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A GEOPHYSICAL STUDY OF THE EARTH'S CRUST IN CENTRAL VIRGINIA: IMPLICATIONS FOR APPALACHIAN CRUSTAL STRUCTURE

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Abstract. A regional seismic reflection line (I-64) across the Virginia Piedmont has provided a stacked section suitable for an integrated interpretation of geophysical data in the region. A highly reflective upper crust, an allochthonous Blue Ridge Province, underlying thrust sheets including the Blue Ridge master decollement, and a basal decollement at a depth of about 9 km (3 s) are confirmed on the seismic data. Immediately east of the Blue Ridge Province, Appalachian structures plunge to as much as 12 km (4 s) depth. The Evington Group, Hardware terrane, and Chopawamsic metavolcanic rocks (Carolina terrane) crop out in the Piedmont Province, and numerous eastward dipping reflections originate from these rocks in the subsurface. These eastward dipping reflectors overlie a gently west dipping (10°-15°), highly reflective zone that varies in depth from 1.5 s (4.5 km) beneath the Goochland terrane to 4 s (12 km) beneath the rocks of the Evington Group. Some of the overlying eastward dipping reflections apparently root in this zone. The zone may include decollement surfaces along which the overlying rocks were transported. Relatively few reflections originate from within autochthonous Grenville basement at the western end of the profile. The Goochland granulite terrane is interpreted to be a westward thrust nappe structure that has overridden a portion of the Chopawamsic metavolcanic rocks. A broad zone of east dipping (20°-45°) reflections bounds the Goochland terrane on the east. These reflections may originate from deformation zones and continue to Moho depths. They appear to be correlative with similar events seen on other Appalachian lines. The pervasiveness of the zone of east dipping events on other seismic reflection lines and the continuity of the adjacent Piedmont gravity high suggest continuity of crustal features along the length of the Appalachians. A major conclusion of this study is that crustal thinning is responsible for the main components of the gravity field in Virginia, that is, the Appalachian gravity gradient and the Piedmont gravity high. The crust thins from about 52 km beneath the Appalachian mountains to about 35 km beneath Richmond, Virginia, and then rethickens by up to 10 km beneath the zone of east dipping reflections (mylonites?) east of Richmond. The I-64 seismic data also contain a sequence of reflections at about 9-12 s, indicative of lower crustal layering; the base of this zone of reflections coincides almost exactly with the Mohorovicic discontinuity interpreted from earlier refraction work. The layering extends about 70 km west from Richmond, Virginia, and is interpreted as a lower crustal transition zone that is believed to persist across most of Virginia.

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Introduction

In studies of mountain belts the central and southern Appalachians are unique because of the relatively large amount of subsurface control provided by regional seismic reflection traverses (Figure 1). In the southern Appalachians, two Consortium for Continental Reflection Profiling (COCORP) profiles [Cook et al., 1979, 1981; Nelson et al., 1985a,b] and several deep industry lines [Behrendt, 1985, 1986] extend from the Valley and Ridge Province across the entire Blue Ridge, Piedmont, and Coastal Plain Provinces. These studies have made major contributions to our understanding of the Appalachians and mountain systems in general and have confirmed the importance of thin-skinned horizontal transport of thrust sheets [Cook et al., 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 of the most prominent such features are the pervasive zone of east dipping reflections seen on the seismic reflection profiles [Ando et al., 1984; Iverson, 1986] and the distinctive gravity profile [Hutchinson et al., 1983; Karner and Watts, 1983; Cook, 1984a] that includes the Appalachian gravity gradient and the Piedmont gravity high.

The dipping events have been imaged to some degree on nearly every seismic reflection profile in the internal metamorphic core of the orogen [Ando et al., 1984; Brewer and Smythe, 1984; Iverson, 1986; Nelson et al., 1986]. They are characterized by a broad (tens of kilometers) region of reflections dipping at angles of 20°-45° on the sections (Figure 2). Probably the best example is on the COCORP western Georgia profile [Nelson et al., 1985a,b]. Several authors have speculated that the reflections represent suture zones [Nelson et al., 1985a,b] or a large ramp structure controlled by a buried continental-oceanic crust transition zone

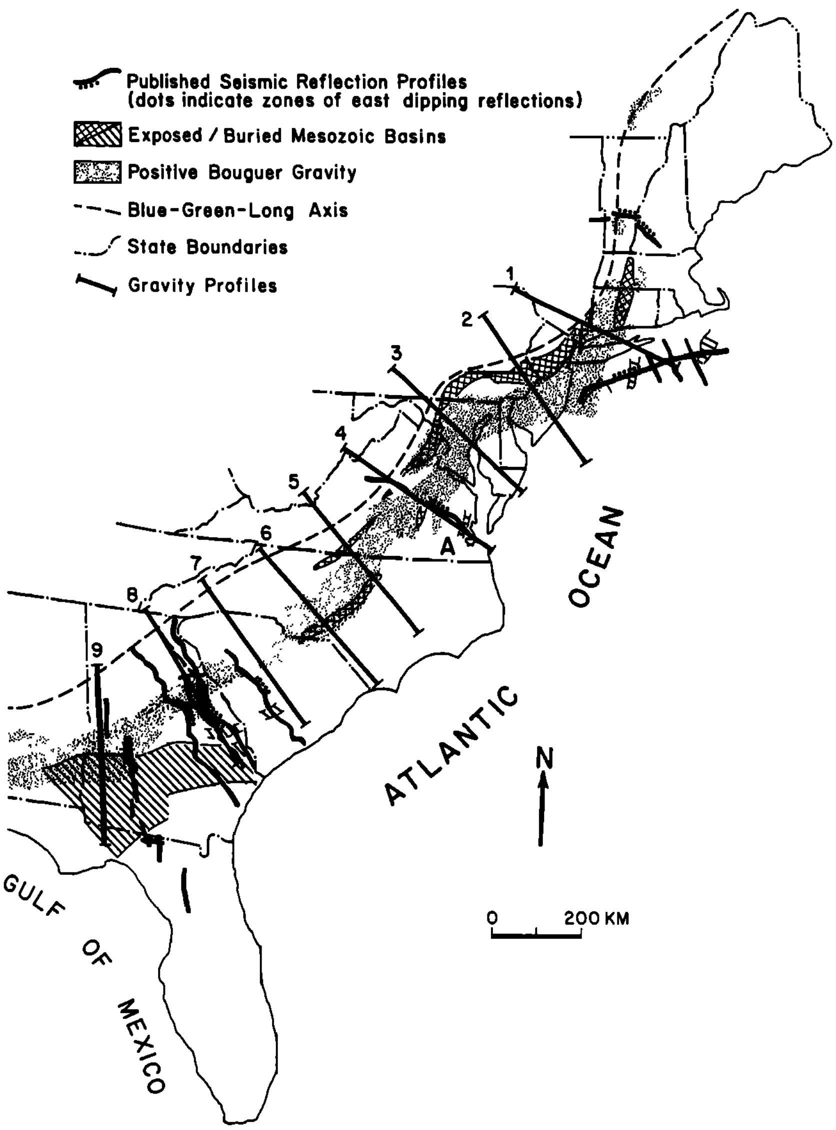

Fig. 1. Map of the eastern United States showing regional seismic reflection profiles (see references in introduction to this paper), the locations of major zones of east dipping events on the profiles, the Blue-Green-Long axis [Rankin, 1976], the larger Mesozoic rift basins [Swanson, 1986; Hutchinson et al., 1986; Williams, 1978; this study], and the Piedmont gravity high (outlined with the 0 mGal contour by Haworth et al. [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 3.

remaining from the original Precambrian rifting event [Ando et al., 1983].

