A Geophysical Study of the Earth's Crust in Central Virginia with Implications for Lower Crustal Reflections and Appalachian Crustal Structure

by

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(ABSTRACT)

Reprocessing of the United States Geological Survey's seismic reflection profile along Interstate 64 (I64) across Virginia with a data extension to 14-sec two-way travel time has provided a stacked section suitable for an integrated interpretation of refraction, earthquake, and blast analyses done by previous workers as well as gravity modelling done in this study. The seismic reflection profile shows a highly reflective upper crust which is consistent with an allochthonous Blue Ridge Province, possibly with underlying thrust sheets, and a basal decollement at about 9 km (3 sec) depth. Immediately east of the Blue Ridge province, the Appalachian structures plunge to up to 12 km (4 sec) depth where their interpretation on the section becomes ambiguous. The Evington Group, Hardware Terrane, and Chopawamsic metavolcanic rocks exposed in the Piedmont Province correspond to numerous reflections which appear to overlie a shallowly (10° to 15°) west-dipping, highly reflective zone dipping from 1.5 sec beneath the Goochland Terrane to 5 sec beneath the Evington Group rocks. Some of the overlying reflections apparently root in this zone which is therefore interpreted to include decollement surfaces along which the overlying rocks were transported. Grenville basement rocks are interpreted to underlie this zone and form autochthonous basement beneath the entire western portion of the profile because relatively few reflections originate from within this region. The Goochland granulite terrane is interpreted as a nappe structure which has overridden a portion of the Chopawamsic metavolcanic rocks. The Goochland terrane is bounded on the east on the section by a broad zone of east-dipping (20° to 45°) reflections which may penetrate to Moho depths and are possibly correlative with similar events seen on other Appalachian lines.

The I64 section contains a layered sequence of reflections at about 9 to 12 sec extending about 70 km west from Richmond, Virginia whose base coincides almost exactly with the Mohorovicic Discontinuity (Moho) interpreted from earlier refraction work. The deep reflections are thus believed to be lower crustal layering forming a 5 to 10 km thick Moho transition zone which is believed to persist across the state. The density contrast of 0.25 gm/cm^3 between the lower crust and upper mantle derived from gravity modelling, the seismic transition zone, and the presence of intrusive rocks of lower crust-upper mantle origin at the surface are consistent with partial melting and contamination of the lower crust with upper mantle material.

The refraction data and gravity modelling are consistent with a crust which thins from about 52 km beneath the Appalachian mountains to about 35 km beneath Richmond, Virginia, and rethickens by up to 10 km beneath the zone of east-dipping events east of Richmond. The pervasiveness of the zone of east-dipping events on other seismic reflection lines and the continuity of the adjacent Piedmont gravity high suggest that a similar crustal profile occurs along the length of the Appalachians.

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Introduction

In our studies of mountain belts the Appalachians are unique because of the relatively large amount of subsurface control provided by regional seismic reflection traverses. In the southern Appalachians, two COCORP profiles (Cook and others, 1979 and 1981; Nelson and others, 1985a and 1985b) and the SEISDATA lines (Behrendt, 1985 and 1986) extend from the Valley and Ridge Province across the entire Blue Ridge, Piedmont, and Coastal Plain. Other lines cross parts of the orogen in New England (Ando and others, 1983 and 1984; Hutchinson and others, 1986a and 1986b; Stewart and others, 1986), Canada (Ando and others, 1984; Stewart and others, 1986; Keen and others, 1986), and in the British Caledonides (Brewer and Smythe, 1984). These studies have made major contributions to our understanding of the Appalachians, and mountain systems in general, such as the confirmation of the importance of "thin-skinned" horizontal transport of thrust sheets (Cook and others, 1979; Harris and Bayer, 1979).

In spite of this relative abundance of geophysical data, however, major features of the internal architecture of the Appalachians remain controversial. Two such features are the pervasive zone of east-dipping events found on the seismic reflection profiles (Iverson, 1986; Ando and others, 1984) and the distinctive gravity profile (Hutchinson and others, 1983; Cook, 1984a; Karner and Watts, 1983).

The dipping events have been imaged to some degree on nearly every seismic reflection profile mentioned above at an internal portion of the orogen (Iverson, 1986; Nelson and others, 1986; Ando and others, 1984; Brewer and Smythe, 1984). They are characterized by a broad (tens of kilometers) region of reflections dipping at angles of 20° to 45° on the sections (Figure 1). Probably the best example is on the newer COCORP southern Appalachian profile (Nelson and others, 1985a and 1985b). Several authors have mentioned these events and speculated that they may represent suture zones (Nelson and others, 1985a and 1985b) or a large ramp structure controlled by a buried continental-oceanic crust transition zone remaining from the original Precambrian rifting event (Ando and others, 1983).

The gravity profile across the Appalachians (Figure 2) consists of a gravity low over the Valley and Ridge Province and an adjacent Piedmont gravity high; the transition between the two is the Appalachian gravity gradient. Several models have been proposed to explain this gravity profile (see reviews in Hutchinson and others (1983), Cook (1984b), and Karner and Watts (1983)) including a suture zone, a rift, lithospheric flexure, mantle upwarp, and upper crustal effects. On the surface, the gravity low coincides with the Appalachian topographic high and could therefore be the result of crustal thickening beneath the Appalachian mountains (Cook, 1984a for example). The gravity high is not as easily explained but has been correlated with exposures of Carolina Slate Belt and related metavolcanic rocks (Long, 1979), Mesozoic rift basins (Griscom, 1963), and with the dipping reflections imaged on the seismic reflection data (Ando and others, 1984; Hutchinson and others, 1986b). In a recent paper, it was suggested that a flexure of the lithosphere because of subsurface and surface loading is the cause of the paired high and low gravity anomalies (Karner and Watts, 1983).

The difficulty in understanding major features such as these can be attributed to several factors. Subhorizontal thrust sheets of uncertain extent and transport distance (Cook and others, 1979; Harris and Bayer, 1979) result in lower crustal rocks often unrelated to the rocks at the surface above them. Downward extrapolation of surface data is unreliable under these circumstances. Second, the seismic profiles have a relatively poor reflection quality, especially in the deep crustal portions. The identification of structures in the lower portions of the crust and understanding their

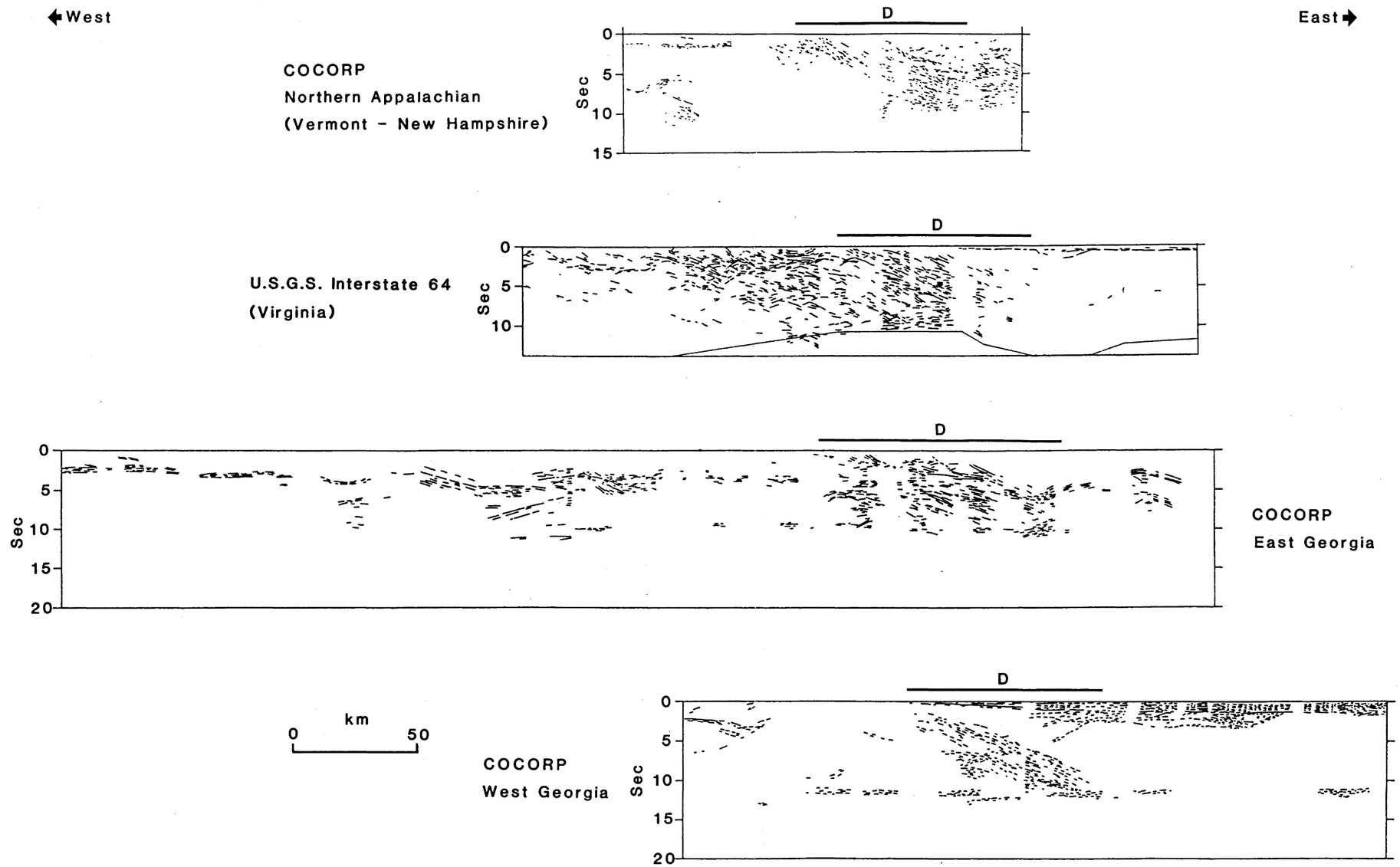

Figure 1. Line drawings of Appalachian seismic reflection profiles: Line drawings of the four seismic reflection profiles which cross the orogen and image the entire thickness of the crust. From top to bottom: COCORP northern Appalachian lines (Ando and others, 1983 and 1984), 164 line (this study), COCORP east Georgia lines (Cook and others, 1979 and 1981), COCORP west Georgia lines (Nelson and others, 1985a and 1985b). Note the presence of east-dipping events reaching deep crustal levels, and possibly penetrating the crust, on all four lines (D in figure).

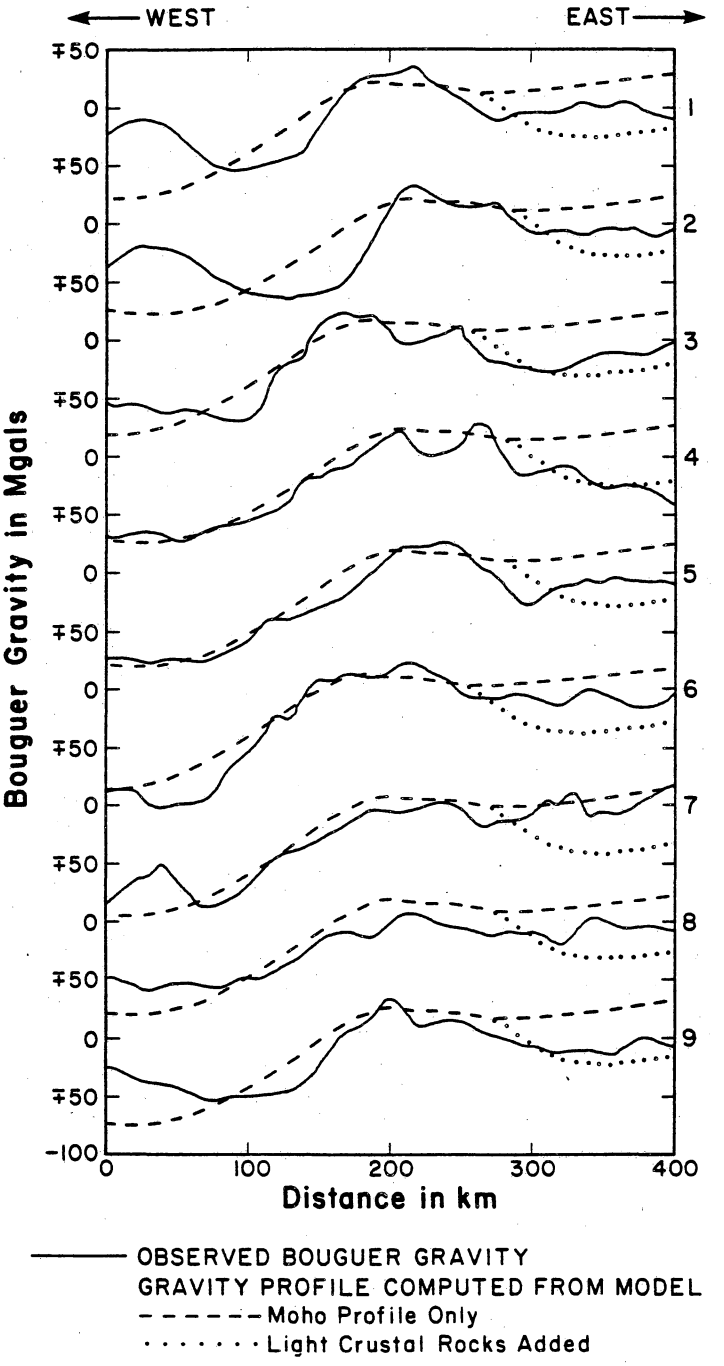

Figure 2. Gravity profiles: Nine Bouguer gravity profiles across the Piedmont gravity high at the locations shown in Figure 15. The dashed curves and dotted curves are computed from two variations of Model A in Figure 12, one with (dotted line) and one without (dashed line) lighter upper crustal rocks on the eastern side of the profiles. Compiled from Haworth and others (1980) and data from the Defense Mapping Agency.

relationship to geologic and potential field data recorded at the surface is therefore ambiguous. Finally, there is a lack of complementary geophysical data such as refraction studies to constrain the interpretation of the reflection data. Attempts to integrate all the available data have therefore been only partially successful to date (Cook, 1984b).

The Virginia region provides an exceptional opportunity to conduct an integrated geophysical study of a portion of the Appalachians because of the abundance and variety of data now available there. In addition to relatively detailed gravity maps of the region published the Virginia Division of Mineral Resources (Johnson, 1971, 1972, 1973, 1975), an extensive set of seismic experiments has been undertaken in the area. The work of James and others (1968) is the most extensive refraction study of the lower crust anywhere in the south-central Appalachians and provides a relatively detailed map of the crustal thickness in Virginia. Several other studies have also been carried out to determine and to test a well-constrained crustal velocity model for use with a regional seismograph network (Bollinger and Sibol, 1985).

Most important, these diverse data sets can be united into a coherent interpretation using the U. S. Geological Survey Vibroseis seismic reflection traverse along Interstate 64 (I64) between Staunton, Virginia and the Atlantic coast (Figure 3) (Harris and others, 1982a and 1982b). This line was recently reprocessed at Virginia Tech using an extended correlation technique to produce 6 extra seconds of data, resulting in 14-sec records. Strong reflections in the 9-12 sec range occur on the section and, in conjunction with the excellent upper crustal reflections (Harris and others, 1982a and 1982b), provide one of the best seismic reflection images of the entire continental crust available at the present time. A comprehensive and detailed geophysical data set is thus available in Virginia through which both upper and lower crustal rocks can be investigated using a multiplicity of data. Conclusions made from this relatively detailed study can, in part, be generalized to the southern and central Appalachians using other available data for extrapolation.

