and references therein; Holland, 1967; Robin, 1978; Fournier and Potter, 1982; Griggs, 1967). Kirby and Kronenberg (1984) noted that microfracturing, the primary mechanism of creep in brittle rock at low temperatures and pressures, is essential to the hydrolytic weakening process and the introduction of (OH) into quartz. The hydrolytic weakening experiments of Briggs and Blacic (1964) and Blacic (1965), and Ord and Hobbs (1986), indicated a substantial penetration of water-related species into quartz from a hydrous environment at 1000-1600 MPa and 1100-1300°K; however, if cracking of specimens is rigorously avoided, no significant penetration of water in quartz can be detected (Paterson, 1986). Mineral defects and fractures of all sizes are therefore essential as reaction species pathways and reaction sites.

Dissolution and reprecipitation of minerals in fractures

Rimstidt and Barnes (1980) showed that the rate of quartz precipitation decreases drastically as temperature increases. Angevine *et al.* (1982) suggested that chemical lithification (cementation) occurs on an active fault between earthquakes. They assume a porous, water-saturated aggregate of quartz grains; pressure solution causes dissolution at the highly stressed grain contacts, and the resulting solute is precipitated nearby as a cement. Angevine and Turcotte (1983) suggested that porosity reduction by pressure solution may play an important role in the lithification of sandstones and limestones, resulting in a decrease of permeability with porosity, hindering fluid migration, and generating excess fluid pressures.

Takahashi *et al.* (1986) presented experimental results on the dissolution rate and on toughness reduction of granite in high-temperature water up to 350 °C that support our basic idea of weakening. They illustrated exponential increases in intrinsic permeability as a function of time over a period of 6 hours for temperatures ranging from 150 °C to 350 °C. Rates of increase of permeability are considerably higher at 350 °C than at 250 °C.

Stress concentrations at asperities should increase the amount of mineral material that can be taken into solution because the formation of microcracks leads to more surface area as well as opportunities for hydrolytic weakening. Solution transfer of a mineral involves (Durney, 1976):

1. Pressure enhanced dissolution at grain contacts,
2. Solute transport through an intergranular fluid film, and
3. Precipitation of solute on a free-grain surface.

As noted by Angevine *et al.* (1982) the history of the solute after pressure solution is not completely known. They suggested that it is immediately reprecipitated locally to decrease permeability and raise fluid pressures; however, there does not seem to be a local mechanism comparable to pressure solution to reprecipitate the mineral. That is, the rate of mineral dissolution can be sped up by localized stress concentrations at asperities, but we may need another mechanism to reprecipitate it. Furthermore, about 80% of the rock volume of a groundwater basin is in the recharge volume. Therefore, in most (80%) of the rock volume of the groundwater basins the temperature will increase in the (downward) direction of flow away from recharge areas; thus, quartz that is placed into solution by stress corrosion in a recharge rock volume (Figure 1) may not precipitate until the smaller (discharge) rock volume is reached.

An illustrative example is quartz, a common mineral in the crystalline rocks of the southeastern U.S. Paterson (1986) developed the thermodynamic model of Doukhan and Trepied

(1985) for water solubility in quartz, based on the $4H^{+} = Si^{+4}$ substitution, i.e., four hydrogens substituted for a silicon (McClaren *et al.*, 1983), and included the variation in oxygen fugacity as well as the solubility of quartz in water. Ord and Hobbs (1986) suggested a less regular structural substitution of (OH) in quartz in defects, rather than a $4H^{+} = Si^{+4}$ substitution.

The solubility of quartz in pure water under different conditions of temperature and fluid pressure is shown in Figure 6. Solubility is plotted on the vertical axis. Fluid pressure is assumed to be hydrostatic in a crust with hydraulic continuity to the surface, and was determined by vertical integration to the desired depth, z, of the density of water over the temperature range defined by the geothermal gradient. Density was obtained from the equation of state of water (Haar *et al.*, 1984). The solubility of quartz in water was then determined from the specific volume of water using the equation of Fournier and Potter (1982). The temperature corresponding to a given solubility on Figure 6 is the product of the geothermal gradient and the depth.