The gravity profile across the Appalachians (Figure 3) 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. Models that have been proposed to explain this gravity profile include a suture zone, a rift, lithospheric flexure, mantle upwarp, and upper crustal effects (see reviews by Hutchinson et al. [1983], and Cook [1984b]). 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 [e.g., Cook, 1984a]. The gravity high is not as easily explained but has been correlated with exposures of Carolina Slate Belt and metavolcanic rocks [Long, 1979], with Mesozoic rift basins [Griscom, 1963], and with the dipping events imaged on the seismic reflection data [Ando et al., 1984; Hutchinson et al., 1986]. Karner and Watts [1983] suggested that subsurface and surface (topographic) loading is the cause of the paired high and low gravity anomalies.

The Virginia region provides an exceptional opportunity to conduct an integrated geophysical study across the metamorphic core of the Appalachians because of the now available reflection seismic, refraction seismic, gravity [Johnson, 1971, 1972, 1973, 1975], and magnetic data. The work of James et al. [1968] is the most comprehensive refraction study of the lower

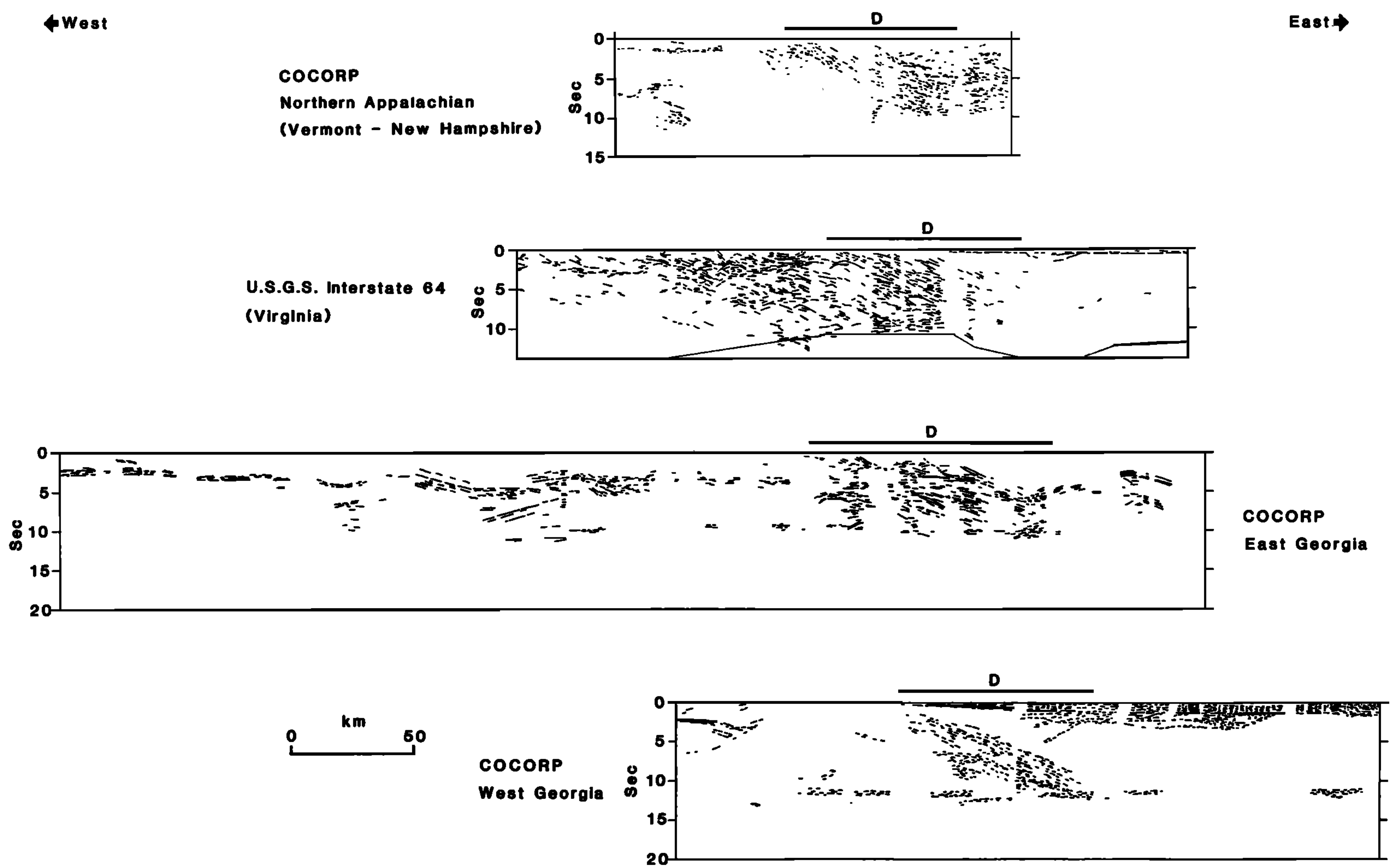

Fig. 2. Line drawings of the four seismic reflection profiles that cross the orogen and image the entire thickness of the crust. From top to bottom: COCORP northern Appalachian lines [Ando et al., 1983, 1984], I-64 line (this study), COCORP east Georgia lines [Cook et al., 1979, 1981], COCORP west Georgia lines [Nelson et al., 1985a, b]. Note the presence of east dipping events reaching deep crustal levels, and possibly penetrating the crust.

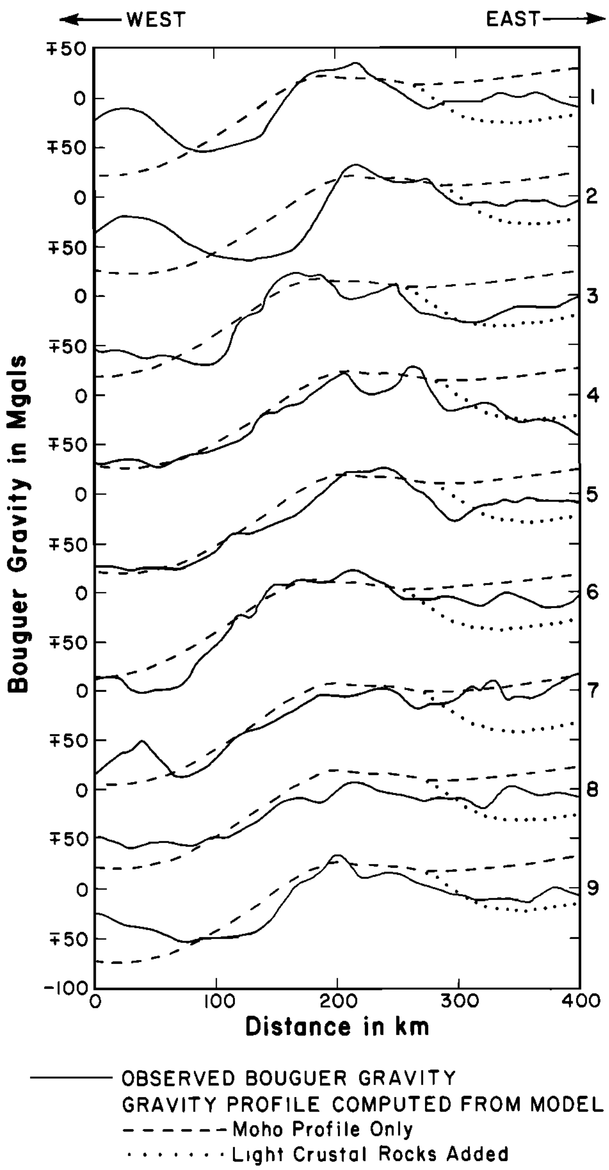

Fig. 3. Nine profiles of the simple Bouguer gravity across the Piedmont gravity high at the locations shown in Figure 1. The dashed curves and dotted curves are computed from two variations of model 1 in Figure 10, one with (dotted line) and one without (dashed line) lighter upper crustal rocks on the eastern side of the profiles. Compiled from Haworth et al. [1980] and data from the Defense Mapping Agency.

crust anywhere in the south central Appalachians and provides a relatively detailed map of crustal thickness in Virginia (Figure 4). Other studies have been made to determine a well-constrained crustal velocity model for use with a regional seismograph network [Bollinger and Sibol, 1985].