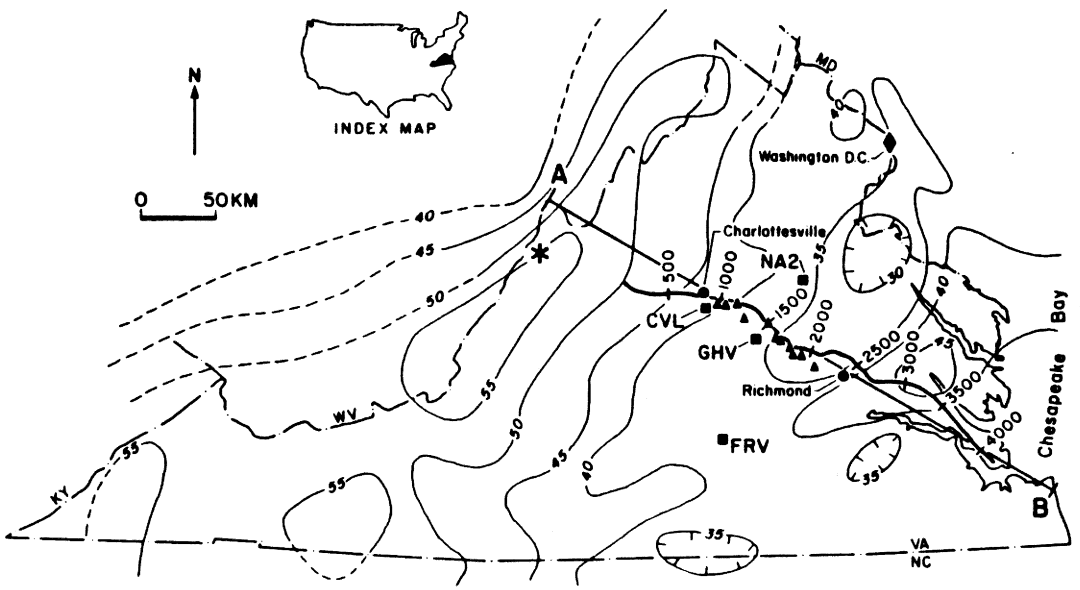

Figure 3. Index map of Virginia: Map of Virginia with crustal thickness contours in km (Figure 7 from James and others, 1968). Also shown are the locations of the I64 seismic reflection line (heavy line) with station numbers indicated, stations on Chapman's (1979) refraction line (triangles), the central Virginia seismograph stations (solid squares) used for Chapman's tripartite array (CVL, GHV, NA2) and the approximate location of the quarry blasts (asterisk) from which Chapman recorded Moho reflections at seismograph stations CVL, GHV, and FRV. Line A-B is the location of the profile discussed in the section on gravity modelling.

Previous Seismic Work

James and others (1968) provided the first extensive investigation of the deep crust in central Virginia. They used the results of previous workers in the Washington, D.C. area and seismic refraction data from the East Coast Onshore-Offshore Experiment (ECOOE) to map the crustal thickness in the mid-Atlantic states. In the latter experiment over 100 shots were fired, mostly at sea, and recorded by eight observing trucks located at various points on an approximate grid with 35-km spacing between gridpoints. Records from 200 of these source-receiver pairs were analyzed using time-term methods (see references in James and others, 1968) to produce a crustal thickness contour map of the area (Figure 3). Because most of the records used in the analysis are from offshore sources, the data are effectively unreversed profiles. According to their interpretation, the crustal thickness along the I64 profile is almost 40 km near the coast, thins to about 35 km beneath Richmond, and increases to 50 km near the western end of the line (Figure 3 and Figure 4). They also derived a subcrustal acoustic velocity of 8.15 ± 0.05 km/sec in the area.

In 1978 a network of seismic stations was installed in Central Virginia (Bollinger and Sibol, 1985). As part of the installation of this network, Chapman (1979) conducted an investigation to determine a crustal velocity model for the central Virginia region for use in locating earthquakes inside the network. Several independent methods for deriving the crustal velocity structure were used; the techniques and results relevant to this paper are summarized below.

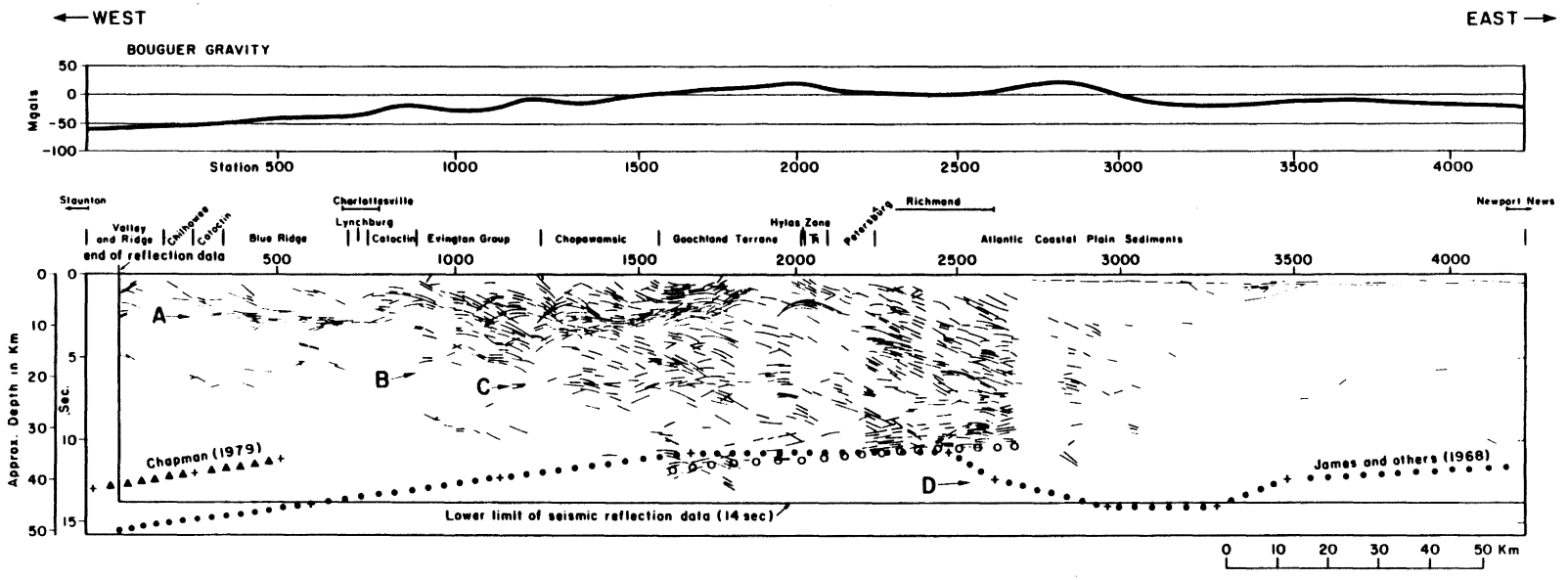

Figure 4. Line drawing of unmigrated seismic reflection data: Line drawing of the unmigrated 164 seismic reflection data with interpreted base of crust (open circles) as well as crustal profiles from James and others (1968) (solid circles) and Chapman (1979) (triangles). Crosses on the James and others (1968) profile are locations of contours on their Figure 7, crosses on Chapman's profile indicate approximate reflection points. Also included is a profile of the Bouguer gravity (Johnson, 1977) and surface geology (Calver, 1963; Farrar, 1984; Glover and others, unpublished data) along the line. Letters A, B, and C denote the major reflection *packages* extending for several hundred stations to the right, and in line with, the arrows (see text). Letter D refers to the region where there is a discrepency between the refraction data (James and others, 1968) and the seismic reflection data (see text). The section is plotted with approximately equal horizontal and vertical scales. The velocity (6.34 km/sec) used for depth-to-time conversion is from Chapman's (1979) determination.

Chapman used two experiments to determine crustal velocities. The first of these was an 80 km reversed refraction line recorded on portable seismographs (Figure 3) using quarry blasts near Richmond and Charlottesville, Virginia as sources. The line was perpendicular to geologic strike, thus sampling a wide range of rock types, and was fortuitously located near the later seismic reflection line. A velocity of 6.06 ± 0.07 km/sec was computed by analyzing records as a single-ended line and 6.09 ± 0.04 km/sec (for the crust below a thin (0-3 km) superficial low velocity layer) when analyzed as a reversed profile. No refractions from deeper layers were interpreted.

The second of Chapman's methods was the use of three of the early central Virginia network stations to measure the apparent velocities of refracted waves from quarry blasts. Velocities of 5.97 ± 0.09 km/sec for sources at distances less than 90 km and 6.5 ± 0.1 km/sec for those greater than 170 km from the network were interpreted. This is in excellent agreement with the velocity determined from the refraction line analysis and suggests a two-layer crustal model with an upper layer about 15 km in thickness. Chapman emphasizes, however, that there is a large degree of uncertainty in the crossover distance (± 20 km) due to the sparsity of stations and thus an error of ± 2 km in the estimate of the upper layer thickness. It is important to point out that no sharp interfaces were observed in the data; the nature of this velocity increase is thus unknown and may be a gradational rather than a sharp boundary.

Chapman determined a subcrustal (Pn) velocity by analysis of first arrivals from the August 29, 1975 magnitude (mb) 3.8 earthquake in northwestern South Carolina. These arrivals are consistent with a velocity of 8.18 ± 0.11 km/sec. An average crustal (Pg) velocity of 6.34 ± 0.29 km/sec was also measured from secondary arrivals from this same earthquake. The latter velocity was further substantiated by analysis of arrivals from other earthquakes from which an average crustal velocity of 6.33 ± 0.21 km/sec was determined.

One other important observation was made by Chapman. Secondary arrivals at three central Virginia stations (CVL, GHV, and FRV) from a well constrained quarry blast in Bath County, Virginia were interpreted to be reflections (PmP) from the Mohorovicic discontinuity (Moho). Chapman's interpretation of these reflections is a Moho which dips 6° to the northwest with depths of 41 km near Staunton, Virginia, 38 km just west of Charlottesville, Virginia, and 39 km midway

between them (Figure 4). These values are about 8 or 9 km less than those given by James and others (1968).[1]

Results from the aforementioned earthquake and refraction work in the central Virginia region thus suggest a 2-layer crustal model. The upper layer is about 15 km in thickness with a velocity of 6.09 ± 0.04 km/sec; the lower layer has a velocity of 6.5 ± 0.1 km/sec and a base at 36 km depth. The character of the transition between these layers is unknown from these experiments and may be gradual or sharp. A mantle velocity of 8.15 ± 0.05 km/sec and average crustal velocity of 6.34 ± 0.29 km/sec were determined. Later workers in the area have substantiated this crustal model (Carts and Bollinger, 1981; Sibol, 1983; Chapman and Bollinger, 1985) and also present evidence that the thicker crust beneath the Appalachians continues south at least into the South Carolina region (Carts and Bollinger, 1981).

Several seismic reflection profiles have also been acquired in the central Virginia region. The 8 sec I64 seismic reflection data were originally interpreted in detail by Harris and others (1982a and 1982b) and defined both the seismic-stratigraphic groups and structural style in the area. Many aspects of the interpretation of the upper crust proposed in this paper are consistent with their work, but significant differences do occur. Other 5 to 8 sec seismic reflection data were acquired in the region by personnel from Virginia Tech and will be the subject of future papers (Costain and others, in prep.; Glover and others, in prep.). These papers will provide a much more extensive discussion of the upper crust than included here; this paper emphasizes the geophysical and deep crustal aspects of the interpretation.

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[1] Figure 16 in Chapman (1979), which indicates a discrepency of only 1 or 2 km between these values, is incorrect (M. Chapman, pers. comm., 1986).

Seismic Reflection Data

The I64 seismic line was acquired in 1981 by Geophysical Service Incorporated (GSI) for the United States Geological Survey. They used 2 to 3 vibrators as an energy source and recorded on a 48 channel split-spread geophone array with a 67-m (220-ft) spacing. The largest offset in the array is only about 1675-m (5500-ft) and provides minimal velocity control because of the small normal moveout. Chapman's (1979) results therefore provide important velocity constraints for depth computations from the seismic reflection data. A 10-sec, 14-to-56 Hz upsweep was used with an 18-sec listening time to produce 8-sec correlated records. The resulting 12-fold stacked section is illustrated and described in Harris and others (1982a and 1982b).

The data were reprocessed at Virginia Tech on a VAX 11/780 computer using the Digicon, Inc. DISCO software package. With the exception of the extended correlation (Appendix A; Okaya, 1986) to produce 14 sec records and the application of Vibroseis Whitening (Coruh and Costain, 1983), a routine processing sequence was followed (Appendix B). The application of Vibroseis Whitening (AGC before correlation) generally results in an improved signal-to-noise ratio but loses the true amplitude information; relative amplitudes, however, remain nearly the same (Coruh and Costain, 1983). Several of the shot records have been recorrelated without spectral whitening and are indicated as such in the figure captions where displayed.

The entire section is too large for meaningful reproduction, but line drawings of the unmigrated and migrated stacks are shown in Figure 4 and Figure 5, respectively, and panels of the actual data are shown where appropriate.

Because of the lack of velocity control, a linear velocity function of 5200 m/sec at the surface and 6500 m/sec at 14 sec was used in a post-stack time migration. This velocity function is about 5% lower than the crustal velocities indicated by the refraction data. It is, however, consistent with experience and results from other workers who find that slightly low velocities produce the most effective migrations, particularly on deep reflections where the near-surface rocks can cause severe distortion of the wave field (Warner and others, 1986). The migration appeared to do an adequate job of collapsing the arcuate diffraction events and uncrossing many of the intersecting events seen on the unmigrated data. The eastern portion of the line, as will be discussed later, has a poor signal-to-noise ratio and migration was therefore not attempted beyond station 3200. The unmigrated stacked section was used for the line drawing of the eastern portion in Figure 5.

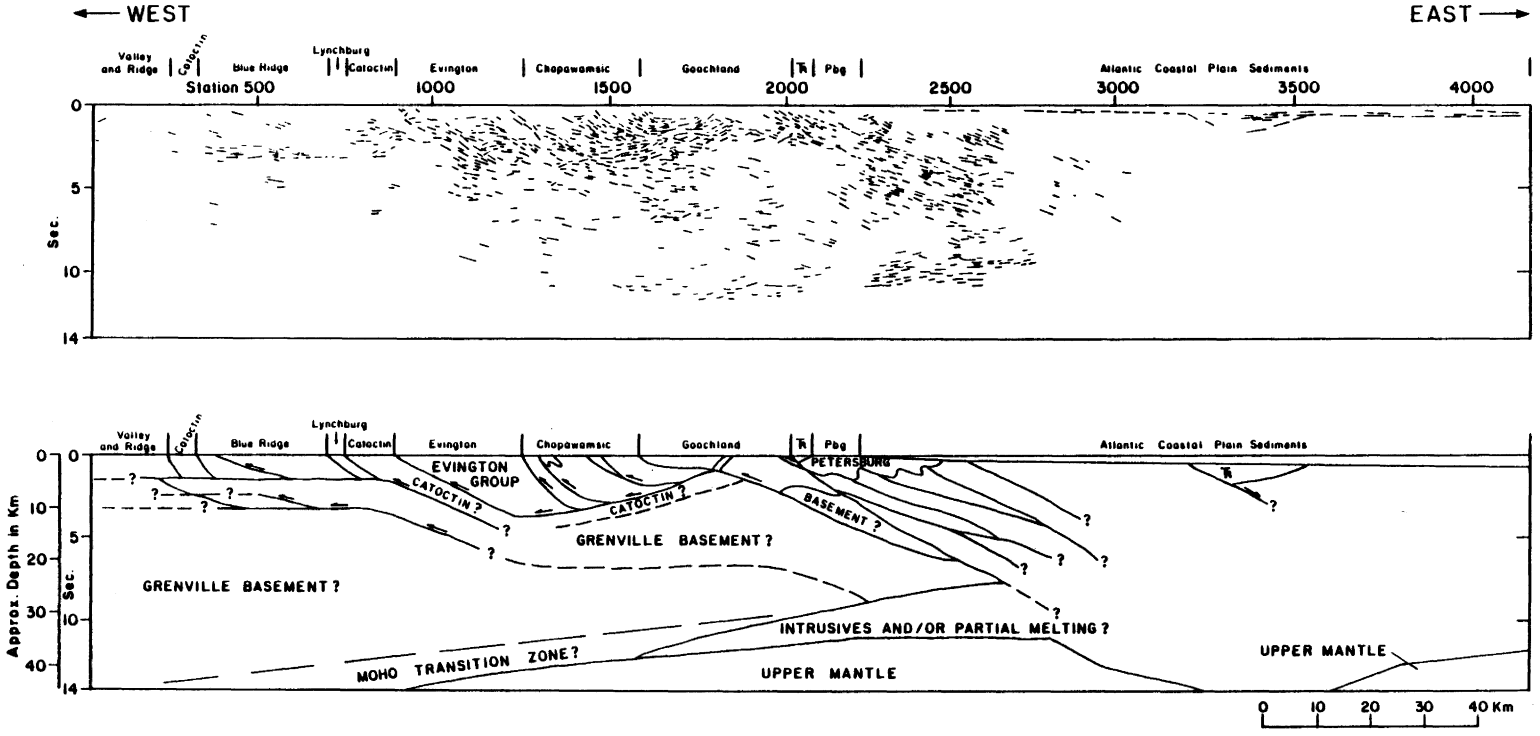

Figure 5. Line drawing of migrated seismic reflection data: Line drawing of the stacked section after post-stack time migration with a possible interpretation drawn below. An alternative interpretation of the upper portion of the data can be found in Harris and others (1982b). The eastern portion of the line drawing (beyond station 3200) is taken from the unmigrated stacked section (see text).