Most conspicuous on Figure 6 are the changes in quartz solubility in geothermal gradients from 20-60 °C/Km. These changes occur over depth ranges that are almost the same as those for maximum intraplate earthquake frequency. In the four seismic zones of the eastern U. S. earthquake frequency peaks at 7-13 km (Figure 2). For the eastern United States (about 20 °C/km) the solubility of quartz in pure water under hydrostatic pressure increases nonlinearly from about 6 to 2300 ppm over the depth range from 0-20 km. Relatively rapid changes in the rate of quartz solubility versus depth occur over the depth interval of 5-1[illegible] km for gradients above about 35°C/Km (Figure 6b, c). In gradients higher than about 55°C/Km discrete solubility maxima occur at depths of 7-9 km (Figure 6a, b).

The results shown in Figure 6 are for pure water. Furthermore, they convey nothing about the kinetics (Rimstidt and Barnes, 1980) of the solubility of quartz. The principal morphology (maximum) on Figure 6 is due to the behavior of water near the critical point. In the crystalline crust of the southeast U.S. the presence of additional chemical components such as CO_2 might shift the critical point of water to shallower depths, in the right direction for a closer correspondence between maxima in quartz solubility and the depth of frequency maxima of earthquakes. At the present time, except for the general direction of shift, the effects of additional chemical components on the critical point of water are poorly known. The effect of shifting the depth of the critical point is probably more important than any change in quartz solubility that might occur. The effects of stress corrosion on mineral solubility have not been incorporated into Figure 6. Wintsch and Dunning (1985) calculated the energy stored in dislocations in quartz as a function of dislocation density. The local high density of dislocations in pileups and tangles may result in the addition of joules per mole that (1) significantly increase the solubility of the local region of the tangle at small fluid/rock ratios, and (2) are capable of driving diffusive mass transfer of aqueous SiO_2 away from the tangle and of driving mineral reactions that are metastable with respect to strain-free quartz. Additional experimental data may be required to establish a quantitative relationship between temperature, fluid pressure, quartz solubility, stress corrosion, corrosion fatigue, and earthquake frequency. The shapes of the quartz solubility curves for high heat flow areas (Figure 6a, b) resemble the depth-frequency distribution of hypocenters as shown by Meissner and Strehlau (1982) for intraplate earthquakes. The available data suggest, therefore, that [illegible] solubility may be a factor controlling the vertical distribution of intraplate earthquakes.

Asperities

Deep (≧20 km) reservoir-induced earthquakes are evidence for preexisting fracture permeability already in place at hypocentral depths, suggesting that the fractures are already open and are or soon become fluid-filled. Open fractures imply fracture roughness, i.e., asperities, that *keep* the fractures open even in an ambient tectonic stress field. There is evidence that cracks in crystalline rocks may not close even at high pressures until ductile conditions are reached (Gale, 1975; Pratt *et al.*, 1977). Gangi's (1978) "bed of nails" model supports this assumption. Carlson and Gangi (1985) reviewed the effect of cracks on the pressure dependence of P wave velocities in crystalline rocks using the "bed of nails" model that describes the pressure dependence of velocity remarkably well up to pressures of 500 MPa. In this model, the geometry of a real crack is approximated by a mechanically equivalent distribution of cylindrical rods of different length, i. e., asperity heights. Gangi's model suggests that the mechanically equivalent distribution of cylindrical rods of different length, i. e., asperity heights, successfully used to model real cracks, might keep cracks open even at high stresses. Higher stress concentrations at and within asperities are a necessary consequence of such a model. It is well known that hydrolytic weakening of quartz is a function of stress and temperature, and would presumably take place within asperities and at asperity contacts. The presence of water in the environment of a crack tip can facilitate crack propagation by promoting weakening reactions.

Regardless of the difference between the maximum and minimum stresses, an increase in fluid pressure in fractures accompanied by mechanical and/or chemical stress corrosion will decrease the shear strength of rocks. In the eastern U.S., we suggest that this is sufficient to trigger earthquakes in a crust already stressed to near-failure.