These diverse data sets can be united into a coherent interpretation using the U.S. Geological Survey vibroseis seismic reflection traverse along Interstate 64 (I-64) between Staunton, Virginia, and the Atlantic coast (Figure 4) [Harris et al., 1982a,b]. This line was recently reprocessed at Virginia Tech for the present study using an extended correlation technique to produce six extra seconds of data, resulting in 14-s records. Strong reflections in the 9-12 s range occur on the section and, in conjunction with the excellent upper crustal reflections [Harris et al., 1982a,b], provide one of the best seismic reflection images of the Appalachians available at the present time. We believe that conclusions made from this relatively detailed study can be generalized to the southern and central Appalachians.

Previous Seismic Work

James et al. [1968] used the results of previous workers and time term analysis of seismic refraction data from the East Coast Onshore-Offshore Experiment (ECOOE) to map the crustal thickness in the mid-Atlantic states (Figure 4). According to their interpretation, the crustal thickness along the I-64 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 (Figures 4 and 5). They determined a subcrustal seismic velocity of 8.15 ± 0.05 km/s in the area.

Chapman [1979] used a variety of seismic methods to determine a crustal velocity model for the central Virginia area for use in locating earthquakes inside a regional seismograph network (Figure 4). Chapman's crustal model consists of a two-layer crust with an upper layer about 15 km in thickness and a velocity of 6.09 ± 0.04 km/s; the lower layer has a velocity of 6.5 ± 0.1 km/s and a base at a depth of 36 km. The character of the transition between these layers is unknown from these experiments; it may be gradual or sharp. Chapman determined a subcrustal (P_n) velocity of 8.18 ± 0.11 km/s in agreement with the earlier value derived by James et al. [1968]. An average regional crustal (P_g) velocity of 6.34 ± 0.29 km/s was also determined using arrivals from this and other earthquakes. Later work substantiated this crustal model [Carts and Bollinger, 1981; Sibol, 1983; Bollinger and Sibol, 1985; Chapman and Bollinger, 1985] and also presented evidence that the thicker crust beneath the Appalachians continues south at least into the South Carolina region [Carts and Bollinger, 1981].

Chapman [1979] also interpreted secondary arrivals at three central Virginia stations (CVL, GHV, and FRV) from a well-constrained quarry blast to be reflections (P_mP) from the Mohorovicic discontinuity (Moho). Chapman's interpretation of P_mP using a simple two-layer model predicts a Moho that 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 5). These values are about 8 or 9 km less than those given by James et al. [1968]. (Figure 16 of Chapman [1979], which indicates a discrepancy of only 1 or 2 km between these values, is incorrectly drawn (M. Chapman, personal communication, 1986).)

Seismic Reflection Data

The I-64 reflection seismic data were acquired in 1981 by Geophysical Service Incorporated (GSI) for the U.S. Geological Survey. They used two to

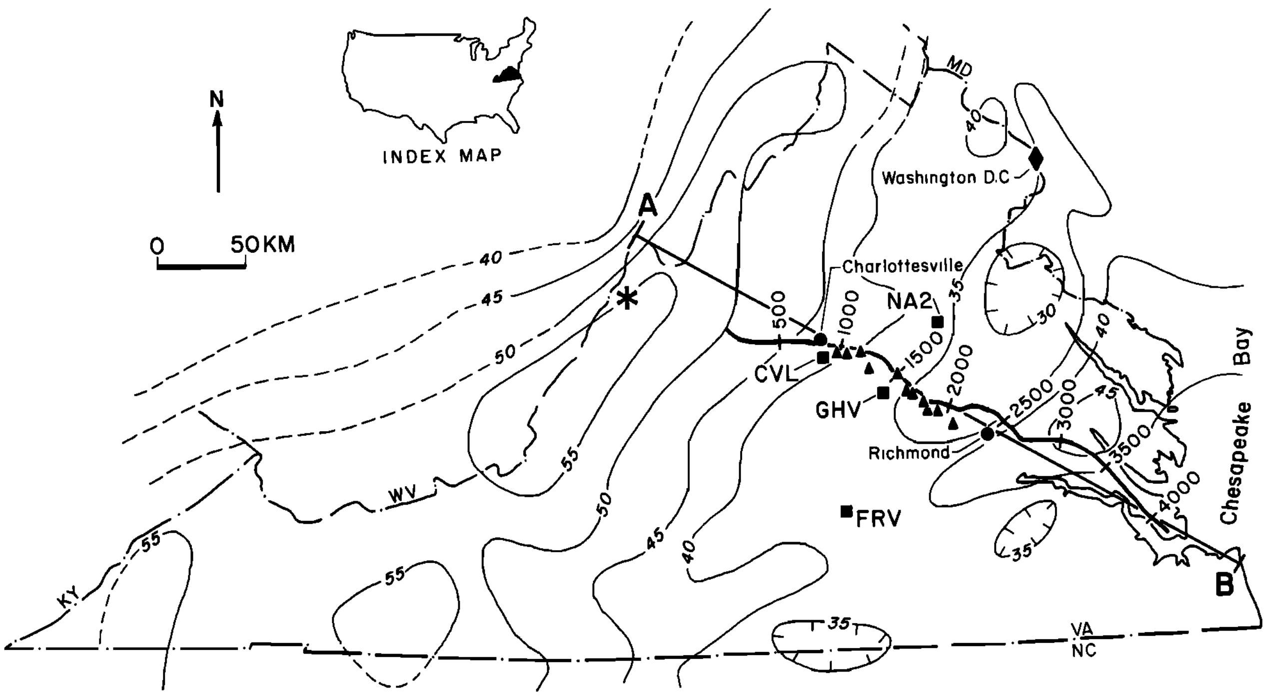

Fig. 4. Map of Virginia with crustal thickness contours (in kilometers) [James et al., 1968, Figure 7]. Also shown are the locations of the I-64 seismic reflection line (heavy line) with station numbers indicated, stations on Chapman's [1979] refraction line (triangles), the central Virginia seismograph 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 modeling.

three large vibrators as an energy source and recorded on a 48-channel, split-spread geophone array with a 67-m (220 foot) spacing to produce a 12-fold stacked section. A 10-s, 14- to 56-Hz upsweep was used with an 18-s listening time to produce 8 s of fully correlated data. These data were interpreted by Harris et al. [1982a,b], who 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.

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 [Okaya, 1986] to produce 14-s records and the application of Vibroseis Whitening (AGC before correlation) [Çoruh and Costain, 1983], a routine processing sequence was followed [Pratt, 1986]. A constant velocity time migration was done after stack on the western portion of the line but was not attempted east of station 3200 because the signal-to-noise ratio is very poor [Pratt, 1986]. The finished section is too large for meaningful reproduction, but line drawings of the unmigrated and migrated stacks are shown in Figures 5 and 6, respectively, and panels of the data are shown where appropriate.