Interpretation of Seismic Reflection Data

Data Aquisition Artifacts (?)

The most noticeable aspect of the reprocessed section (Figure 4) is the poor reflectivity recorded east of station 2700 where Harris and others (1982b) terminated their published profile. A strong Coastal Plain - basement reflection and a weakly imaged basinal structure, interpreted to be a Mesozoic rift graben because nearby drill holes bottomed in Triassic sediments (Johnson, 1975), are visible; otherwise the data are virtually featureless. The boundary between good and poor data is abrupt and nearly vertical on the stacked section. A thin vertical strip of good data also appears slightly farther east, beneath station 2800 (Figure 4).

Such vertical boundaries seem unlikely on first inspection but cannot be ruled out immediately. Major faults with a strike-slip component of motion are found in the Piedmont Province (Gates and others, 1986a) and could conceivably create near-vertical boundaries through the crust with differing terranes on each side. Inconsistent with this interpretation, however, is a lack of significant diffractions on the unmigrated section from the reflectors truncated at these hypothetical vertical boundaries. A feature of this nature and magnitude is also not readily apparent on other profiles along strike (Hutchinson, 1986b; Cook and others, 1979 and 1981; Behrendt, 1986).

A more likely explanation for the non-reflective portions of the data is that they are due to a data acquisition problem, although no changes of any kind in the field acquisition parameters have yet been recognized. Common-source records spanning the good-poor transition do not show an abrupt change in reflection amplitude between adjacent records, which suggests that it is not a source or recording parameter (vibrator force, amplifier gain, etc.) which was changed between records. An electrical or mechanical acquisition problem, such as vibrator desynchronization, is a possibility but its persistence over several days of shooting, as here, is difficult to imagine.

One possible cause is that a transmission problem is associated with the ground along the last third of the line, except for the narrow zone near station 2800. A negligible amount of energy is transmitted or received except where unusually bright interfaces are present such as those associated with the Mesozoic basin and the Coastal Plain sediments. Because the energy does not penetrate below these sediments, the barrier is constrained to lie within them (including the basal reflector) or on the surface. Energy transmission, judging from a qualitative evaluation of reflector strength, is otherwise good across the profile. Fluctuations in the reflection amplitudes do occur across the line, thereby creating narrow vertical strips of excellent quality, probably due to near surface effects (ground coupling, statics, etc.).

The Upper Crust

A generalized geologic map of the east-central portion of Virginia is shown in Figure 6 with the seismic reflection profile marked with a heavy line. The western portion of the profile begins in the Virginialley and Ridge sediments (Colton, 1970) and crosses the adjacent Blue Ridge anticlinorium (Espenshade, 1970; Espenshade and Clarke, 1976; Wehr and Glover, 1985). A zone of subhorizontal reflections at about 3 sec on the western end of the section (A on Figure 4 and Figure 5) is interpreted to originate from autochthonous(?) lower Paleozoic sedimentary rocks and perhaps a decollement at the base of the overlying allochthonous thrust sheet(s). This is the same structure consistently imaged beneath the Blue Ridge on other southern Appalachian seismic re-

flection data (Cook and others, 1979; Harris and Bayer, 1979; Coruh and others, in prep.) and may be continuous to the north where a similar reflector geometry is seen on the COCORP northern Appalachian line (Ando and others, 1983 and 1984). Relatively unaltered platform strata beneath the Blue Ridge in North and South Carolina were suggested because of carbonate blocks brought to the surface in the Brevard fault zone (review in Hatcher, 1978). The Appalachian Deep Core Hole (ADCOH) seismic reflection data (Coruh and others, in prep.). clearly imaged subhorizontal reflectors interpreted to be these platform sediments. Though individual, continuous shelf strata and thrust surfaces are not discernible on the I64 data, a structural complexity similar to that imaged on the ADCOH data, which showed a stacking of several thrust sheets (Coruh and others, in prep.), is suggested by the few reflections which are seen.

Steeply east-dipping reflections beneath station 1000, which reach depths of 12 km (4 sec), define an eastern limit to these subhorizontal thrust sheets and indicate that the fault sufaces must turn down east of the Blue Ridge anticlinorium. These data do not, unfortunately, adequately constrain the interpretation to tenably answer the important question of where the exposed basement rocks, which may represent the ancient continental hinge zone (Wehr and Glover, 1985), originally lay. After turning down, the thrust surfaces could be interpreted to merge with either of the large reflection packages marked B or C on Figure 4 and Figure 5. Reconstructing the late Precambrian position and subsequent transport of the Blue Ridge anticlinorium is therefore unresolved by this data at present. It is tempting to speculate, however, that the ramp-like structure imaged here is correlative with the mid-crustal dipping events imaged on the COCORP line beneath the Kings Mountain Belt (Cook and others, 1979); both sets of dipping events reach mid-crustal levels and coincide with the Appalachian gravity gradient.

The Grenville basement interpreted to underlie the thrust sheet(s) has a relatively low reflectivity with only a few isolated reflections recorded, a characteristic of Precambrian cratons in other areas as well (Gibbs, 1986). A change in source strength or energy penetration is unlikely as a cause for this acoustic transparency because the amplitudes of the overlying reflections are consistent across the line (at least as far west as station 300). On this basis, pristine Grenville crust is interpreted to underlie the entire western portion of the traverse (Figure 5).

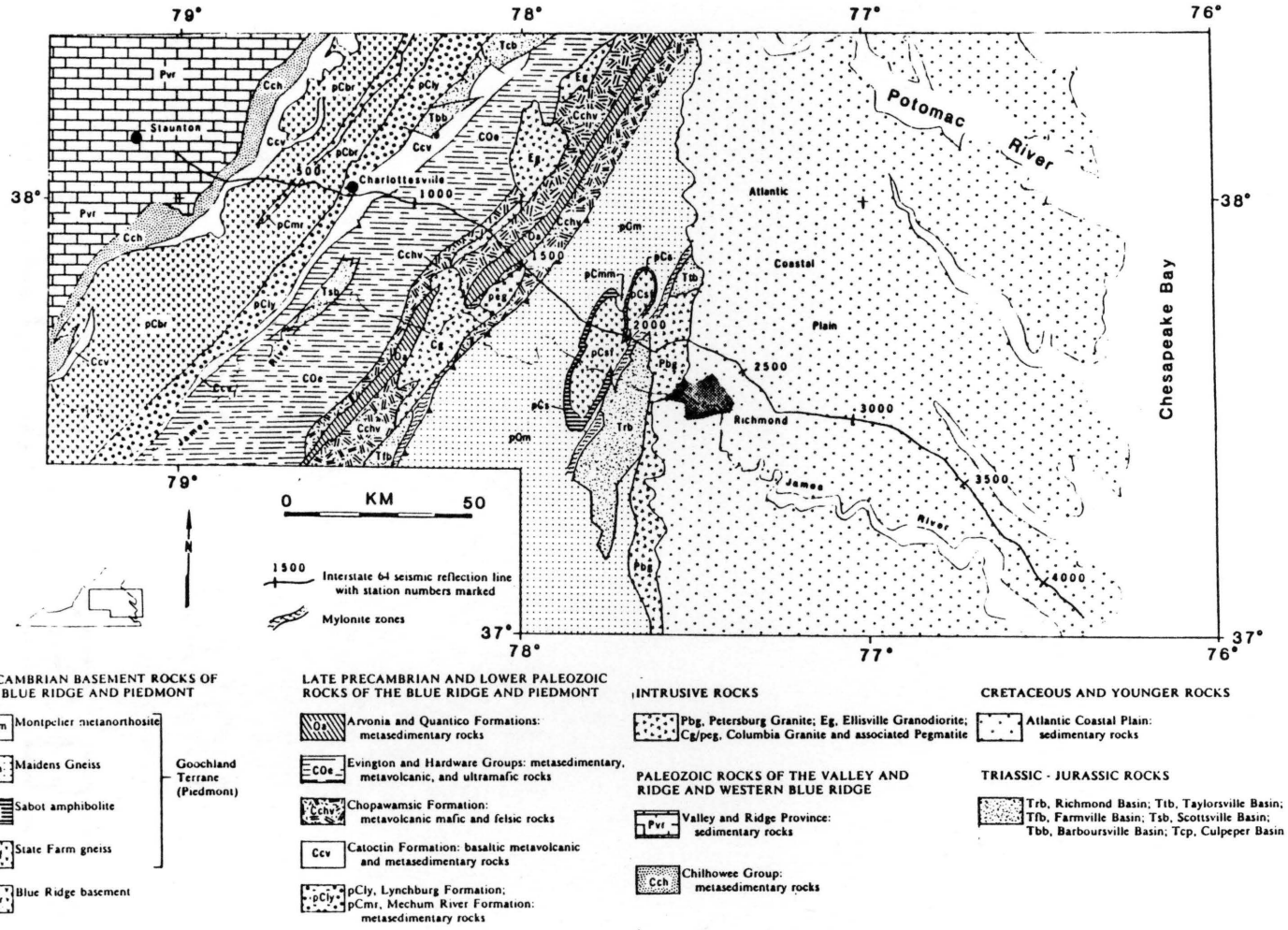

Figure 6. Geologic map of part of Virginia: A geologic map of the east-central part of Virginia compiled from Calver (1963), Bourland (1976), Poland (1976), Reilly (1980), and Farrar (1984). The I64 seismic reflection line is shown as a heavy black line with station numbers marked for reference.

On the surface east of the Blue Ridge are exposures of Lynchburg metasedimentary, Catoctin metavolcanic, Evington Group (including Candler and Hardware) metasedimentary and metavolcanic, Chopawamsic metavolcanic rocks (Bland and Blackburn, 1979; Brown, 1970; Conley, 1978; Evans, 1984; Wehr, 1983; Wehr and Glover, 1985). The Lynchburg-Catoctin sequence is interpreted as Late Precambrian to Cambrian sediments and basalts emplaced on Blue Ridge (Grenville) basement during the opening of the Iapetus ocean (Espenshade, 1970; Fitcher and Diecchio, 1986; Rankin, 1976; Wehr and Glover, 1985). The Evington Group (Candler) metasedimentary and metavolcanic rocks are not as well understood, but are believed to be deeper-water equivalents to the lower Valley and Ridge carbonate shelf strata (Brown, 1970; Evans, 1984). Their eastern edge has been divided into the Hardware Terrane, which may include and ancient subduction zone complex (Shores Melange) possibly formed in a small ocean basin (Bland and Brown, 1977; Evans, 1984). The Chopawamsic metavolcanic rocks have an island-arc composition and are interpreted to have formed along the continental margin (separated by a small ocean basin) or as a separate arc; they are believed to have been accreted to North America during the Taconic collisional event in the mid-to-late Ordovician (Bland and Blackburn, 1979; Pavlides, 1981). They were intruded by the Columbia granite, probably at about 454±9 Ma (Mose and Nagel, 1982), and are partly overlain by deep water shales of the Arvonia Formation in the Arvonia and Columbia synclines (Conley, 1978; Brown, 1970) (for simplicity the Columbia granite and Arvonia Formation are not labelled on the Seismic sections). All of these units have undergone metamorphism and deformation in either the Taconic, Acadian or Alleghanian events (Glover and others, 1983; previous references).

Reflections which project to the surface where the Lynchburg and Catoctin rocks are exposed can be interpreted to be truncated at about 1 sec on the section except those on the eastern side of the Catoctin which dip eastward and merge with either reflection package B or C on Figure 4. A layer of Catoctin rocks within reflection package B would explain the high reflectivity of the layer (Brennan, 1985) and is consistent with the Catoctin rocks underlying the Evington group metasediments. The Lynchburg Formation was apparently deposited within Precambrian rift basins and may vary considerably in thickness or pinch out completely in a short lateral distance

(Wehr and Glover, 1985; Wehr, 1983). Interpreting this formation on the record section is therefore difficult unless directly correlated with surface exposures.

The Evington and Chopawamsic rocks correspond to a highly reflective area on the seismic profile and appear to lie within a basin-shaped synform beneath their surface exposures (Figure 4 and Figure 5). Dipping events within the synform, at least some of which can be interpreted to be faults on the basis of convergence and truncation of reflections, appear to root in a complex and extremely reflective west-dipping zone (B in Figure 4, Figure 5, and Figure 7). Reflections from below this zone are markedly less abundant, implying a different, more acoustically transparent rock type. The reflective zone (B) is thus being interpreted as the base of the Chopawamsic - Evington sequence, possibly with a reflective sequence of Catoctin metabasaltic rocks at their base. This zone may also be a basal decollement, later tilted to the west, along which the overlying rocks were transported.

Below this highly reflective zone (B) on the section is a comparatively non-reflective region, interpreted as Grenville basement rocks, whose upper surface defines a broad arch-like geometry on the reflection profile. The top of this arch occurs near station 1900 where the State Farm Gneiss is exposed at the surface. The State Farm Gneiss is composed of granulite-grade gneisses, schists, and plutonic rocks dated at 1031 ± 94 m.y. (Glover and others, 1978 and 1982) and probably of North American origin (Farrar, 1984). The State Farm Gneiss is overlain by the Sabot amphibolite and Maiden's gneiss; together the three units, plus the intrusive Montpelier metanorthosite, make up the Goochland Terrane (Glover and others, 1982; Farrar, 1984). The Goochland Terrane may be interpreted from the seismic data as a nappe which broke through and overrode the Chopawamsic metavolcanic rocks (Figure 5). The continuity of a Paleozoic amphibolite grade metamorphic event between the Goochland and Chopawamsic Terranes suggests that the two were juxtaposed before the Alleghanian event (Farrar, 1984). The relatively low reflectivity within the arch and the correspondence of its apex with a nappe of basement material (State Farm Gneiss) suggest that it is composed predominantly of high-grade basement material. The east-dipping events between 3 and 7 sec within the Grenville basement rocks interpreted to core the arch may be deformation zones (mylonites) indicative either of nappe structures, one of which now reaches