In most (80%) of the rock volume of the groundwater basins hypothesized to encompass flowlines along which earthquakes are triggered in the Hydroseismicity hypothesis, the geothermal gradient will increase in the direction of deep flow from recharge areas; therefore, quartz placed into solution along a downward flowline from a recharge area (Figure 1) should not precipitate. As the flowlines return to the surface the solubility of quartz will decrease as the temperature decreases; this model is consistent with widespread

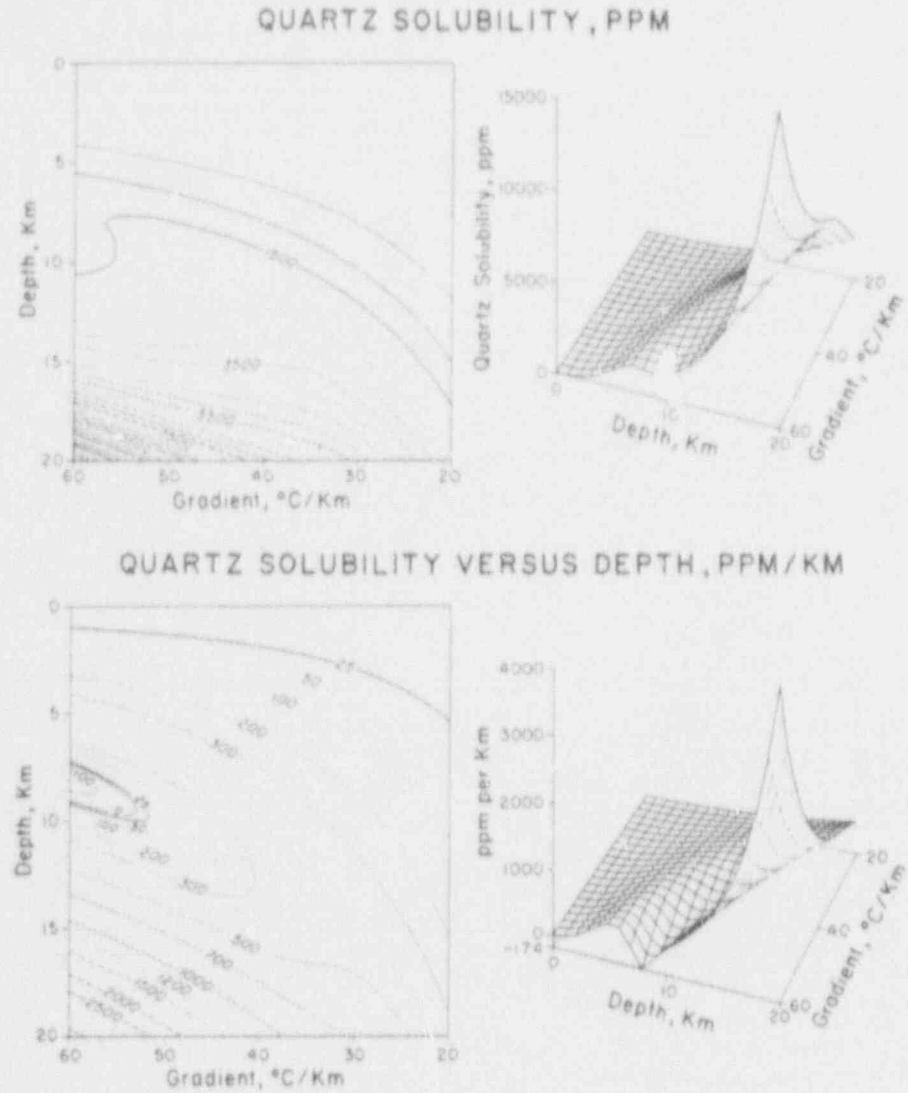

Figure 6. Solubility of quartz in pure water. a) Contour plot of quartz solubility versus depth for geothermal gradients from 20-60 °C/Km. Solubility of quartz in parts per million (ppm). b) Perspective plot of (a). Presence of CO_2 or methane in water will change the critical point of water and thus the depths in the crust at which maxima in solubility occur. c) Contour plot of the rate of quartz solubility versus depth for geothermal gradients from 20-60 °C/Km. Contour interval is ppm /km of depth. d) Perspective plot of (c).

occurrences of fine-grained SiO_2 that fill small fractures commonly observed in the crystalline Piedmont.