Interpretation of Seismic Reflection Data

Most noticeable on the reprocessed section (Figures 5 and 6) is the poor reflectivity recorded east of station 2700, where Harris et al. [1982b] terminated their published profile. A strong Coastal Plain-basement reflection and a weakly imaged basin structure, interpreted to be a Mesozoic rift graben because nearby drill holes bottomed in Triassic sediments [Johnson, 1975], are visible; otherwise, the data are almost 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 5).

Such vertical boundaries seem unlikely on first inspection but cannot be ruled out. Major faults with a strike-slip component of motion are found in the Piedmont Province [Gates et al., 1986a] and could conceivably create near-vertical boundaries through the crust with differing terranes on each side. A far more likely explanation, however, is that a data acquisition problem has caused the vertical panelling of data [Pratt, 1986]. This could be due to a problem with the seismic acquisition equipment or may be a consequence of near-surface energy transmission characteristics. The exact cause remains unknown.

Upper Crust

The location of the seismic reflection profile is shown on the generalized geologic map of the east central portion of Virginia (Figure 7). The profile begins in the Valley and Ridge [Colton,

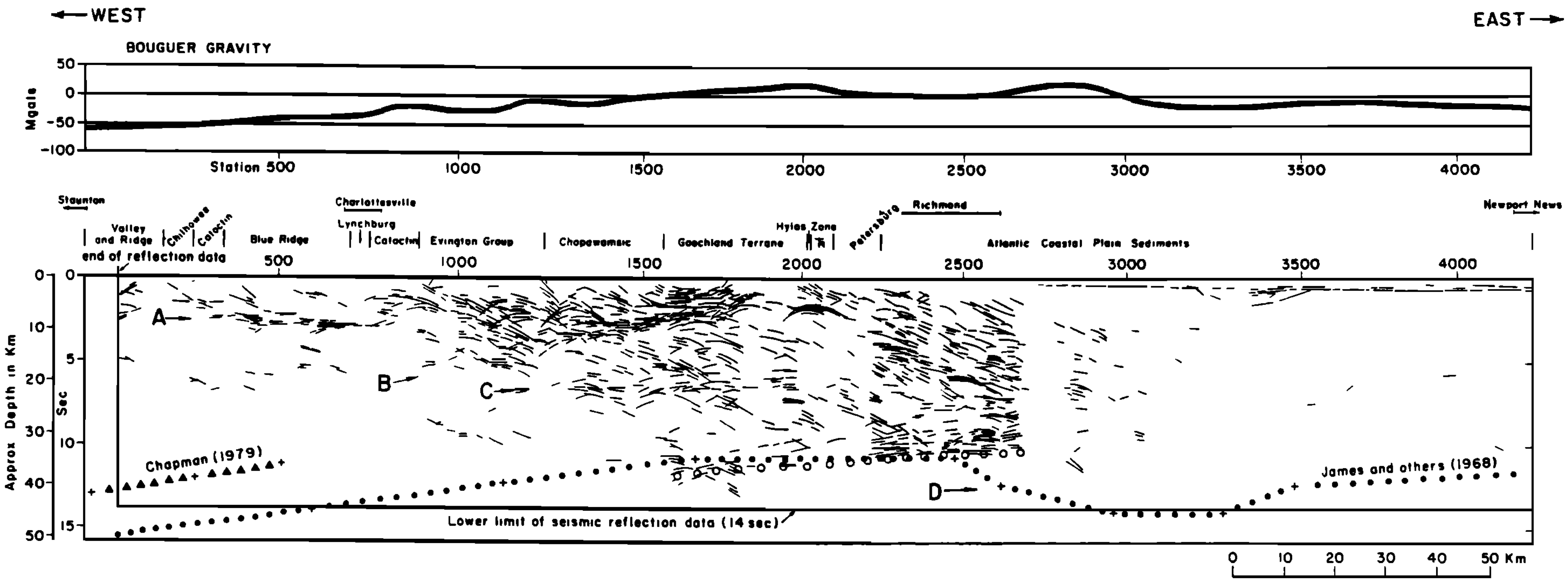

Fig. 5. Line drawing of the unmigrated I-64 seismic reflection data with interpreted base of crust (open circles) as well as crustal profiles from James et al. [1968] (solid circles) and Chapman [1979] (triangles). Crosses on the James et al. [1968] profile are locations of contours on their Figure 7; 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] along the line. A, B, and C denote the major reflection packages extending for several hundred stations to the right of, and in line with, the arrows (see text). D refers to the region where there is a discrepancy between the refraction data [James et al., 1968] and the seismic reflection data (see text). The section is plotted with approximately equal horizontal and vertical scales. The velocity (6.34 km/s) used for depth-to-time conversion is from Chapman's [1979] determination.