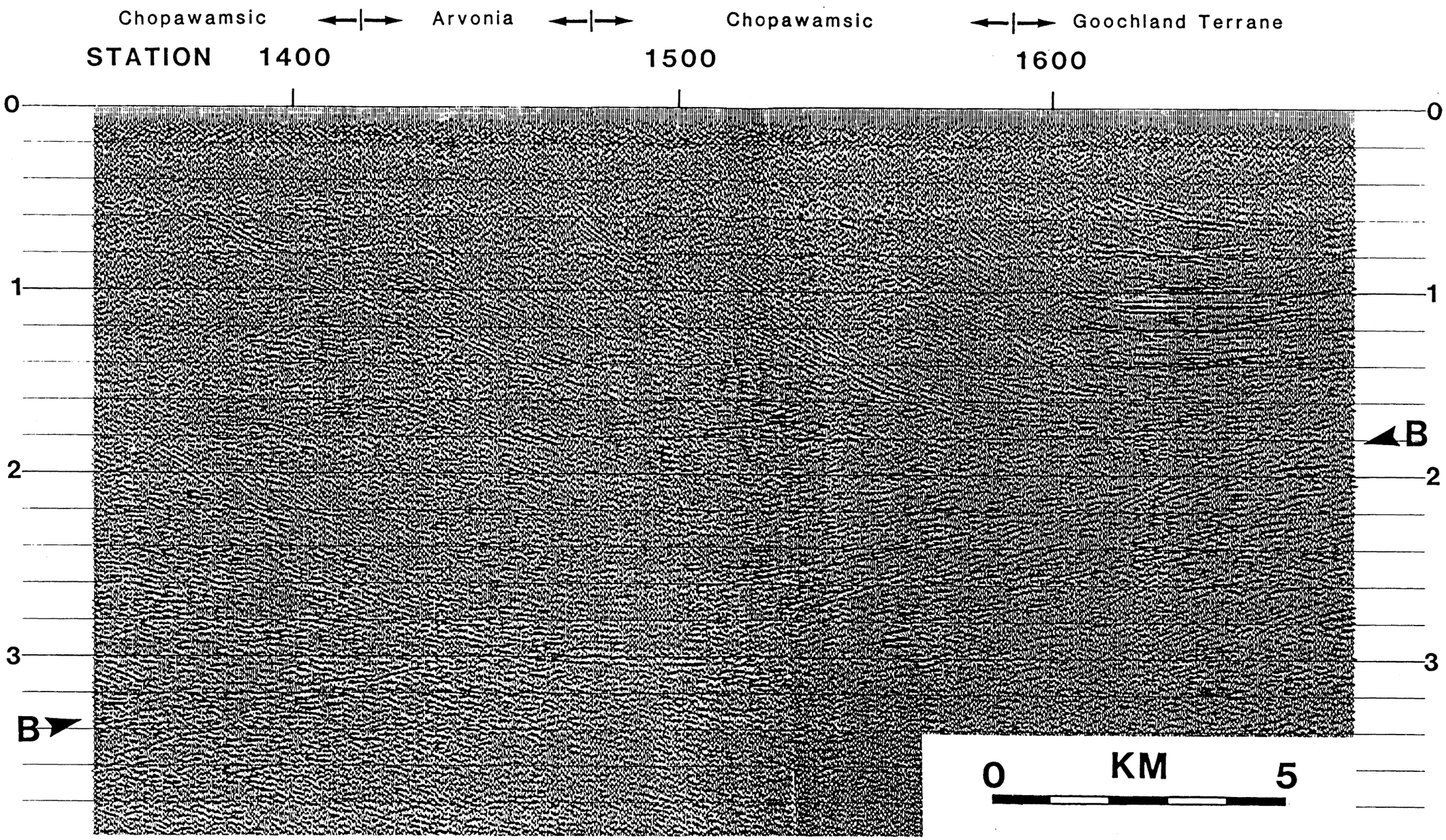

Figure 7. A portion of line I64: Upper portion of migrated seismic reflection data acquired over the Chopawamsic metavolcanic rocks and Goochland Terrane. The letters 'B' mark the complex zone of dipping events with the same label in Figure 4 and Figure 5. Note that some of the shallow dipping events appear to merge with zone 'B', such as the event below station 1600 at 1.7 sec.

the surface as the Goochland Terrane, or Alleghanian strike-slip movement (Gates and Glover, 1986; Gates and others, 1986a). The basement uplift is probably Alleghanian or younger in age because at that time the Goochland Terrane and its possible southern equivalent, the Raleigh Belt, were cooling after amphibolite-grade metamorphism at 5-7 kb pressure (Durrant and others, 1980; Farrar, 1984 and 1985); indicative of up to 17-25 km of burial (Farrar, 1984 and 1985). Possible modes of uplift include duplication of basement rocks by thrust stacking or an arching of the entire crust.

The band of sub-horizontal reflections at mid-crustal levels which turns down towards the east (C in Figure 4 and Figure 5) may originate from either deformation zones entirely within the basement, or from rocks deposited on thinned (rifted) crust (Lynchburg, Catoctin, or Lower Paleozoic sediments?) overridden during later thrusting. An intriguing possibility is that the basement rocks bounded by reflections B and C in Figure 4 and Figure 5 (including the State Farm Gneiss) is a piece of rifted (transitional) North American crust which was thrust onto similar crust, possibly during the Taconic collisional event.

The eastern edge of the Goochland Terrane is bounded by the Hylas ductile deformation zone (Bobyarchick and Glover, 1979). The Hylas is a major boundary in the area which underwent at least one episode of ductile deformation and was later reactivated as a brittle normal fault during Triassic rifting (Bobyarchick and Glover, 1979; Gates and Glover, 1986). It now forms the western boundary of the Richmond Mesozoic rift basin and juxtaposed the Petersburg granite against the Goochland Terrane (Bobyarchick and Glover, 1979). The Hylas is also part of the Eastern Piedmont Fault system (Hatcher and others, 1977), which includes the Modoc, Augusta, and other faults interpreted to have a similar deformation history.

Reflections reaching the surface near the Hylas zone form the western boundary of a region of prominent events which dip to the east at angles of 20° to 45°. Individual reflections are generally short segments, but the entire reflection package appears to penetrate to Moho depths without flattening, although there is a data loss at the eastern portion of the zone. These dipping events obviously represent a significant crustal boundary and may correlate with similar events imaged

beneath the Atlantic Coastal Plain on other Appalachian lines (Iverson 1986; Nelson and others, 1986).

The interpretation favored here for these dipping events is that they delineate the late Precambrian continental margin and the later Taconic collision zone along which the island arc terranes (Chopawamsic metavolcanic rocks, Carolina Slate Belt) were accreted to North America (Bland and Blackburn, 1979). This crustal boundary is east of it equivalent surface structure, the eastern edge of the Chopawamsic metavolcanic rocks (Evans, 1984), but westward transport of accreted material via nappe emplacement would be expected in a collision of this nature and appropriate thrust structures are evident on the seismic section. This interpretation would require all the material east of the zone to have been accreted or formed since the beginning of the Taconic event.

Onlapping the Petersburg granite near Richmond, the Atlantic Coastal Plain sediments (Brown and others, 1972) cover most of the older basement surface, making it accessible only by drilling and other geophysical techniques. Where drilled or exposed, however, it is seen to be composed of synmetamorphic Alleghanian granitoids (Petersburg and Portsmouth granites in this area), Slate Belt-like metavolcanic rocks (eastern Slate Belt), mafic rocks, and Mesozoic sedimentary rocks (Bonini and Woolard, 1960; Gleason, 1979; Johnson, 1973 and 1975; Russell and others, 1985; Williams, 1978).

The seismic data acquired over the Petersburg granite contains dipping events which reach shallow levels, indicating that the granite either does not penetrate more than several kilometers in depth or it is unusually reflective, possibly due to mylonite zones like those seen at the surface within the granite. The gravity modelling discussed in a later section is consistent with a body about 5 km or less in thickness, indicating that at least some of the shallow reflections originate from within the granite body. East of the granite, beneath the Atlantic Coastal Plain sediments, the only feature visible on the section is a basin-shaped reflection package about 22 km in width and with a 1.2 sec two-way travel time. It coincides with drill holes which penetrated Triassic sediments (Johnson, 1975) and is therefore interpreted as a Mesozoic rift basin about 2.7 km in thickness.

The Lower Crust

Origin of Late Arrivals

Before interpreting the later arriving reflections it is important to insert a few words of caution and mention that several possible origins for the 8 to 12-sec events are possible other than primary reflections (Jones, 1985). These include: coherent noise, multiples or reverberations, reflections from out of the plane of the section (sideswipe), and reflected refractions. The possibility that the "deep" reflections are actually one of the above forms of noise should be considered before interpreting the data.

Coherent noise due to the Vibroseis method does not seem to be a likely explanation for several reasons. The amplitudes are as large or larger than the events above them (Figure 8), and do not seem to have any geometric relation to the shallower events (Figure 4). Harmonic distortion (Seriff and Kim, 1970), one common form of coherent noise, could produce "ghost" images of a primary reflection at earlier or later arrival times, but these would be lower in amplitude and similar in geometry to the primary. The events seen in this data satisfy neither of these criteria.

Similar arguments can be made against the arrivals being multiples of the shallower reflections. They are not related to the events above them on the section in time or dip angle (Figure 4). The amplitudes of the deeper events also appear to be much larger than the overlying events (Figure 8), which would be unlikely if they are multiples. Because of the small receiver offsets (1675m maximum) and relatively high velocities of the crystalline rocks, velocity analyses can not be used to recognize multiple events. The amplitude and apparent lack of any relationship with the overlying reflections, however, is strong evidence against the reflections being multiples, even without this last test.

Reflections from out of the plane of the section are possible because the source and geophone arrays were in-line. There was thus no discrimination in the field against energy coming from the side. The arguments against the reflections being due to sideswipe are: 1) a lack of correspondence between the reflections and changes in the line direction and 2) the consistent strike of the surface

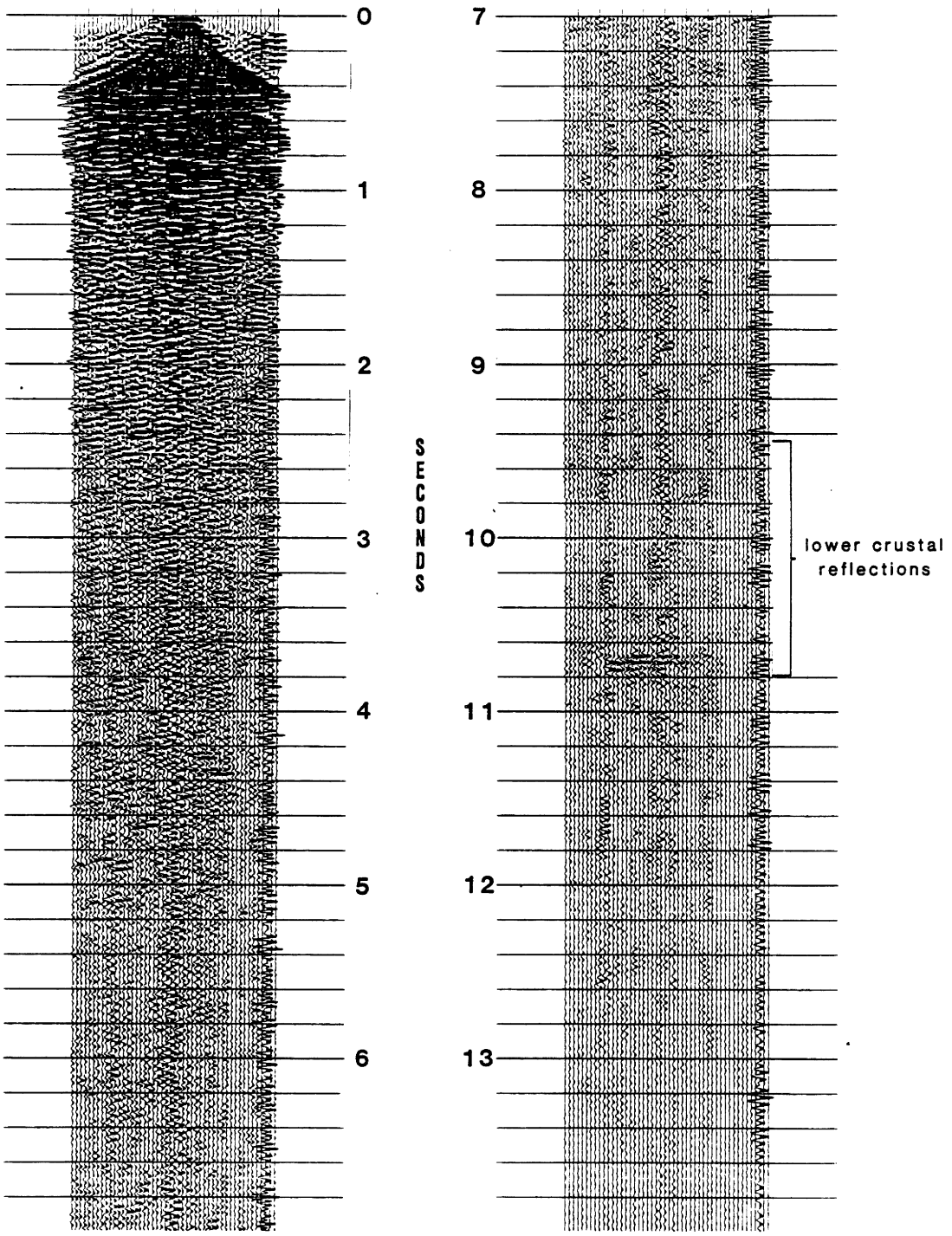

Figure 8. **Shot records showing Moho reflections:** Records with no Vibroseis Whitening (see text) showing the late arriving (9-12 sec two-way traveltime) reflections at true amplitude. They are interpreted as lower crustal layering.

geologic contacts in the area. The interfaces from which the energy was reflected are located at considerable distances (about 25 to 40 km) from the source and, because they are subhorizontal on the section, they must be oriented nearly parallel to the line. Layered reflectors to the side of the line with the necessary orientation would therefore intersect the Appalachians at an oblique angle and require the presence of a major anomalous structure. Invoking Occum's razor, a more likely location for the reflectors is sub-horizontal and in the plane of the section, especially considering the positive correlation with the previously mentioned blast and refraction data (Figure 4).

Reflected refractions, waves travelling along an interface, as a refraction and reflecting from an interface are another possible explanation for the deep arrivals. The relatively large amplitude of the deep events supports this explanation because refracted waves spread in two, rather than three, dimensions and therefore retain their energy over longer distances. The same reasoning as with the sideswipe argument is valid here though: there is no apparent correlation with changes in line direction and reflected refractions require a much more complex and extremely fortuitous subsurface geometry than a simple reflection from deep, sub-horizontal interfaces. The late arrivals (8 to 12 sec) are thus interpreted as primary reflected events from the lower crust.

The Mohorovicic Discontinuity and Lower Crustal Layering

Shown on the line drawing in Figure 4 are the locations of the base of the crust as interpreted by James and others (1968, their Figure 7) and Chapman (1979). Depths for both of these were converted to time using a single-layer average crustal velocity of 6.34 km/sec determined by Chapman for the central Virginia area and then plotted on the time section. This velocity function differs from that of James and others (1968) who utilized a spatially varying velocity function ranging from 6.30 to 6.55 km/sec and is slightly less than the 6.37 km/sec average velocity of Chapman's (1979) crustal model. The difference between these three crustal models, however, is less than 4% and produces a maximum effect of only about 0.5 sec on the time section (about 1.6 km in depth). There is also some lateral error, estimated to be about $\pm$ 100 stations ($\pm$ 7 km), in the location of Chapman's reflection points because the precise location of the sources (quarry blasts) is not known and because the points are projected onto the cross section along strike.

On the I64 reflection data, James and others' (1968) Moho is coincident with the base of a layered set of reflections at 10 to 12 sec. No strong reflections appear on the stacked section below this layered sequence except dipping events which migrate to a position above it, consistent with an acoustically transparent upper mantle. The layered subhorizontal reflections are therefore constrained to lie within the lower crust. Unfortunately, it is not possible with the present data set to determine the velocity structure within this layered sequence except that there is an upper mantle velocity of 8.15 km/sec at its base.