Transient increases in fluid potential in the fracture must induce flow from the site of pressure solution to the interior of an asperity, in the same manner as conceptually assumed for a double-porosity groundwater flow model. This flow is reversible, as it is in a double-porosity model. Rates of fluid transfer will depend on relative fracture and asperity permeabilities, and on the local hydraulic gradient at the fluid-asperity interface, but if an ambient tectonic stress is continually replenished, then ratios of fluid pressure to "intergranular" stress within asperities might reach values close to unity if asperity permeability is low enough to prevent the escape of pore fluid and thus decrease the structural integrity of the asperity. Permeability within an asperity may be reduced by a mechanism similar to that proposed by Angevine and Turcotte (1983). Transient increases in fluid pressure within asperities will decrease rock strength and promote rock failure for a fracture fabric favorably oriented with resp[illegible] ambient tectonic stress field.

Failure of brittle materials by maximum tensile stress [illegible] the boundaries of asperities in fractures filled [illegible] fluid (stress corrosion) could further weaken the [illegible] fracture. Regional planes of weakness might therefore be expected to develop at preferred orientations with respect to an ambient tectonic stress field; such directions would be strongly influenced by pre-existing fractures and structural fabric. Bollinger (1986) has, in fact, observed that focal mechanisms in the southeast U.S. can be separated out into preferred orientations.

Possible trigger mechanisms for Hydroseismicity thus include:

1. An increase in fluid pressure at hypocentral depths caused by an increase in the height of the water table in recharge area(s). The pressure increase is transmitted relatively rapidly via the network of crustal fractures.

2. Hydrolytic weakening by corrosion, stress corrosion, or corrosion fatigue leading to structural weakening of the crust. Changes in the chemical and physical conditions necessary for the corrosion are brought about by the slow, continual introduction of new groundwater. Mineral defects and rock fractures of all sizes serve as reaction species pathways and reaction sites.

The hydrolytic weakening is a slow process, reducing the shear strength of the rocks probably to near failure. The fluid pressure transients would then serve to "trigger" an earthquake in a conventional sense.

SUMMARY AND CONCLUSIONS

A new hypothesis termed Hydroseismicity is proposed to explain some intraplate seismicity. Its basis is a correlation between 1) seismogenic crustal volumes, 2) gravity-driven river basins that provide an adequate supply of water to the upper- and mid-crust, and 3) a crust with fracture permeability that is tectonically stressed close to failure. It is suggested that in crustal volumes with a combination of connected fractures and adequate groundwater, natural increases in hydraulic head in recharge areas of groundwater basins can be transmitted to depths of 10-20 km in the crystalline upper crust, and thereby trigger earthquakes, via a flow-path geometry that resembles except for scale the model familiar to groundwater hydrologists for near-surface flow. Furthermore, an increase in head in the discharge area (river) would propagate a pressure pulse back toward the recharge area. Possible trigger mechanisms for Hydroseismicity include small increases in fluid pressure at hypocentral depths caused by increases in the height of the water table in recharge area(s), and hydrolytic weakening of minerals that leads to structural weakening. Fluid pressure transients from storms are hypothesized to propagate downward from recharge areas via fracture permeability to depths of at least 20 km. We hypothesize that diffusion of pore pressure provides a small transient increase in fluid pressure and modification of ambient chemical conditions [illegible] fractured crustal volume already under a tectonic [illegible] the shear strength of rocks is decreased by these [illegible] and triggers earthquakes.

The four principal seismogenic volumes in the southeastern U. S. are within groundwater basins capable of providing an abundant supply of water to the crust, and that also intersect Eocambrian or Mesozoic rifted crust. The groundwater basins have the largest surface recharge areas as well as rivers with the highest average stream gradients as measured from their headwaters to the Fall Line.

The various types of evidence for deep fluid flow reviewed above indicate that crystalline rocks can have significant permeability to depths of 10 to 20 km during and subsequent to a metamorphic and igneous event. The permeability of the rocks is highly variable; fluid is most likely present in fractures which continuously open as older ones close. Although in some small areas fractures may be absent, it is clear that large volumes of the crust should have sufficient numbers of fractures to be quite permeable. The volume of fluid is likely to be significant and thus able to change the isotopic and perhaps elemental composition of the rocks. The water is also likely to be meteoric in a stable area such as the southeast U.S.