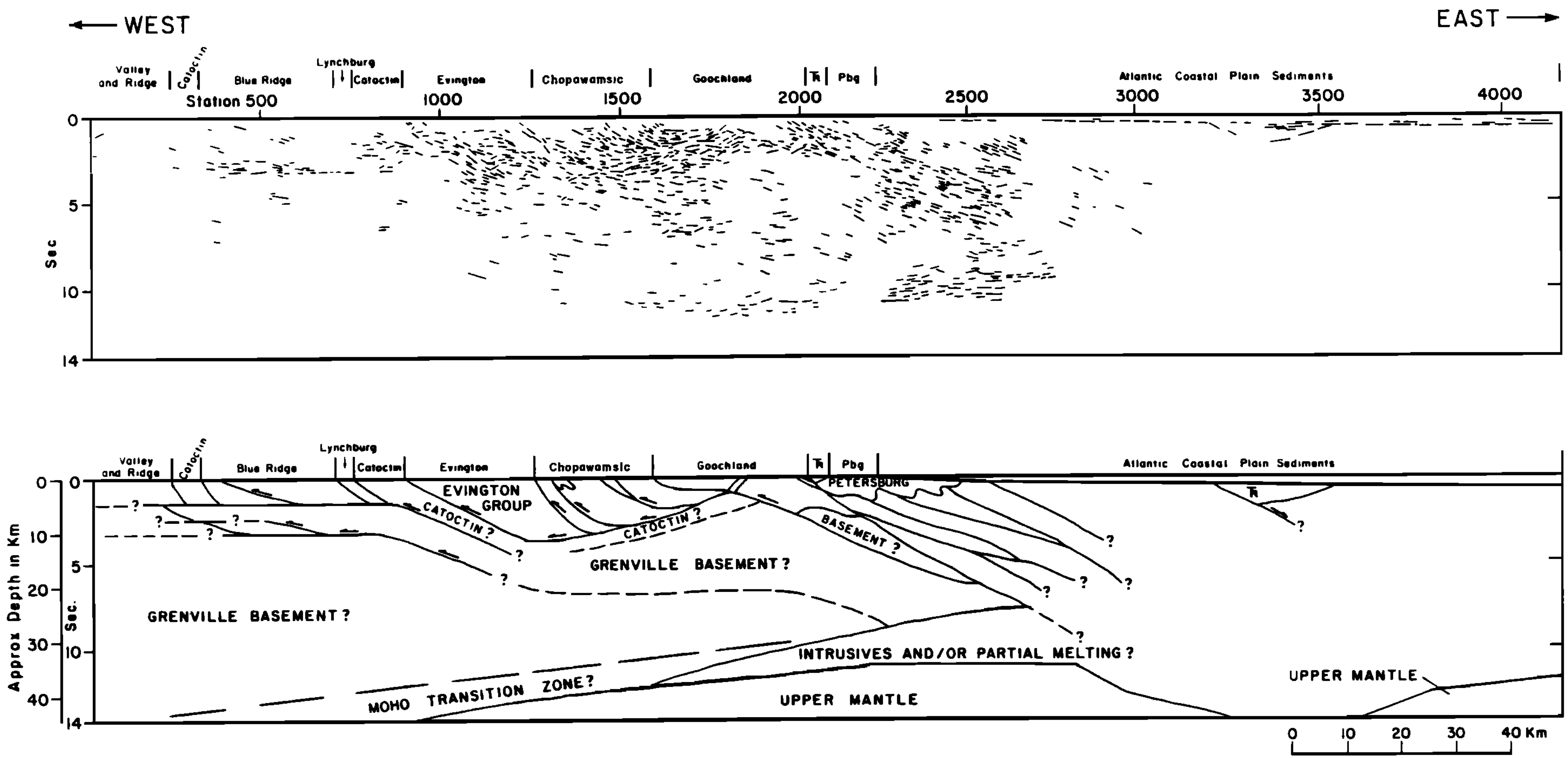

Fig. 6. Line drawing of the stacked section after post stack time migration with a possible interpretation drawn below. An alternative interpretation of the upper portion of the data is given by Harris et al. [1982b]. The eastern portion of the line drawing (beyond station 3200) is taken from the unmigrated stacked section (see text).

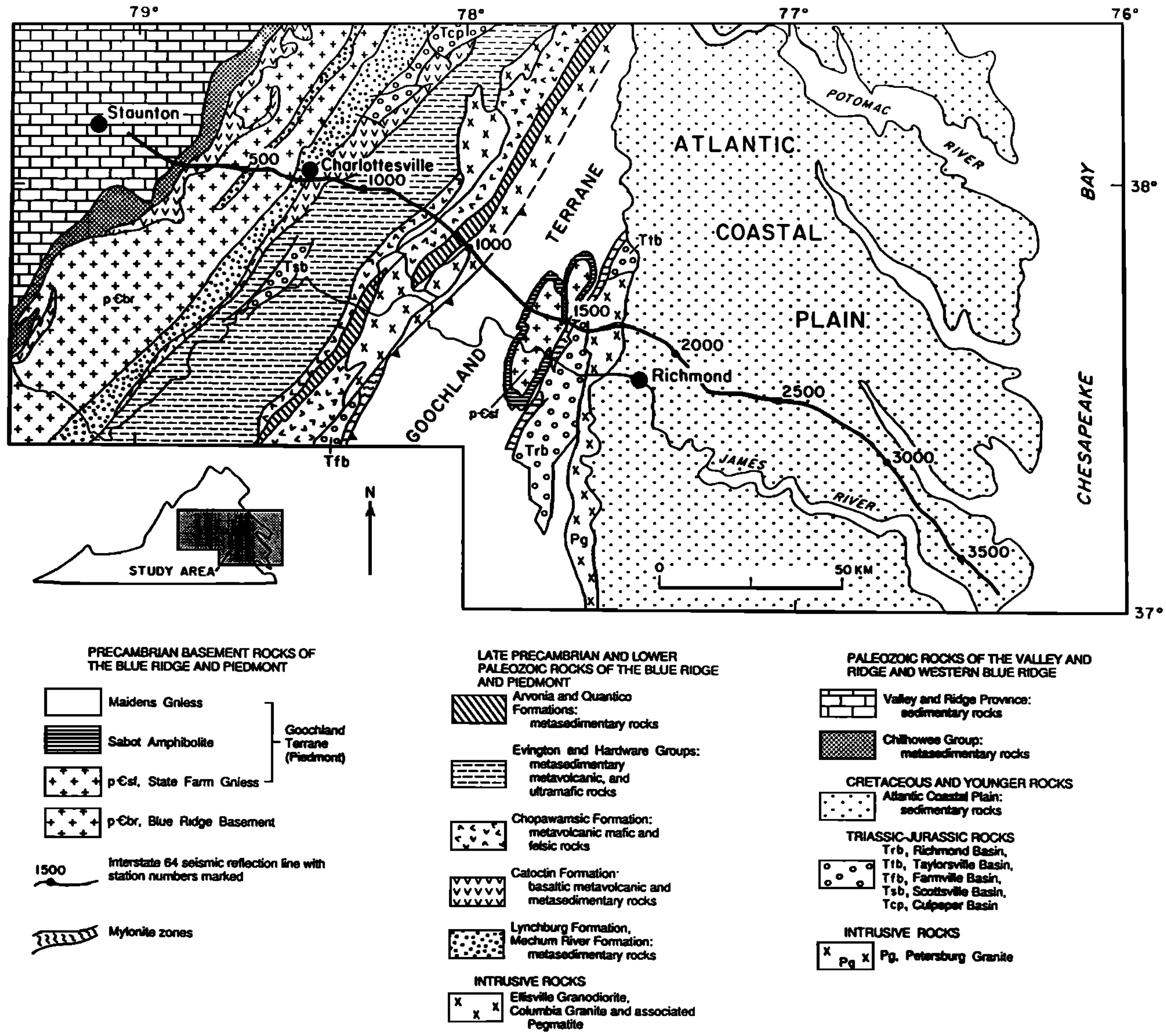

Fig. 7. Geologic map of the east-central part of Virginia compiled from Calver [1963], Bourland [1976], Poland [1976], Reilly [1980], and Farrar [1984]. The I-64 seismic reflection line is shown as a heavy black line with station numbers marked for reference.

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 s on the western end of the section (A in Figure 5) is interpreted to originate from para-autochthonous, relatively unmetamorphosed lower Paleozoic shelf strata. The Blue Ridge master decollement (decollement with greatest horizontal transport) is at a depth of about 3 km at the base of the allochthonous crystalline thrust sheet(s). This is the same structural setting consistently imaged beneath the Blue Ridge on other southern Appalachian seismic reflection data [Cook et al., 1979; Harris and Bayer, 1979; Çoruh et al., 1987]. The Appalachian Deep Core Hole (ADCOH) seismic reflection data [Çoruh et al., 1987] more clearly imaged subhorizontal reflectors from these same platform strata beneath the Blue Ridge master decollement in South Carolina. Although continuous reflections from shelf strata and thrust surfaces are not discernible on the I-64 data, a structural setting similar to that imaged on the ADCOH data, which showed a stacking of several thrust sheets [Çoruh et al., 1987], is suggested by the few reflections that are evident.

Steeply east dipping reflections beneath station 1000 that extend to depths of 12 km (4 s) define an eastern limit to these subhorizontal thrust sheets and indicate that the fault surfaces must turn down and enter basement rocks east of the Blue Ridge anticlinorium. After turning down, the thrust surfaces could be interpreted to merge with either of the large reflection packages marked B or C on Figure 5. Reconstructing the Late Precambrian position and subsequent transport of the Blue Ridge anticlinorium, which may represent the ancient continental hinge zone [Wehr and Glover, 1985] is therefore ambiguous at present. It is tempting to speculate, however, that the ramplike structure imaged here is correlative with the midcrustal dipping events imaged on the COCORP line beneath the Kings Mountain Belt [Cook et al., 1979]; both sets of dipping events reach midcrustal levels and are similarly located with respect to the Appalachian gravity gradient.

The autochthonous Grenville basement beneath the Blue Ridge has a relatively low reflectivity with only a few isolated reflections. 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 the basis of this low reflectivity, pristine Grenville crust is interpreted to underlie the entire western portion of the traverse (Figure 6).