The easternmost portion of the (good quality) multichannel reflection data suggest a considerably thinner crust than the profile of James and others (1968) (position D in Figure 4). This discrepancy could be due to:

1) A lack of reflections from the deepest part of the crust. The deeper portion is therefore seen in the refraction data but not imaged on the reflection data. This seems an unlikely explanation given the consistency of the Moho reflections and their close match with the refraction data over a span of almost 70 km immediately to the west.
2) The crustal velocity structure created a misinterpretation of the crustal thickness. One possible cause would be a large velocity anomaly which delays the horizontally travelling (refracted) waves but not the near-vertical reflected energy; the retarded arrivals were then interpreted as a thicker crust. The anomaly, if there is one, may be in the upper mantle (where the refracted waves travelled) and is east of the seismic network and recorded earthquake and quarry blasts, so no independent velocity control is available to check this hypothesis.
3) The structure in the Moho caused a "shifted" interpretation of the refraction data and the thicker crust actually lies farther to the east, where the reflections are too weak to image it. Structural distortion is a common problem in time term analysis over a complex surface, particularly when most of the refraction records are unreversed, as in the ECOOE data. James and others (1968) discuss this distortion and try to avoid it by applying corrections. It is possible, however, that their corrections, which were only approximate, were not sufficient in this area. Shifting the Moho interface interpreted from the ECOOE refraction data

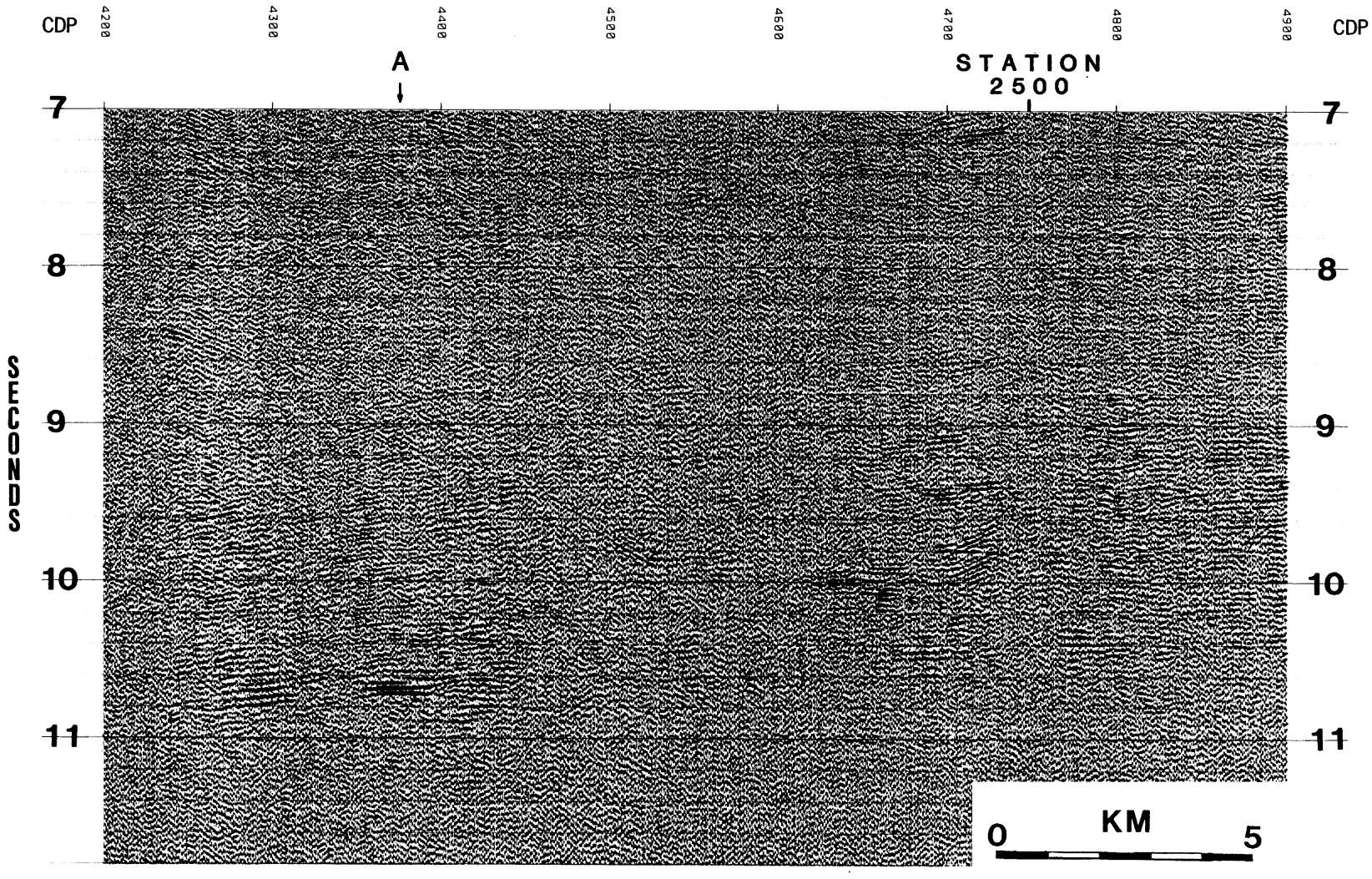

Figure 9. Moho reflections: Portion of the stacked section showing the layered reflections interpreted to be from the lower crust, the base of which is interpreted as the Mohorovicic discontinuity. The letter 'A' marks the location of the shot record shown in Figure 8.

about 250 surface stations (17.5 km) to the east produces a much better fit with the I64 reflection data and is used in the final interpretation (Figure 5).

On the western segment of the profile the multichannel reflection data do not image the Moho where Chapman interpreted reflections from it, possibly due to two major differences between the experiments. Chapman apparently measured events which, at epicentral distances of 120 to 170 km and a crustal thickness of 40 km, impinged upon the Moho with incidence angles (to the vertical) of about 60°. This is near the 52° critical angle for the Moho interface (computed from Chapman's velocity model) and, as illustrated in Figure 10 by synthetic modelling of possible Moho velocity structures, would produce a reflection of nearly the maximum possible amplitude; the near-vertical multichannel data would have a much lower reflection coefficient. Chapman also used seismometers with a natural frequency of 1 Hz (M. Chapman, pers. comm., 1986) which can record well frequencies lower than the 14-56 Hz I64 reflection data. Figure 11 shows results from synthetic modelling which demonstrate that data of this sort can be much more sensitive to a gradational boundary, particularly in the case of a ramp-like velocity function, than the higher frequency data.

One explanation consistent with the data and with results from similar experiments (Bamford and Prodehl, 1977; Braile and Chaing, 1986; Gibbs, 1986) is that the Moho is characterized by a smoother transition (a velocity ramp) beneath the Grenville basement on the western portion of the section than under the crust to the east (a layered velocity function). Synthetic modelling (Figure 11) demonstrates that a ramped velocity function, representing a gradational velocity increase with traveltime, can produce a sizeable reflection from the top and bottom of the ramp on relatively low frequency data yet remain nearly invisible on higher frequency data. Chapman may have recorded reflections from the top of such a gradational boundary (he picked the first arrivals) and the refracted energy may have travelled along the base (the most rapidly travelling wave was picked). Though there is uncertaintly in the accuracy of both data sets, a total thickness of approximately 8 to 9 km could be interpreted for the crust-mantle transition zone. This is nearly the same thickness as the layered reflections farther east on the I64 data and on data from other areas (Bamford and Prodehl, 1977; Braile and Chaing, 1986).

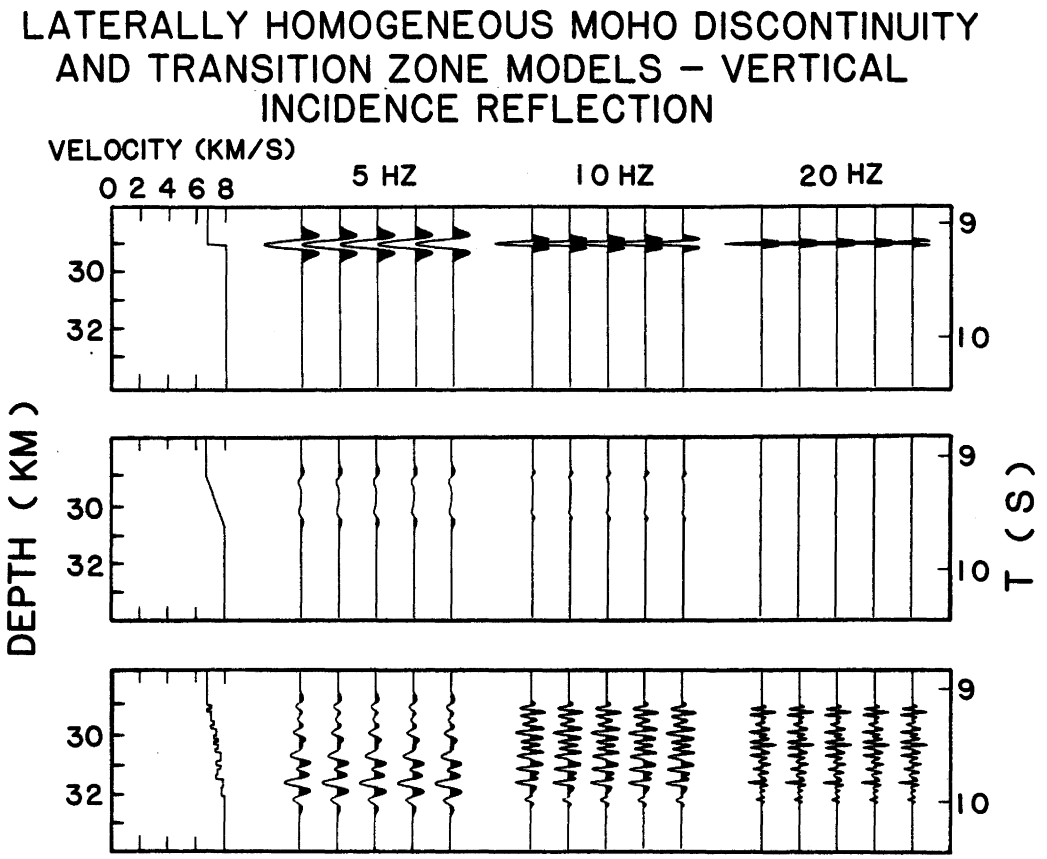

Figure 10. Moho transition zone reflection characteristics: Figure from Braile and Chaing (1986) demonstrating by synthetic modelling the effects of source-receiver distance (or incidence angle) and frequency on reflections from three possible Moho transition zones at depths of 29-32 km. The line and dot patterns show the approximate frequency and distance ranges of vertical incidence (multichannel) and wide-angle reflection data. Model I corresponds to a sharp interface; the reflections show a large amplitude variation with distance but no frequency dependence. Model II is a smooth velocity gradient (ramp function); distance (incidence angle) has a strong effect on reflection amplitudes as in Model I but there is also a strong frequency dependence, particularly on near-vertical data (light line). Model III are layered transition zones; there is also a strong effect on reflection amplitudes due to distance (incidence angle), but both graphs show the results of interference effects which will depend on the interface in question.

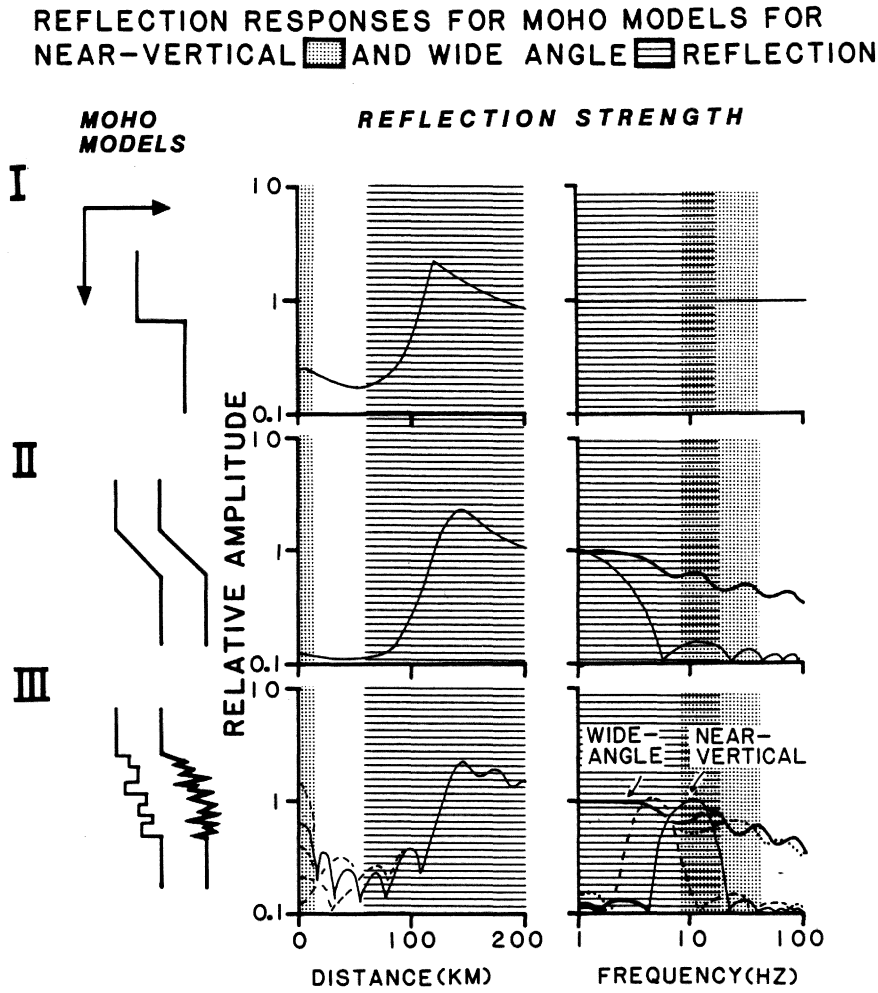

Figure 11. Frequency dependence of Moho reflections: Figure from Braile and Chaing (1986) showing synthetic models of three Moho transition zones and the vertical incidence reflections expected from them for three wavelets with different frequencies. All amplitude scaling is constant. The models show the frequency dependency of reflections from a velocity gradient and the differing interference effects for each wavelet from a layered transition zone. Note, however, that the layered transition zone is imaged by all three wavelets whereas the gradient is barely imaged on the higher frequence data.

Gravity Modelling

Further constraints can be placed on the crustal structure by gravity modelling. Two-dimensional rather than three-dimensional modelling was performed because the major features of the gravity field are consistent along strike and because little control is available on the lateral continuity of the structural features interpreted on the seismic data. Inspection of the gravity field suggests that the two-dimensional approximation should be adequate in the Valley and Ridge, Blue Ridge, and Piedmont Provinces. The Coastal Plain region, on the other hand, has a gravity field which changes significantly along strike, presumably indicating subsurface masses which likewise vary. In particular, granites beneath the cities of Petersburg (Calver and others, 1963) and Portsmouth, Virginia (Russell and others, 1985) are coincident with elliptical gravity anomalies (Johnson, 1977) which may indicate an ovate shape to the intrusives. Results of the modelling of the eastern portion of the profile, therefore, will be useful only in a general sense.

A 400-km long transect nearly perpendicular to strike of the surface geologic units and coincident with the location of the seismic reflection line was chosen for modelling (line A-B in Figure 3). The simple Bouguer gravity field along this line was taken at 10 km intervals from the maps of Johnson (1971, 1972, 1973, and 1975) and a geologic cross section was constructed based on the interpretation of the seismic reflection data. The gravitational field associated with the two-dimensional model was then computed using the method of Talwani and others (1959) and

aligned with the observed field by subtracting a constant which equalized the values at the 110 km point of the transect. An approximate crustal profile outside the limits of profile A-B on Figure 3 was found neccesary to reduce edge effects and, at points beyond the control derived from James and others (1968), consisted of a 42-km thick crust beneath the Allegheny Plateau, a gently thinning crust beneath the offshore continental margin, and a thickening wedge of Atlantic Coastal Plain sediments offshore (Grow and others, 1979).

The first model (Figure 12, Model A) consists of a uniform crust except for mass units representing the Moho profile, the Atlantic Coastal Plain sediments, the Mesozoic basin imaged on the reflection data, and a light mass at the eastern edge of the model (the Portsmouth granite and perhaps other Mesozoic basins). The outlines of the first three of these units are well constrained by the seismic data; the shape of the latter was determined by altering the outline until the model's gravity field matched the observed data. The model reproduced the observed gravity to within 10 mgals across the entire profile except for the local anomalies coincident with exposures of the Catoctin metavolcanic rocks and the Petersburg granite, and a positive anomaly which probably results from mafic rocks in the basement beneath the Coastal Plain sediments (Gleason, 1979; Johnson, 1973 and 1975).