The most important fracture permeability in the crystalline rocks of the southeastern U. S. is along zeolite-lined fractures developed primarily during the latest rifting event. These minerals occur as euhedral fracture fillings and in open vugs, indicating open fractures. If fluid pressure in a hydraulically connected, fractured crust is more nearly hydrostatic (from a fluid of density $\rho = 1$ gm/cm^3) rather than lithostatic (rock density $\rho = 2.67$ gm/cm^3), the maximum depth at which zeolite mineral assemblages can form in the fractures is 15-20 km. Observed deep ($\cong 20$ km) reservoir-induced earthquakes are evidence for preexisting fracture permeability already in place at hypocentral depths. Open fractures imply fracture roughness, i.e., asperities, that keep the fractures open even in an ambient tectonic stress field. Higher stress concentrations at and within asperities are a necessary consequence of such a model. Hydrolytic weakening of quartz is a function of stress and temperature and should take place at and within asperities.

The four principal seismogenic volumes in the southeastern U. S. are within relatively high average-stream-gradient gravity-driven groundwater basins that can provide an abundant supply of water to the crust, and that intersect known or suspected Eocambrian or Mesozoic rifted crust. The host basins also have the largest surface recharge areas. Seismicity in the southeast U. S. is characterized spatially by steeply-dipping nodal planes and *diffuse* alignments and clusters of epicenters. These are all characteristics that are consistent with a steep to vertical fracture fabric that is currently being reactivated by pore pressure diffusion from the surface recharge of groundwater basins.

The hypothesis of Hydroseismicity incorporates a diversity of physical phenomena located well within the crust where they are difficult to study. Nevertheless, Hydroseismicity is a testable hypothesis. Measurement of the elevation of the water table is a direct measurement of the energy per unit weight of groundwater at the well point. No attempt has yet been made to correlate regional spatial and temporal changes in the elevation of the water table or other potentiometric surface with seismicity. Although these data may already be available at adequate spatial sampling, the temporal sampling interval may not be adequate. The data that are available may be difficult to interpret if well points sample locally confined as well as unconfined systems. Details of well construction may not be known, and local anisotropy in fracture permeability should be recognized. It may be several years before the 3-dimensional regional movement of a water table surface can be reliably compared with the temporal and spatial distribution of seismicity. There is evidence for reservoir-induced seismicity because both the elevations of the water level in the reservoir and the locations of the induced earthquakes are more readily available.

Six seismic events occurred during January and February, 1981, in the vicinity of Scottsville, Virginia (western edge of the central Virginia seismic zone) and within 5 km of the James River. Of these, the 3 largest [mb(Lg) = 3.4, 3.2, 2.9] occurred within a period of 3 minutes and were felt with MMI = IV. These events appear to define a tabular source volume at a depth of between 5 and 10 km that strikes northeasterly and dips steeply to the southeast (Sibol and Bollinger, 1981). We do not yet have the meteorological data to correlate with this seismic episode, but given its temporally protracted nature (19 Jan-12 Feb) it could represent response to a propagating, transient fluid pressure pulse.

It is important to note that the Hydroseismicity model has the potential for forecasting earthquake occurrence. That is, should the model prove viable, and the crustal diffusivity characteristics that follow changes in the elevation of the water table can be established, then earthquakes subsequent to basin recharge could in principle be anticipated.

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11. ABSTRACT (200 words or less)

A new tectonic model of the Appalachian orogen indicates that only one terrane boundary is present in the Piedmont and Blue Ridge of the central and southern Appalachians. This terrane boundary is the Taconic suture, which has been transported in the allochthonous Blue Ridge/Piedmont crystalline thrust nappe and is repeated at the surface by faulting and folding caused by later Paleozoic orogenies. The suture passes through the lower crust and lithosphere somewhere east of Richmond. It is spatially associated with seismicity in the central Virginia seismic zone, but is not comformable with earthquake focal planes and appears to have little causal relation to their localization.

Subsurface structure in the central Virginia seismic zone differs from that along strike in the aseismic Roanoke River traverse. The metamorphic plate is 9 km thick in the central Virginia seismic zone but only 3 km thick near the Roanoke River. The central Virginia seismic zone may be more pervasively broken by distributed high angle normal faults than the Roanoke River area. Preliminary attempts to fit a single regional p-axis to all of the planes of Munsey and Bollinger (1985) give an apparently good fit for a N55° E trending p-axis. This is subparallel to the dominant NE regional p-axis west of the Appalachians.

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