A rift sequence composed of Lynchburg metamorphosed sandstones and Catoctin metamorphosed basalts and sandstones crops out on the eastern side of the Blue Ridge [Conley, 1978; Wehr, 1983; Wehr and Glover, 1985]. The Lynchburg-Catoctin sequence was emplaced on Blue Ridge (Grenville) basement during the Late Precambrian opening of the Iapetus ocean [Espenshade, 1970; Rankin, 1976; Wehr and Glover, 1985; Fitcher and Diecchio, 1986]. 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, 1983; Wehr and Glover, 1985]. Interpreting the Lynchburg formation on the record section is therefore difficult unless it can be correlated directly with surface exposures.

The Evington Group (including Candler and Hardware) and Chopawamsic (Carolina terrane) metavolcanic and metasedimentary rocks crop out east of the Blue Ridge anticlinorium in the western Piedmont Province [Brown, 1970; Conley, 1978; Bland and Blackburn, 1979; Evans, 1984]. The Evington Group metasedimentary and metavolcanic rocks are not well understood but may be deeper-water equivalents of the lower Paleozoic Valley and Ridge carbonate shelf strata [Brown, 1970; Evans, 1984]. Their eastern edge has been divided further into the Hardware terrane, which may include an ancient subduction zone complex (Shores Melange) possibly formed in an ocean basin [Bland and Brown, 1977; Evans, 1984]. The Chopawamsic metavolcanic rocks have an island arc composition and are interpreted to have formed along an ancient 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 [Brown, 1970; Conley, 1978]. (For simplicity the Columbia granite and Arvonia Formation are not labeled on the seismic sections.) All of these units were metamorphosed and deformed during either the Taconic, Acadian, or Alleghanian events [Glover et al., 1983].

Most of the reflections that correlate with surface exposures of Lynchburg and Catoctin rocks can be interpreted to be truncated at about 1 s on the section. The exceptions are those events on the eastern edge of the Catoctin; these dip eastward and may merge with either reflection packages B or C on Figure 5. A layer of Catoctin rocks within reflection package B would explain its high reflectivity [Brennan, 1985], and this interpretation is consistent with the normal stratigraphic position of the Catoctin beneath the Evington Group metasediments.

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 (Figures 5 and 8). Dipping events within the synform, at least some of which can be interpreted to be faults on the basis of their geometry of convergence and truncation of reflections, appear to root in an extremely reflective west dipping zone (B in Figures 5 and 8). Reflections from below this zone are markedly less abundant, implying a different, more acoustically transparent medium. The reflective zone (B) is thus interpreted as the base of the Chopawamsic-Evington sequence, possibly with a reflective sequence of Catoctin at their base. This zone may also be the location of a basal decollement, reactivated during the Alleghanian, and tilted to the west, along which the overlying rocks were transported.

Below the highly reflective zone (B) on the section is a relatively nonreflective region, interpreted as basement of Grenville age, whose upper surface defines a broad archlike geometry on the reflection profile. The top of this arch is located beneath 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 et al., 1978, 1982] and is probably of North American origin [Farrar, 1984]. It is overlain by the Sabot amphibolite and Maiden's gneiss; together the three units, plus the intrusive Montpelier meta-anorthosite, make up the Goochland terrane [Glover et al., 1982; Farrar, 1984].

From the seismic data, we interpret the Goochland terrane to be a westward thrust nappe that ruptured and overrode the Chopawamsic metavolcanic rocks (Figure 6). The relatively low reflectivity within the arch and the coincidence of its apex with a nappe of basement material (State Farm Gneiss) suggest that it is composed predominantly of high-grade basement material. The eastward dipping reflections in these interpreted Grenville basement rocks between 3 and 10 s are interpreted to originate from deformation during the Alleghanian dextral strike-slip transpression [Gates and Glover, 1986; Gates et al., 1986a; Secor et al., 1986a,b]. The basement arching(?) 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 kbar pressure [Durrant et al., 1980; Farrar, 1984, 1985]; indicative of up to 17-25 km of burial [Farrar, 1984, 1985].

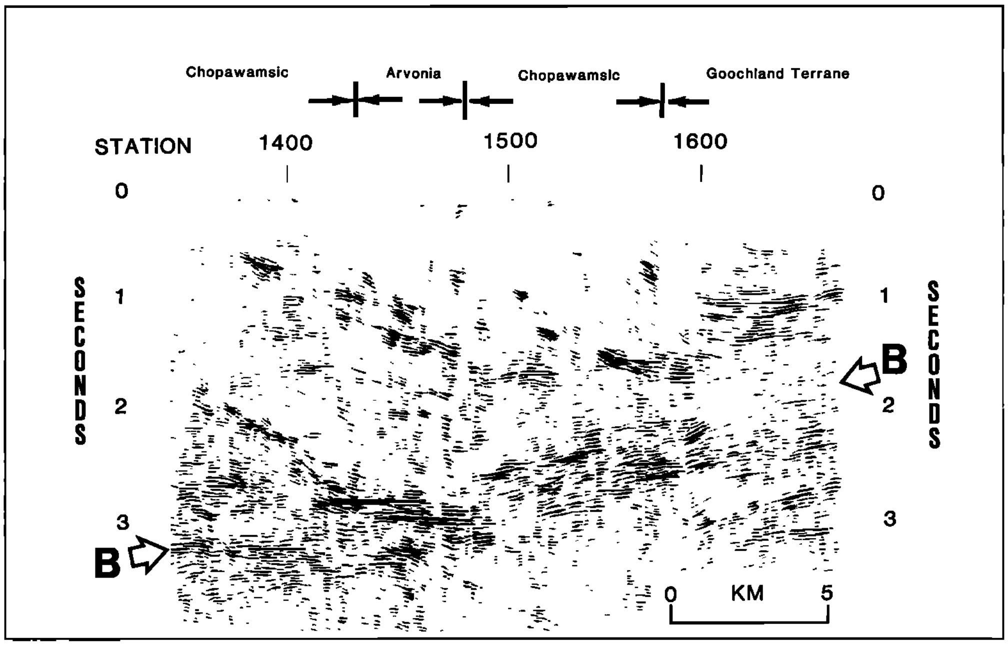

Fig. 8. Upper portion of the automatic line drawing of the seismic reflection data acquired over the Chopawamsic metavolcanic rocks and Goochland terrane. The "B"s mark the complex zone of dipping events with the same label in Figure 5. Note that some of the shallow dipping events appear to merge with zone "B," such as the event below station 1600 at 1.7 s.

The eastern edge of the Goochland terrane is bounded by the Hylas ductile deformation zone [Bobyarchick and Glover, 1979]. The Hylas is a major structural boundary that 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 juxtaposed the Petersburg granite against the Goochland terrane [Bobyarchick and Glover, 1979] and now forms the western boundary of the Richmond Mesozoic basin. The Hylas is also part of the Eastern Piedmont fault system [Hatcher et al., 1977], a system that includes the Modoc, Augusta, and other faults interpreted to have a similar deformation history.

Reflections that extend to the surface near the Hylas zone form the western boundary of a region of prominent events that dip to the east at angles of 20°-45°. Individual reflection segments are generally short, but the entire reflection package appears to penetrate the lower crust without flattening, although there is the data loss at the eastern portion of the zone.

The eastward dipping reflections obviously represent a significant crustal boundary and may correlate with those discovered on other Appalachian lines [Iverson, 1986; Nelson et al., 1986]. Nelson et al. [1985a,b] suggested that similar reflections in Georgia are the seismic signature of the Alleghanian suture. Ando et al. [1983] favored a crustal-scale ramp and buried continental margin interpretation for a similar reflection package in New England.