The model demonstrates that major crustal units of differing density, such as suture zones or other such masses, are not required to reproduce the main features of the gravity profile across the Valley and Ridge, Blue Ridge, and Piedmont Provinces, a result which confirms conclusions from simpler models (Cook, 1984a for example). The eastern portion of the transect, on the other hand, has significantly lower gravity values than those produced by the crustal profile derived from the James and others' (1968) refraction data. To compensate for this in the model requires either a thicker crust than interpreted from the refraction data or, as in the model, a significant amount of lighter upper crustal material in the crust such as granitic bodies. The confirmation of the refraction profile by agreement with the other seismic data and by the match with the gravity field across most of the model suggests that lighter material in the crust is the correct explanation. Drills penetrating the Coastal Plain sediments have reached a basement of light granitoids and Mesozoic sediments in several widely scattered areas (Russell and others, 1985; Johnson, 1973 and 1975;

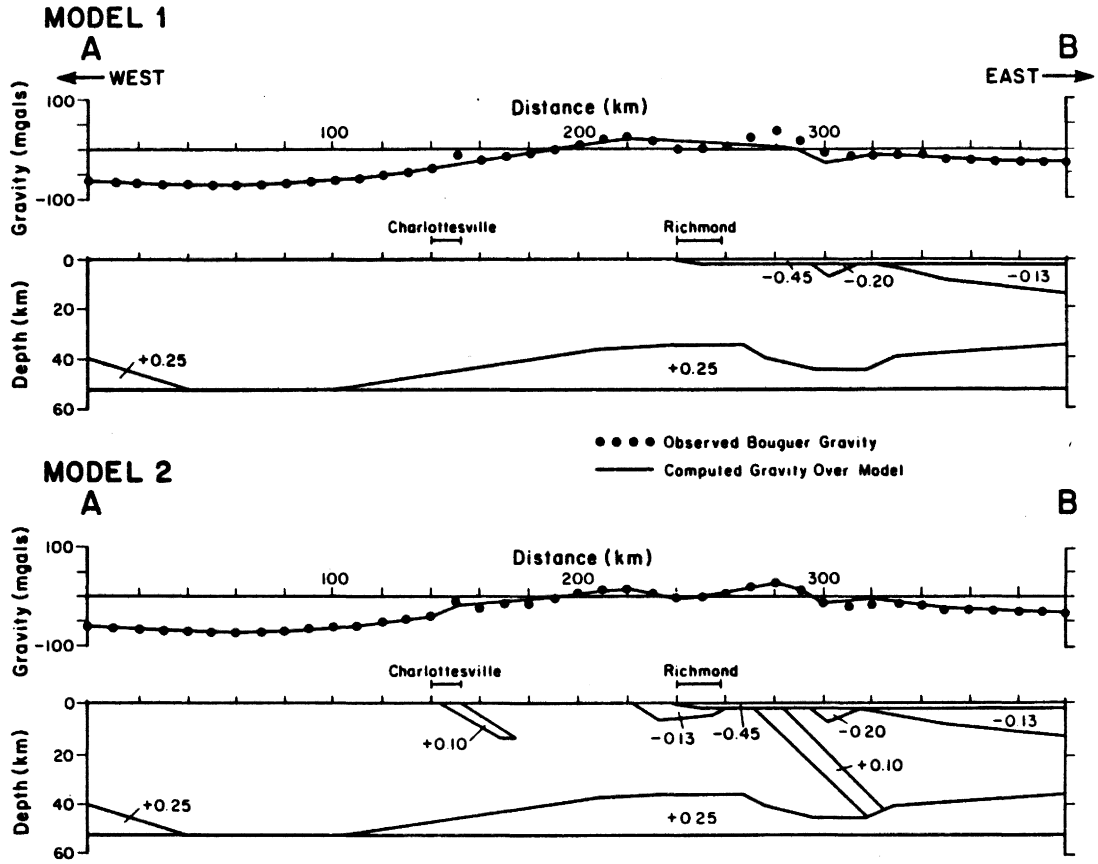

Figure 12. **Gravity model of the Virginia crust:** A) Model of the crust in central Virginia. Moho profile was taken from James and others (1968) after shifting it eastward about 17.5 km (see text). Anomalous bodies represent the Atlantic Coastal Plain sediments (thickness exagerated in the figure for clarity), a buried Mesozoic basin, and a large granitic body (Portsmouth Granite).
B) Same as previous model except for additional bodies representing the Petersburg granite, Catoctin metabasaltic rocks, and "mafic rocks" beneath the Atlantic Coastal Plain sediments.

Gleason, 1979), providing further support for this hypothesis. The thickening of the lighter material on the eastern edge of the model, required to match the gravity values, however, is suspicious because it mirrors the rising Moho profile. This may be indicating that the sloping upper crustal body is compensating for an incorrect Moho profile in the model. Obviously, more control on the crustal structure is required to resolve this issue.

An important result from this exercise is that a density contrast of only 0.25 gm/cm^3 between the upper mantle and lower crustal rocks produced the best fit to the observed gravity field. Higher contrasts, because of the general eastward thinning of the crust, produce an unacceptably high gravity gradient between the Valley and Ridge and Coastal Plain Provinces, which would in turn require a progressively larger amount of light material in the crust toward the eastern portion of the profile. This latter alternative cannot be discarded, but the model with the 0.25 gm/cm^3 crust-mantle density contrast is more pleasing because of its simplicity.

This relatively small contrast may be an indication of a lower crust with a higher density than the upper crust. Chapman's (1979) 6.5 km/sec mid-crustal velocity lends strong support to this hypothesis and is otherwise unexplained. Laboratory measurements and other gravity modelling place the density of the upper crust at about about 2.77 gm/cm^3 (Keller and others, 1985 and references therein). For the upper mantle velocity of 8.15 km/sec measured by Chapman (1979) and James and others (1968), results from laboratory measurements predict a density of 3.25 gm/cm^3 or greater (Birch, 1960 and 1961). The 0.25 gm/cm^3 lower crust-upper mantle density contrast therefore indicates that the lower crust has a minimum density of about 3.0 gm/cm^3, or about 0.23 gm/cm^3 greater than the upper crust.

During the modelling, the model's gravity field is matched with the observed field by removal of a constant value. This makes the modelling procedure completely insensitive to density changes which occur as a uniform horizontal layer because the contribution to the gravity field from such a body is constant across the profile. In other words, a "standard" horizontally layered crust of unknown density structure is assumed, and only perturbations to this crustal profile which produce laterally variant anomalies are modelled. A density increase in the lower crust would therefore need to occur as a nearly horizontal density stratification to not contradict the gravity modelling. The

exception to this is in the Atlantic Coastal Plain Province where lighter material is required and could lie in the lower crust.

The suggestion by Hatcher and Zietz (1979) that a thick mafic crust underlies large areas of the southeast may be correct in view of these results, but a simple density gradient which increased with depth is also consistent with the evidence and does not require major changes in crustal composition. Dense mafic intrusions or depletion of light material from the lower crust by partial melting and intrusion into the upper crust, though they may be important factors in causing the lower crustal reflectivity (see discussion in the next section), either have a minor effect on lower crustal densities or are distributed such that their overall contribution to the anomalous gravity field is small. The latter possibility seems fortuitous at first glance because it requires an even distribution of denser material across the profile which the lower crustal reflections, because they are concentrated in one region, contradict. If the discrepency between the refraction data and wide-angle reflections on the western end of the profile (Figure 4) is indicative of a transition zone, however, a consistent layer of denser material may be present in the lowermost crust which would also be consistent with the gravity modelling. More detailed velocity information on lower crustal velocities is needed to resolve these issues and could make significant contributions to our understanding of the lower crust.

A second model (Figure 12B) was easily constructed whose computed gravity field matched the observed gravity profile to within 10 milligals. This model was identical to the first except for additional masses representing the Catoctin metavolcanic rocks, the Petersburg granite, and the mafic rocks beneath the Atlantic Coastal Plain sediments. The shapes of the first two of these masses are constrained to some extent by mapped surface exposures and by the seismic reflection data; the outlines used in the model are probably the minimum sizes of these bodies. The distribution of mafic rocks is totally unconstrained except for several drill holes which penetrated the Coastal Plain sediments and bottomed in "biotite schist" and "metavolcanic rocks" (Johnson, 1973 and 1975; Gleason, 1979). The outline of the mafic rocks shown on the model is made under the assumption that they penetrate the entire crust at about the same angle as the east-dipping events imaged to the west of them. A strong magnetic anomaly (Figure 13) is coincident with the top of

this mafic body and is consistent with a relatively narrow body extending for about 175 km along strike.

Other, larger borders to these rock units were constructed which create a feasible solution, but they are more complicated and require changes in the density contrasts or the addition of extra compensating bodies. As Keller and others' (1985, their Table 1) also found, the only units in the Piedmont which seem to have anomalous bulk densities are the granitic bodies, the Mesozoic sediments, and possibly the Catoctin metavolcanic rocks. The bulk densities of the other units are apparently similar and therefore can not be used effectively to constrain the interpretation by gravity methods.

The gravity modelling thus demonstrates that the main features of the gravity field can be explained by the Moho profile using a crust-mantle density contrast of about 0.25 gm/cm^3. Anomalous crustal rocks are required only in the Atlantic Coastal Plain Province where the model predicts that less dense rocks are present. Drill holes in the Coastal Plain area, consistent with this result, often enter basement rocks composed of relatively light granitoids and Triassic sediments (Johnson, 1973 and 1975; Russell and others, 1985).

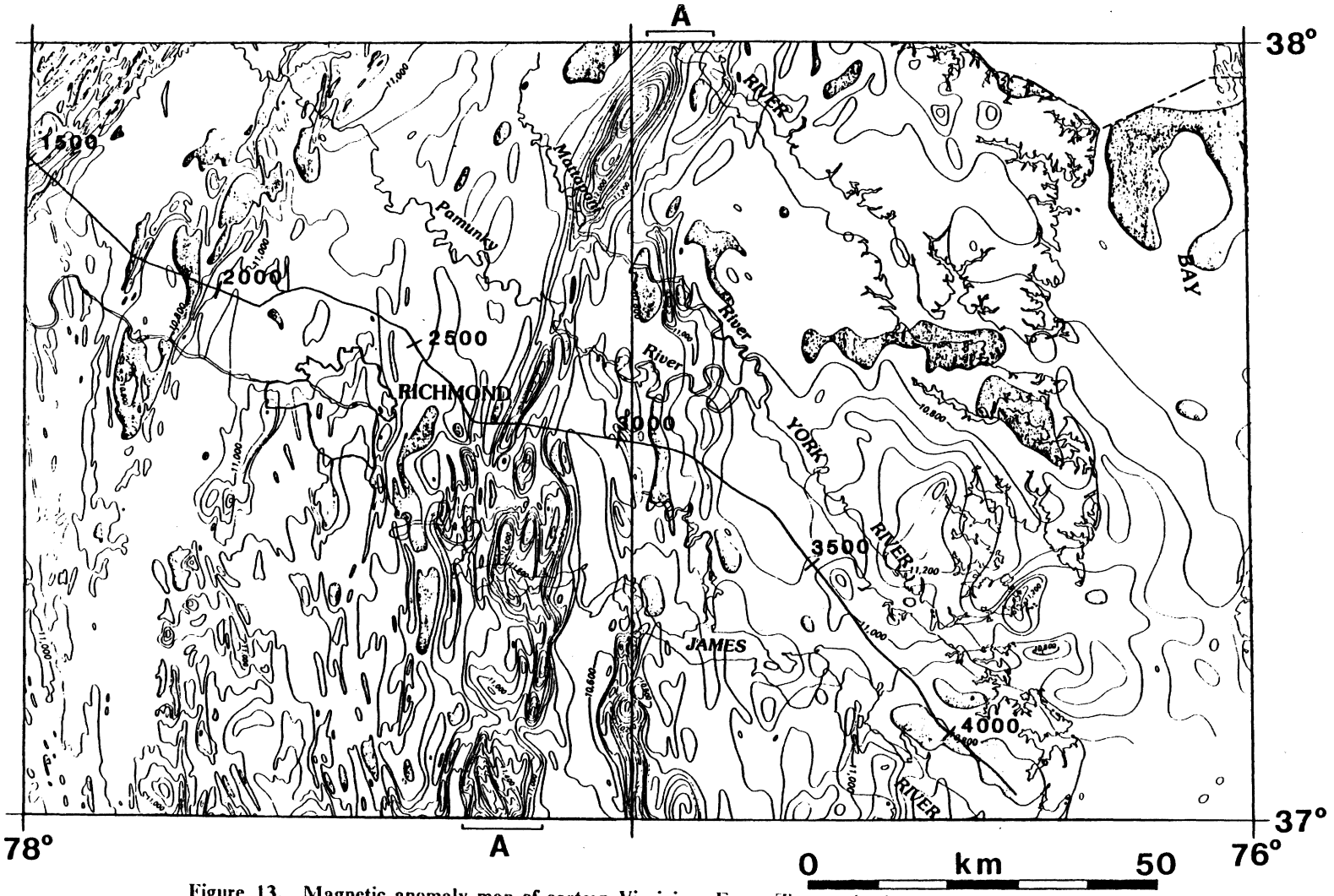

Figure 13. Magnetic anomaly map of eastern Virginia: From Zietz and others (1977) showing the strong north-south magnetic anomaly (A) coincident with the mafic body outlined in the gravity model (Model 2 in Figure 12). Contours are in hundreds of gammas with low values shaded.

Discussion

Implications for the Lower Crust

Meissner and Wever (1986) suggest that "reflections in the crust seem to represent the latest tectonic event." Above the Moho reflections on the I64 data, the rocks at the surface have undergone amphibolite-grade metamorphism at pressures (5-7 kbars) indicative of burial to depths of up to 17-25 km during the Alleghanian event (Farrar, 1984 and 1985), with cooling occurring soon thereafter (Farrar, 1985; Gates and others, 1986b). Also interpreted from the surface rocks above the Moho reflections is Carboniferous strike-slip movement on the Hylas and other ductile deformation zones (Gates and others, 1986a; Gates and Glover, 1986), possibly related to escape tectonics during more northerly collision (Burke and Sengor, 1986; Gates and others, 1986b). These, and the presence of the Mesozoic rift basins, suggests that the area has undergone strike-slip motion, uplift, and initiation of rifting in the past 300 million years. The region has been relatively quiescent since the Mesozoic rifting event.

The Moho reflections on the I64 seismic reflection data thus follow the distribution pattern found in other areas; namely, strong reflections beneath the most recently active regions of the crust, particularly areas of extension (Meissner and Wever, 1986; Smithson and Johnson, 1986; Bamford and Prodehl, 1977). Stable craton regions apparently have a smoother crust-mantle transition,

though the discontinuity is still evident on refraction and wide-angle reflection data (this study; Braile and Chiang, 1986; Bamford and Prodehl, 1977). These observations imply that the lower crustal layering is a relatively short-lived phonomenon which fades to a more transitional boundary, a second order discontinuity on reflection data, beneath the older crust. Possible models are summarized in Klemperer and others (1986) and Matthews and Cheadle (1986) and include:

1) The Moho layering is related to anisotropy, possibly due to shearing at the crust-mantle boundary (Braile and Chaing, 1986; Chowdhury and Hargraves, 1981). The material eventually equilibrates to form a second order discontinuity on the reflection data beneath cratons.

2) Reflections may be from compositional layering originally formed in active regions by intrusion of basaltic (mafic) magma into the lower crust (McKenzie, 1984), accretion of material to its base (underplating) (Fyfe, 1986; Furlong and Fountain, 1986), or metasedimentary layers (Hale and Thompson, 1982). A relatively viscous zone in the lower crust (mechanical mixing) or mobility of elements between layers (chemical mixing) eventually redistributes the material in the layering to create a smoother crust-mantle transition.

3) The layering is a temperature effect creating layers of differing viscosity (melts) but of nearly the same composition throughout. This could be caused by differential melting of material in the lower crustal region (Meissner and Wever, 1986). Differential melting could also remove some of the lighter material from the lower crust via intrusion into the upper crust. Eventually, phase equilibrium is reached and the sharper contrasts are lost.

4) Fluids could enter the lower crust via several mechanisms (Fyfe, 1986) and produce the density and velocity contrasts necessary to create the deep reflections (Matthews and Cheadle, 1986; Hall, 1986). This model is appealing because it accounts for the high electrical conductivity of deep crustal rocks (Shakland and Ander, 1983). The fluids could later be removed by metamorphism or inclusion in intrusive magmas.

Several observations can be made about the deep crustal reflections on the seismic reflection data which constrain some aspects of the above models. First, comparison of the refraction and

reflection data indicates that mantle velocities (8.15 km/sec) are reached only at the base of the layered sequence, not within or above the layering. The velocities within the layered sequence must therefore be lower crustal (6.5 km/sec) or gradational between lower crustal and upper mantle. This observation rules out a simple underplating model, which would be expected to produce upper mantle velocities within the layering (Furlong and Fountain, 1986), or requires that the underplating material not have a mantle composition.

The layering has an indistinct upper surface but a comparatively sharp, well defined lower boundary. Also, the reflection strength appears to generally increase downward; the brightest reflections are in the lower half of the sequence whereas reflections from the upper part of the layering are weak. This reflector geometry requires an explanation which tends to concentrate the reflectors in the lowermost portion of the crust. Fluids, for instance, would have to preferentially collect in the lower portion, requiring them to either move downward through the crust or move up from the upper mantle and accumulate in this zone. Likewise, if the layering does decrease in amplitude with age, the youngest reflectors evidently form at the bottom of the crust.

If the layering is due to motion in a viscous lowermost crust, the last observation would seem to require an asymmetric movement within the zone, with maximum shearing occurring at the lower boundary of the viscous zone where the brightest reflections are. The upper mantle material either did not undergo strain to produce the layering, or it has not been permanently recorded. The lateral decay in reflection amplitudes also implies a similar decay in the motion over several tens of kilometers.