One interpretation for these dipping events in Virginia is that though they were modified during subsequent deformational events, they fundamentally delineate the Late Precambrian continental margin and thus the 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 its equivalent surface exposure (the eastern edge of the Chopawamsic metavolcanic rocks), but westward transport of accreted material via nappe emplacement would be expected in a collision of this nature and appropriate thrust geometry is 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 Cretaceous and younger Atlantic Coastal Plain (ACP) sediments [Brown et al., 1972] conceal most of the older rocks. Where drilled or exposed, however, the region is known 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; Johnson, 1973, 1975; Gleason, 1979; Williams, 1978; Russell et al., 1985].

The seismic data acquired over the Petersburg granite contain dipping events that reach shallow levels, indicating that the granite either does not penetrate more than several kilometers in depth or it is unusually reflective, probably due to mylonite zones like those seen at the surface within the granite. The gravity modeling discussed in a later section is consistent with a body about 5 km in thickness, indicating that at least some of the shallow reflections originate from within the granite body. The base of the granite mass is difficult to define on the

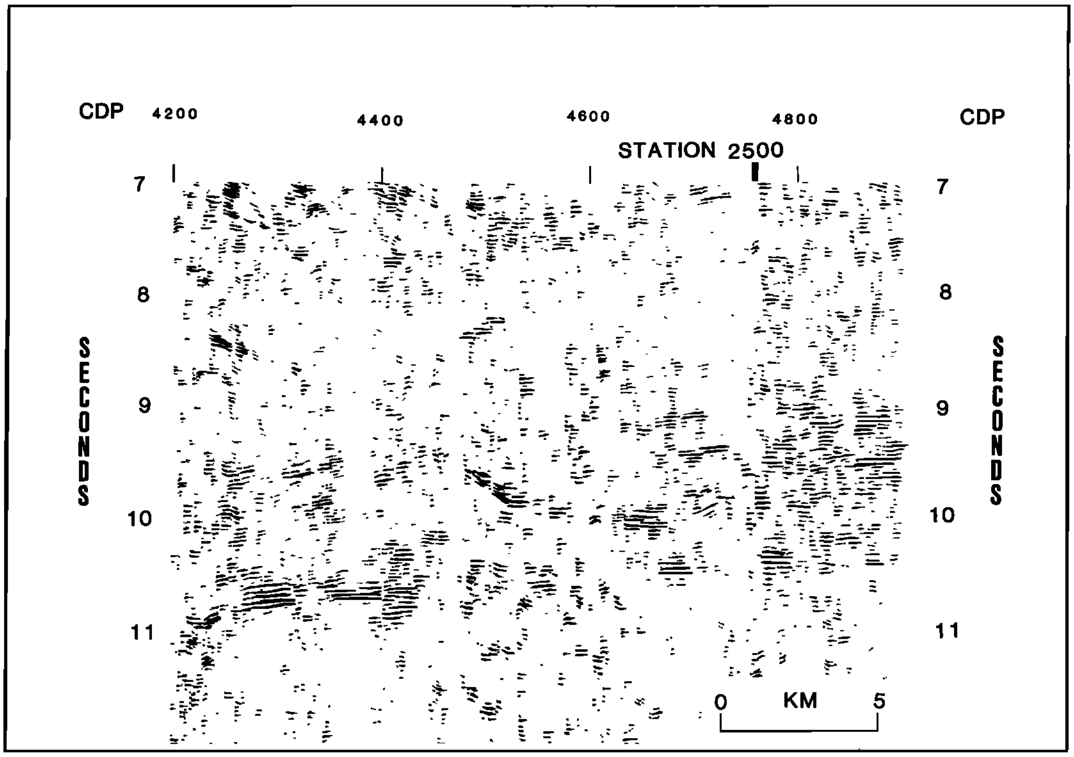

Fig. 9. Portion of the automatic line drawing of the stacked section showing the layered reflections interpreted to be from the lower crust, the base of which is interpreted as the Mohorovicic discontinuity.

seismic section. East of the granite and beneath the ACP sediments, the only feature visible on the section is a basin-shaped reflection package about 22 km in width and with a 1.2-s two-way travel time. It is coincident with drill holes that penetrated Triassic sediments [Johnson, 1975] and is therefore interpreted as a Mesozoic rift basin about 2.7 km in thickness (Figure 6).

Lower Crust

Depth to the Moho as interpreted by James et al. [1968, Figure 7] and Chapman [1979] are shown in Figure 5. Depths were converted to time and then plotted on the time section using a single-layer average crustal velocity of 6.34 km/s [Chapman, 1979]. This velocity function differs slightly from that of James et al. [1968] and Chapman's [1979] two-layer model, but the difference between these three crustal models is less than 4% and produces a maximum effect of only about 0.5 s on the time section (about 1.6 km in depth). There is also some lateral error, estimated to be about ±100 stations (±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 I-64 reflection data, the Moho of James et al. [1968] is coincident with the base of a layered set of reflections at 10-12 s on the section (Figure 9). These events are interpreted as primary reflections from the lowermost crust, with a relatively transparent mantle below. 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 refraction velocity of 8.15 km/s at its base.

The easternmost portion of the (good quality) I-64 data suggests a considerably thinner crust than the profile of James et al. [1968] (position D in Figure 5). There are several possible explanations for this apparent discrepancy [Pratt, 1986], including a lack of reflections from the lower crust in this region or a velocity anomaly which caused a misinterpretation of the refraction data. At present, however, we feel that the most likely explanation is that relief on the Moho caused a shifted interpretation of the refraction data and that the thicker crust seen in the refraction data actually lies farther to the east, where the seismic reflection data are poor. Structural distortion of this nature 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 et al. [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 about 250 surface stations (17.5 km) to the east produces a much better fit with the I-64 reflection data as well as the gravity data (see below) and is used in the final interpretation (Figure 6).

On the western portion of the profile the multichannel reflection data do not image the Moho where Chapman [1979] interpreted reflections from it, possibly due to differences between the experiments. Chapman apparently measured events that, at epicentral distances of 120-170 km and a crustal thickness of 40 km, impinged upon the Moho with incidence angles (with respect to the vertical) of about 60°. This is near the 52° critical angle for the Moho interface (computed from Chapman's velocity model) and would produce a reflection of nearly the maximum possible amplitude [e.g., Braile and Chaing, 1986]; 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, personal communication, 1986) that can record frequencies lower than the 14- to 56-Hz I-64 reflection data. This lower-frequency data can be more sensitive to a gradational boundary; in particular, a ramped velocity function can remain nearly transparent on high-frequency data yet produce strong low-frequency reflections [e.g., Braile and Chaing, 1986].

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 smooth transition (a velocity ramp) beneath the Grenville basement on the western portion of the section, whereas the crust to the east has a sharper, laminated boundary (a layered velocity function). Chapman may have recorded reflections from the top of such a gradational boundary (he picked the first arrivals) and the refracted energy may have traveled along the base (the most rapidly traveling wave was picked). Though there is uncertainty in the accuracy of both data sets, an approximately 8- to 9-km-thick velocity gradient could be interpreted in the lower crust in this region. This is nearly the same thickness as the layered reflections farther east on the I-64 data and for similar gradients observed on data from other areas [Bamford and Prodehl, 1977; Braile and Chaing, 1986].