The layering also has a slight wedge-shaped appearance to it, thickest in the eastern portion where the Moho is shallowest and thinning and fading in amplitude to the west, beneath the thickest crust. The obvious conclusion from this observaton is that the layering and crustal thinning are interrelated, the layering a byproduct, or the cause, of uplift.

Though none of the above models may be eliminated as a cause of these deep, layered reflections, the model which seems most consistent with all of these observations collectively is partial melting of, and/or intrusion of upper mantle material into, the lower crust during uplift. Assuming the uplift indicated in the surface rocks involved the entire thickness of the crust (which the Moho

rise suggests), the lower crustal and upper mantle material would undergo a decrease in pressure at a nearly constant temperature, which may result in melting as the material crosses the solidus line in pressure-temperature (P-T) space (Furlong and Fountain, 1986). The subhorizontal attitude of the reflections would then be attributable to pressure effects determined by the gravity field (Meissner and Wever, 1986). The Alleghanian plutons and Mesozoic dikes (Speer and others, 1979; Fullagar and Butler, 1979; Ragland and others, 1983) indicate that lower crustal-upper mantle melting was indeed occurring at that time. If the discrepency between Chapman's (1979) and James and others' (1968) depths on the western portion of the section is due to an older transition zone which was once similar to the highly reflective lower crust farther east, it indicates a lower crust at least partly contaminated by relatively high velocity upper mantle material. If the layering seen on the reflection profile has persisted since the Mesozoic event, it would suggest either a very long freezing time for the melted material or, more likely, the necessity of another tectonic event to decrease the reflectivity.

Implications for the Appalachians

The gross geometry interpreted from the geophysical data in Virginia (Figure 5) consists of a crust which thins eastward to a point beneath the Fall line (western limit of the Atlantic Coastal Plain sediments) and then rethickens to the east. This mantle upwarping is mimicked in the upper crust where the highly reflective zone (B on Figure 4) parallels the eastern upslope of the Moho profile, suggesting a common origin involving uplift or arching of the entire crust. The apparent uplift is greatest at the exposed basement terrane (Goochland Terrane), consistent with geologic data which show a 6 to 8 km uplift of the basement rocks relative to the surrounding rocks (Gates and others, 1986b). The eastern edge of the uplift may be bounded by the zone of east-dipping events on the profile. These same dipping events appear to reach the Moho where the crust rethickens.

The results from gravity modelling demonstrate that in Virginia the Piedmont gravity high may be primarily due to the thinner crust. Reversing this logic, an indication of the continuity of this crustal profile along strike can be tested by examination of gravity data along the length of the Appalachians. Figure 15 is a map of the eastern United States with the Piedmont gravity high shown as a dot pattern, and Figure 2 shows gravity profiles across the high at the 9 locations marked on Figure 15. The gravity high varies in amplitude and width along the southern and central Appalachians, leading some authors to conclude that there are major changes in crustal thickness or composition along strike (Cook, 1984a). The large amounts of Carolina Slate Belt metavolcanic rocks and granitoids which crop out in the Piedmont (Williams, 1978) and the variations in crustal thicknesses interpreted from the refraction data (James and others, 1968) are two such features which obviously effect the gravity field by producing significant local anomalies. Note, for example, the locallized thin crust beneath station FRV on Figure 3 which corresponds with a small eastward extension of the Piedmont gravity high ('A' on Figure 15). This same perturbation is somewhat offset by the presence of the Petersburg granite, which probably causes the elongate negative anomaly encroaching on the gravity high adjacent to the exposed Richmond Mesozoic basin. Likewise, the northernmost profiles in Figure 2 show the effects of a gravity high at Scranton, Pennsylvania on their western side which partially counters the otherwise low values. The persistence of the paired high-low gravity anomalies along strike, however, is indicative of an overall consistent crustal profile only locally perturbed by these irregularities.

Superimposed on the profiles in Figure 2 are the computed gravity values from the model in Figure 12A with (dotted line) and without (dashed line) the light granitic material on the eastern edge. The model profile was aligned with each of the observed gravity profiles to produce the best match between the two. The comparison shows that even a simple model including only the Moho profile and the Atlantic Coastal Plain sediments produces a close match with the paired high-low anomalies along most of the length of the Appalachians. The eastern portions of the profiles generally have Bouguer gravity values between those that the two model variations produce, indicating even thicker crust beneath the Atlantic Coastal Plain sediments than indicated by the model or the

presence of lighter material in the crust. Further work is necessary to determine which of these alternatives is responsible.

Karner and Watts (1983) recently presented a crustal model which invoked flexure of the lithosphere at a continental collision zone, with related surface and subsurface loads, to explain the paired high-low gravity profile over the Appalachians. The model consists of a subducted plate which is bent downward, creating thicker crust at the collision zone and a flexural bulge inland from it. The obducting plate has the opposite geometry, being bent upward at the collision zone. Filling in the spaces in front of and on the obducted plate are sedimentary basins. The gravity low, according to their model, is caused primarily by a thicker crust where the subducted plate downturns. The gravity high, they predict, is due in a large part to the basement rock of the obducted block (their "subsurface load") displacing lighter upper crustal rocks.

The I64 profile crosses the Appalachians at a location corresponding to the eastern portion of the Karner and Watts (1983) model, beginning near the edge of the thickest crust and extending out onto the proposed obducted block. The arguments presented in this paper, however, are inconsistent with their model in one major respect. The crustal profile derived here, one of thinning beneath the Blue Ridge and Piedmont Provinces and rethickening beneath the Atlantic Coastal Plain, has been shown to adequately explain the gravity anomalies with relatively minor alteration. One of the conclusions from this work, therefore, is that the gravity high is due mostly to crustal thinning and that major intra-crustal density anomalies are probably not present. Karner and Watts (1983), in contrast, predict that the gravity high is caused primarily by the "subsurface load" created by the obducted block and its relatively high density in comparison to the surrounding rocks.

There are two less important disagreements between the Karner and Watts (1983) model and the geophysical data presented in this paper. First, the crust-mantle density contrast of 0.6 gm/cm^3 which they used is large in comparison to that derived by the gravity modelling done here. Small modifications to the densities used in their model could probably remedy this discrepency. Second, the simple geometry of the obducted plate in Karner and Watts' (1983) model is questionable given the zones of east-dipping events imaged on the seismic reflection profiles along the entire length of the Appalachians. There is obviously a lot more complexity involved than their model takes into

account because of the multiple tectonic events which have affected the Appalachians. The reason this may be a minor point, however, is that the geometry and complexity of the obducted block (their "subsurface load") has little effect on their main arguments, which rest on the flexure of the subducted plate. The latter point therefore requires a modification of their subsurface load but the basic principles behind their model remain intact.

The results of this study thus contradict some aspects of the Karner and Watts (1983) model. The main concept of their model, the flexure of the subducted lithosphere, is untested by the geophysical data presented in this paper, but an alternative hypothesis for the cause of the gravity high has been presented which differs from their explanation. Incorporating the arguments presented here into their model, perhaps by substituting the crustal thinning for the subsurface load as the cause of the gravity high, may be possible.

Regional seismic reflection profiles are indicated as heavy lines on Figure 15 with the adjacent dots denoting the easterly zones of dipping reflections seen on those sections (Figure 1). The spatial relationship between the gravity high and the dipping events along the entire length of the Appalachians is striking and certainly supports a relationship between the two features. The geometry suggested here is that, as in Virginia, there is mantle upwarp, and therefore thinner crust, adjacent to (immediately west of) the dipping events.

The thinning of the crust between the Blue Ridge and Piedmont is well documented on refraction data (Figure 14) and is probably the cause of the steep gravity gradient (see reviews in Cook and others (1984a or 1984b) and in Hutchinson and others (1983)), but little control on the Moho depths beneath the Piedmont is available from refraction data outside of Virginia. The refraction data are therefore inconclusive regarding a rethickening of the crust east of the Piedmont high, but what evidence there is suggests a generally thicker crust beneath the Atlantic Coastal Plain Province than beneath the Piedmont (Figure 14). Two other authors, however, also found it convenient to invoke crustal thinning below the gravity high when modelling the crust in Georgia using gravity data (Iverson and Smithson, 1983a; Long, 1979).

Only a few segments of Moho reflections are seen on the multichannel data (Nelson and others, 1985a; Cook and others, 1979 and 1981) but slightly longer travel times to the Moho re-

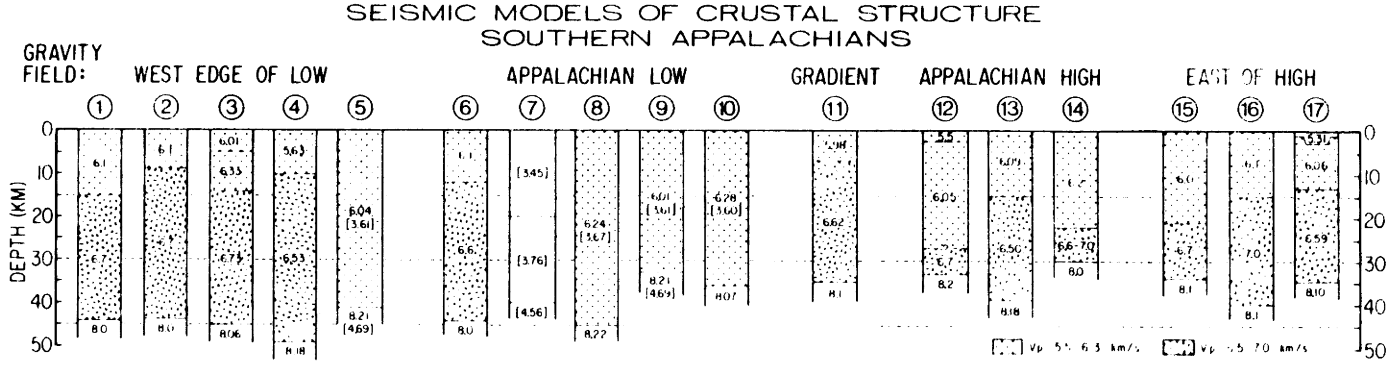

Figure 14. Crustal thicknesses from refraction data: Figure from Hutchinson and others (1983) showing a compilation of crustal thicknesses determined from refraction data in the Appalachians, grouped by provinces. Note that the crustal thicknesses in the Atlantic Coastal Plain Province are generally thicker than those in the Piedmont Province. A crustal thinning between the Valley and Ridge and Piedmont provinces is also evident. References are given in Hutchinson and others (1983).

flectors seem to be indicated beneath the highly reflective zones of dipping events on both southern Appalachian COCORP (Figure 1) (Cook and others, 1981; Nelson and others, 1985a). Cook and others (1981) show that this apparently thicker crust coincides with a gravity low (their Figure 3), creating a geometry similar to that in Virginia. Iso-velocity contours of their stacking velocities, which indicate up to a 10%velocity decrease, however, led them to conclude that upper crustal density deficiencies (due to metasedimentary rocks) and related velocity pulldown are an equally probable explanation for the observations. Iverson and Smithson (1983b) disagree with this interpretation because they found generally high stacking velocities in this region when reprocessing the COCORP data. In Virginia, moreover, the rocks above the apparently thicker crust would need a velocity almost 25% lower than the surrounding rocks if the apparent rethickening is due entirely to velocity pulldown. There is no evidence at present for a velocity decrease of this size. An anomaly of similar magnitude appears necessary to explain the apparent crustal rethickening on the COCORP line, but this is difficult to determine without the original data. Better control on the velocities and crustal thicknesses across the region, perhaps from refraction or wide angle reflection experiments, is neccessary to determine the amount of velocity pulldown which is occurring, if any.

Fault systems in the south-central Appalachians also exhibit continuity along most of the Piedmont gravity high even though the rocks exposed at the surface are often oblique to it (note the lack of correspondence between the gravity high and the Blue-Green-Long axis in Figure 15). Because high-angle faults can cut through overlying thrust and nappe structures, they could conceivably provide a more direct indicator of deeper crustal trends than the surface rocks, which may have been transported. The eastern Piedmont fault system (Hatcher and others, 1977), of which the Hylas zone forms the northern part, is one such group which is proximal to the zones of east-dipping events on the seismic sections (Cook and others, 1981; Nelson and others, 1985a; Behrendt, 1986; this study) and may in fact be part of their surface expression. The gravity high is also flanked on both sides along most of its length by Mesozoic normal faults and their associated sedimentary basins (Figure 15). The Mesozoic basins persist beneath the Atlantic Coastal Plain and offshore regions, but their western limit is strikingly coincident with the gravity high. Uplifted Grenville basement rocks occur in at least two positions in the Piedmont, the Goochland-Raleigh belts in

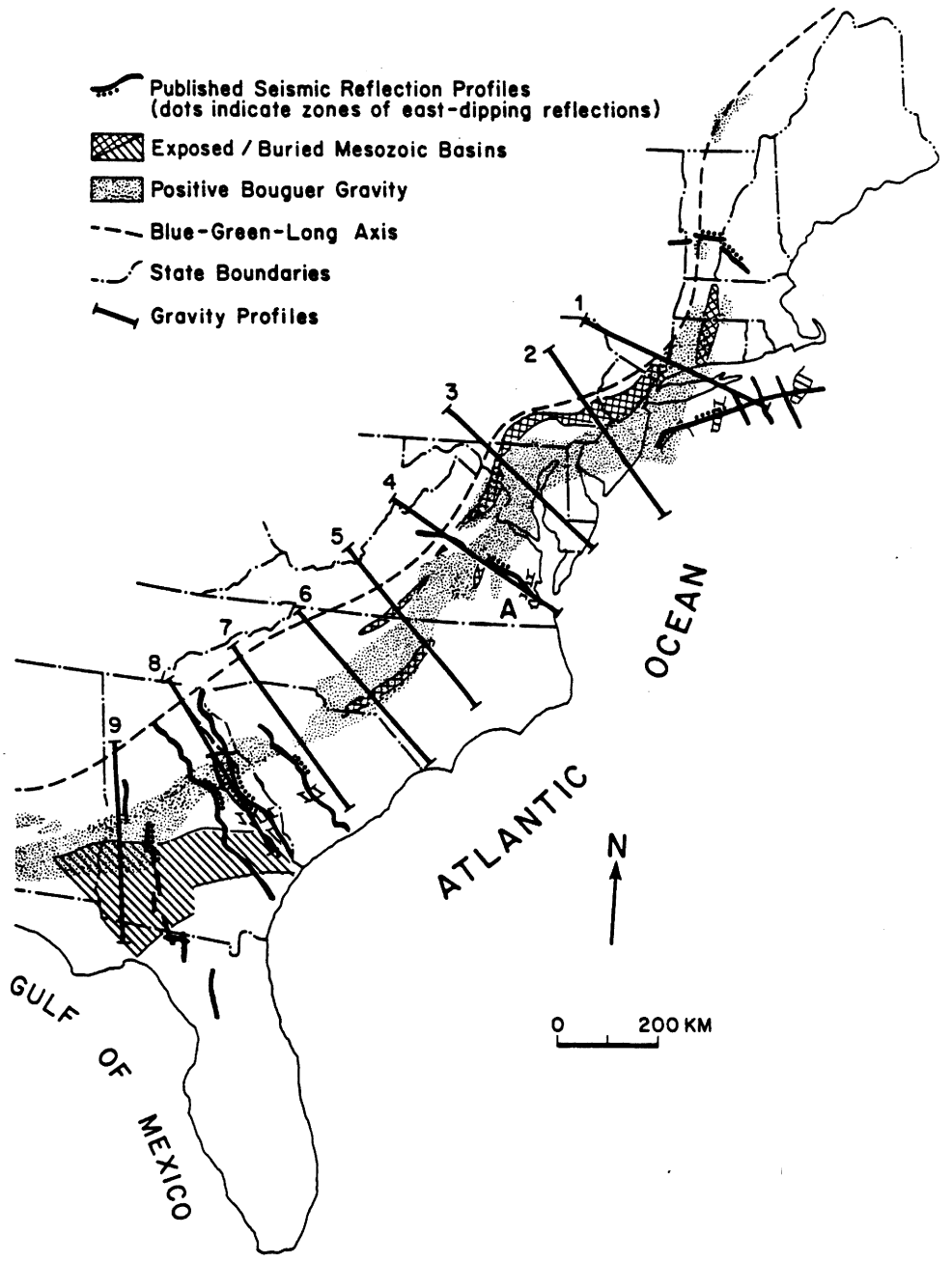

Figure 15. Regional map of the south-central Appalachians: Map of the south-central Appalachians showing regional seismic reflection profiles (see references in introduction to this paper), the locations of major zones of east-dipping events on the profiles, the Blue-Green-Long axis (Rankin, 1976), the larger Mesozoic rift basins (Swanson, 1986; Hutchinson and others, 1986b; Williams, 1978; this study), and the Piedmont gravity high (outlined with the 0 mgal contour on Haworth and others, 1980). 'A' is a perturbation in the gravity field discussed in the text. Lines 1-9 are the locations of the gravity profiles shown in Figure 2.

Virginia and North Carolina (Farrar, 1984 and 1985) and the Pine Mountain Belt in Georgia (Schamel and others, 1980; Nelson and others, 1985a), both of which coincide with the gravity high.

These features, all seemingly related, can be interpreted as marking a fundamental zone of crustal weakness which, because of the range of ages of these features (Ordovician to Triassic), may have persisted since the Precambrian. As mentioned earlier, an interpretation whereby these dipping events and the apparently related gravity high mark the position of the late Precambrian-early Paleozoic continental margin are favored here. The continental hinge zone, by necessity, was the site of the Taconic collision zone in the Ordovician when island-arc volcanic rocks (Chopawamsic and Carolina Slate Belt metavolcanic rocks) were apparently accreted to North America, with some of them being thrust to a position west of the collision zone (Bland and Blackburn, 1979). The variation in distance between the Blue-Green-Long axis, which may be the ancient continental edge (Wehr and Glover, 1985), and the dipping events on the sections (with the related gravity high) is likely a result of differential transport of the exposed basement and related cover rocks.

The old continental edge continued as a line of weakness during subsequent tectonic events. As mentioned earlier, recently interpreted evidence in the Virginia region indicates the the latest events have included uplift along this zone during the transpressional Alleghanian event. This ductile event continued into the Triassic when brittle normal faulting opened rift basins on the flanks of the uplift (Gates and others, 1986b).

At least two different origins for uplift along the gravity high are suggested here. The first of these is a breaking of the crust during initial Triassic rifting, possibly via a simple-shear mechanism (Wernicke, 1985) with the shear planes occurring within the already existing crustal weakness marked by the zone of east-dipping events. The actual uplift may be due to mantle processes, a heating of the lower crust, or isostatic rise during unloading in the initial stages of rifting. Supporting this suggestion, the western edge of the dipping reflections, the Hylas zone, forms the western border of the Richmond Mesozoic basin (Bobyarchick and Glover, 1979), and is downdropped to the east. The numerous other Mesozoic basins exposed east of the Richmond basin and buried beneath the Atlantic Coastal Plain sediments indicate that the faulting occurred along a substantial zone, not a single surface.

Another possible alternative is that the uplift is related to the strike-slip movement recorded on the Hylas zone and other fault zones in the area (Gates and others, 1986b). This hypothesis is consistent with the simultaneous ages of movement on the Hylas zone and cooling of the basement rocks which indicate that uplift occurred during the transpressional strike-slip motion. Mesozoic basins in the region may be the latest phase in the uplift event as it gave way to tensional brittle faulting and are thus related to the gravity high.

The gravity high deviates from this pattern in two locations. In the Connecticut region the anomaly bifurcates, with one arm of the anomaly continuing northward into Vermont where east-dipping events and the Taconic Allochthon are imaged on the COCORP line acquired there (Ando and others, 1983 and 1984). The other portion of the gravity high continues in a northeastward direction where it remains coincident with the Mosozoic faults and sedimentary basins. In Georgia, where the effects of a collision with a southern landmass may also be present, the gravity high cuts obliquely across the surface faults and Mesozoic basins. Studies of the anomaly and its related structures in these regions may, because of their apparent separation, be more decisive as to the exact origin of the proposed mantle upwarping and related gravity high.

Conclusions

The United States Geological survey Interstate 64 seismic reflection line across Virginia was reprocessed at Virginia Tech using an extended correlation technique to produce 14-sec records. The line is the cornerstone of an integrated study of the Virginia crust in which geologic, gravity, and seismic refraction data were also incorporated in the interpretation.

The interpretation of the upper crustal section has the Blue Ridge as a relatively shallow thrust sheet ($\cong$3 km in thickness) possibly underlain by other thrust sheets, with autochthonous Grenville basement encountered at about 3 sec (9 km) on the section. These thrust structures turn down on the east side of the Blue Ridge anticlinorium where east-dipping reflections penetrate to mid-crustal levels, but the data are too ambiguous for a detailed interpretation of this region. The Evington Group metasedimentary and Chopawamsic metavolcanic rocks are highly reflective on the section and appear to occupy a fault-bounded synformal structure reaching depths of up to 10 km. The Goochland terrane is interpreted to be a nappe structure which overrode the Chopawamsic rocks from the east. East of the Goochland terrane, a zone of prominent east-dipping reflections are seen on the section which appear to penetrate to Moho depths, although an apparent data loss on the eastern third of the line obscures the lowermost portions of these events. Grenville basement is relatively transparent on the section and may core an arch-like structure in the mid-crust with a structural high at the Goochland terrane.

Strong reflections in the 9 to 12 sec range on the section are interpreted as lower crustal layering about 5-10 km in thickness. They have a well defined base to them which almost exactly coincides with the Mohorovicic discontinuity derived from earlier refraction work. Arrivals interpreted to be wide-angle Moho reflections were interpreted by other workers on seismograph records of quarry blasts in the western portion of the state. These arrivals are consistent with a discontinuity approximately 8-9 km shallower than indicated on the refraction data, and may be indicating a second-order lower crustal transition zone not imaged on the seismic reflection data. This lower crustal transition zone is interpreted as being due to mafic intrusions and/or partial melting in the lower crust which has equilibrated to a second-order discontinuity beneath the craton in the western portion of the state.

The Mohorovicic discontinuity is interpreted on refraction data by earlier workers to be about 55 km deep beneath the Valley and Ridge Province, reaches a shallow level ($\cong$35 km) beneath the Goochland terrane, and then rethickens to the east. This profile mimicks the arching seen at mid-crustal levels on the I64 reflection data. Results of gravity modelling suggest that this Moho profile is responsible for the main features of the gravity field, namely the Appalachian gravity low, the Piedmont gravity high, and the gravity gradient between them.

The mantle upwarp on the Moho profile, and the overlying crustal arching, are bounded on the east by the pronounced zone of east-dipping reflections. The continuity of this crustal geometry is indicated on regional maps of the southeastern United States by the consistency of the Piedmont gravity high and the pervasiveness of the east-dipping events on regional seismic reflection profiles. Coincident with these structures on the surface are the eastern Piedmont fault system and Mesozoic sedimentary rift basins. These features are interpreted as delineating a zone of crustal weakness which was the continental edge in Precambrian times and has persisted as a zone of tectonic activity during subsequent events. The mantle upwarping may be due to simple shearing of the crust along this weak zone during the Mesozoic rifting event, or an effect of Late Paleozoic strike-slip faulting concentrated in this zone.

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Appendix A. Extended Correlation Technique

In a normal correlation, the sweep trace is used to correlate only the reflections in which the full returning sweep is recorded at the surface (Figure 16). Partially recorded reflections (R2 in Figure 16) are ignored during the correlation process. An extended correlation retrieves these reflections.

To accomplish this, several seconds of zeros are appended to the uncorrelated field records. When the data are correlated, the sweep trace is allowed to run off the end of the recorded data and onto the zeros (Figure 16). This allows a correlation to be made with any deep reflections in which the full sweep was not recorded rather than stopping the correlation at the deepest reflector where the entire sweep was allowed to return.

Because only part of the sweep is used in the correlation of the deepest events (after the "normal" data length), this extra data should show a time-varying loss in the frequency content. In the case of an upsweep, the higher frequencies are missing; when using a downsweep, the lower frequencies are not recorded. The energy in this portion of the record has, of course, travelled the farthest distance and therefore suffered the most attenuation. Higher frequencies are attenuated more rapidly in the earth, so the energy reflected from the deeper events should be naturally richer in the lower frequencies. The use of an upsweep, as with the I64 data, is therefore preferable when using this technique.

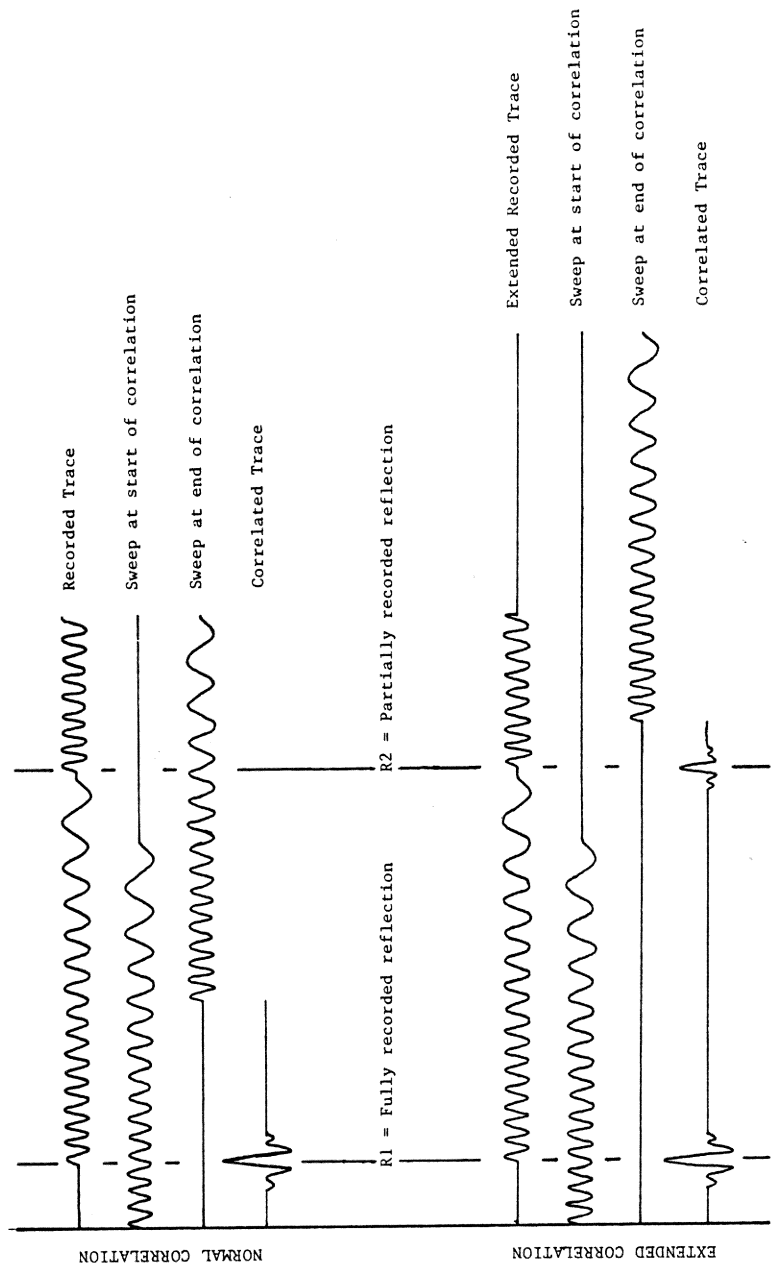

Figure 16. Extended correlation technique: Drawing of a standard correlation and the extended data technique. Explanation is given in text.

In the case of the I64 records, all data above 8 sec contains the full 14-56 Hz bandwidth. Below 8 sec, 4.2 Hz/sec (42 Hz change in a 10 sec sweep) is lost from the upper portion of the bandwidth per second of data. The bandwidth at 11 sec is 14-43.4 Hz, and at 14 sec the data has a 14-30.8 Hz bandwidth (Figure 17). Note that this still retains more than one octave of data at 14 sec and has a full 1.5 octaves where the deepest reflections are recorded at 11 sec.

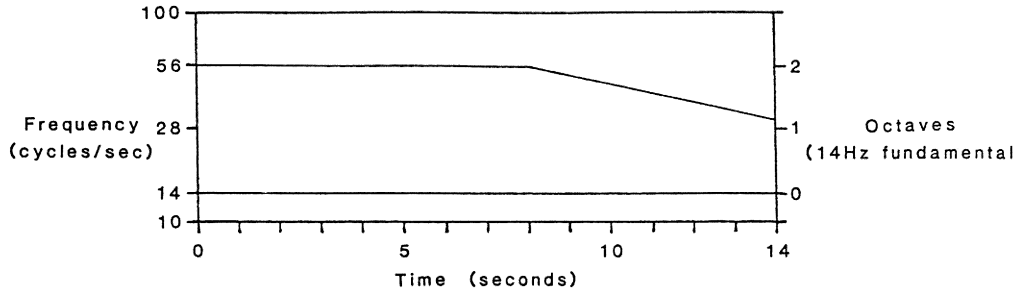

Figure 17. Frequencies in data after correlation: Graph of the frequency content in the I64 data after the extended correlation. The first 8 sec contains the full 14-56 Hz bandwidth (as with a normal correlation), the frequency content then drops off on the high frequency end of the spectrum at a rate of 4.2 Hz per second until there is only a 14-30.8 Hz bandwidth at 14 sec.

Appendix B. Seismic Reflection Data Parameters

Aquisition Parameters

SOURCE PARAMETERS

vibrators	2-3
sweeps per source array	15(3 vibs)-20(2 vibs)
sweep length	10 sec
sweep frequencies	14-56 Hz
taper length	1 sec
source array length	440′ (134 m)
source array spacing	440′ (134 m)

RECORDING PARAMETERS

geophones	GSC-20D (10 Hz)
instruments	DFS V
receiver group spacing	220′ (67 m)
configuration	48 channel, split spread
offsets	1676 - 201 - 201 - 1676 m
	5500 - 660 - 660 - 5500 feet
recording filter	12 - 18 - 90 - 72

Processing Parameters

Demultiplex
AGC 1000
Vibroseis Correlation and data extension
    final data length = 14 sec
Datum statics
    datum elevation = 300 m
    replacement velocity = 5000 m/sec
CDP sort
    station spacing 33.5 m (110′)
    fold = 12
Deconvolution
    gap = 56 ms
    filter length = 160 ms
    design window = 600-4000 ms
    application window = 0-14000 ms
Bandpass filter
    low ramp = 10-14 Hz
    high ramp = 48-56 Hz
Velocity analysis
    multiple analyses alternating with residual statics analyses
Automatic residual statics analysis
    multiple applications alternating with velocity analyses
Mute
Stack
Migration
    post-stack finite difference
    layer thickness = 48 ms
    panel width = 72 traces
    velocity = linear, 5200 m/sec at surface, 6500 m/sec at 14 sec

DISPLAY
    traces/inch = 37
    inches/sec = 2.5
    AGC length = 1000 ms

Appendix C. Seismic Reflection Sections

SEISMIC REFLECTION SECTION ..... IN POCKET

GEOLOGIC CONTACTS

(estimated accuracy ≅ 5 CDP stations (160 meters))

UNIT	CDP Numbers of contacts
Valley and Ridge sediments	43-329
Rome Formation	329-351
Shady	351-370
Chilhowee	370-465
Catoctin Formation	465-630
Swift Run	630-640
Blue Ridge Basement (Pedlar)	640-700
Rockfish Valley Fault	700
Blue Ridge Basement (Lovingston)	700-950
Mechum River Formation	950-980
Blue Ridge Basement (Lovingston)	980-1340
Lynchburg Formation	1340-1442

Catoctin Formation 1442-1715

Evington Group 1715-1830

Hardware Terrane 1830-2425

Chopawamsic Formation ..... 2425-2510?

Granite ..... 2510?-2580?

Chopawamsic Formation ..... 2580?-2734

Arvonia Formation 2734-2850

Chopawamsic Formation 2850-3063

Maidens Gneiss 3063-3532

Sabot Amphibolite 3532-3558

State Farm Gneiss 3558-3810

Sabot Amphibolite 3810-3829

Maidens Gneiss 3829-3882

Hylas Mylonite Zone 3882-3924

Richmond Mesozoic Basin 3924-4006

Petersburg Granite 4006-4284

Atlantic Coastal Plain 4284-7924

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