Gravity Modeling

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 two-dimensional gravity modeling (line A-B in Figure 4). The simple Bouguer gravity field along this line was sampled at 10-km intervals from the maps of Johnson [1971, 1972, 1973, 1975], and a geologic cross section was constructed on the basis of the interpretation of the seismic reflection data. The gravitational field associated with the interpreted model was then computed using the method of Talwani et al. [1959] and aligned with the observed field by subtracting a constant that equalized the values at the 110-km point of the transect (near the midpoint of the gravity gradient). An approximate crustal profile outside the limits of profile A-B on Figure 4 was found necessary to reduce edge effects during modeling and, at points beyond the control derived from James et al. [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 ACP sediments offshore [Grow et al., 1979].

The first model (Figure 10, model 1) was designed to test the effects of crustal thickness and consisted of a uniform crust except for mass units representing the upper mantle, the ACP sediments, the Mesozoic basin imaged on the reflection data, and a light mass at the eastern edge of the model (the Portsmouth, and perhaps other, Alleghanian granitoids). Units representing the ACP and Mesozoic sediments were included because their geometry is well constrained by the seismic data and a reasonable density contrast could be assumed. The light mass was found to be necessary to match the model field with the observed field; its shape has no geologic implications but was simply determined by altering the outline until the two gravity fields were reasonably matched. 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 that probably results from mafic rocks in the basement beneath the ACP sediments [Johnson, 1973, 1975; Gleason, 1979].

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 that confirms conclusions from simpler models [e.g., Cook, 1984a]. The eastern portion of the transect, on the other hand, has significantly lower gravity values than those produced only by the crustal profile derived from the James et al.'s [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 crustal material 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. Drill holes that have penetrated the ACP sediments encountered a basement of low-density granitoids and Mesozoic sediments in several widely scattered areas [Johnson, 1973, 1975; Gleason, 1979; Russell et al., 1985], providing further support for this hypothesis. The progressive thickening of the lighter material toward 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 Mg/m^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, produced an unacceptably high gravity gradient between the Valley and Ridge and Coastal Plain Provinces. Compensating for this effect would in turn require the addition of large amounts of light material in the eastern side of the crustal model or a significant alteration in the Moho profile. These latter alternatives cannot be discarded, but the model with the 0.25 Mg/m^3 crust-mantle

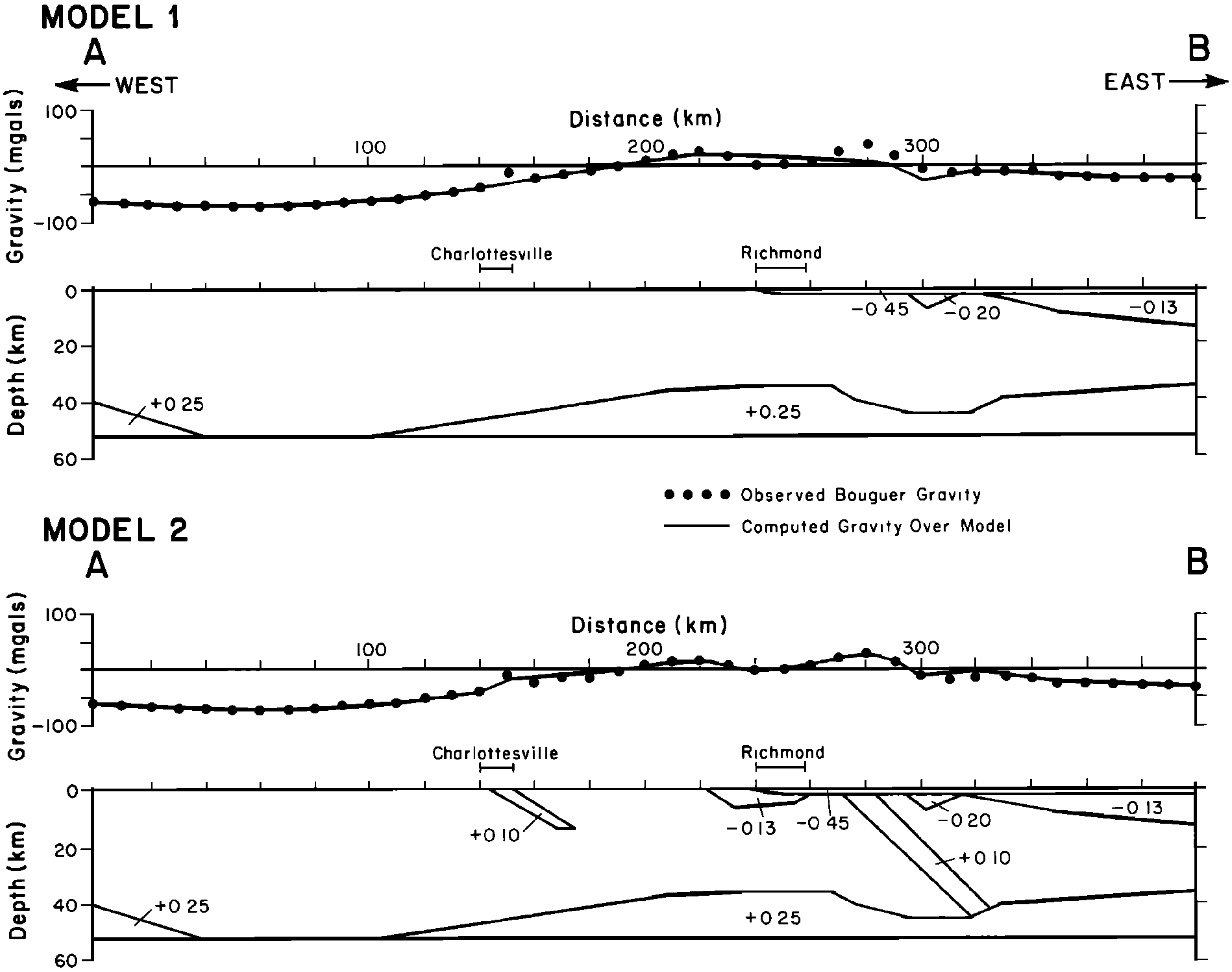

Fig. 10. Model 1 is a gravity model of the crust in central Virginia. Moho profile was taken from James et al. [1968] after shifting it eastward about 17.5 km (see text). Anomalous bodies represent the Atlantic Coastal Plain sediments (thickness exaggerated in the figure for clarity), a buried Mesozoic basin, and a large granitic body (Portsmouth Granite). Model 2 is the same as model 1 except for additional bodies representing the Petersburg granite, Catoctin metabasaltic rocks, and mafic rocks beneath the Atlantic Coastal Plain sediments. See text for explanation of the shapes of these bodies.

density contrast is more pleasing because of its simplicity and consistency with the refraction Moho.

This relatively small contrast may be an indication of the presence of a relatively light upper mantle or of a relatively dense lower crust. Laboratory measurements and other gravity modeling place the density of the upper crust at about 2.77 Mg/m^3 [Keller et al., 1985, and references therein]. Therefore, if the crust has a nearly uniform density throughout (about 2.8 Mg/m^3), the upper mantle must have a density of only 3.0-3.1 Mg/m^3. For the upper mantle with a velocity of 8.15 km/s [Chapman, 1979; James et al., 1968], however, results from laboratory measurements [Birch, 1960, 1961] predict a density of 3.25 Mg/m^3 or greater. Reversing this argument, perhaps the 0.25 Mg/m^3 lower crust-upper mantle density contrast is indicating that the lower crust has a minimum density of about 3.0 Mg/m^3, or about 0.23 Mg/m^3 greater than the upper crust. Chapman's [1979] 6.5 km/s midcrustal velocity lends strong support to this hypothesis and is otherwise unexplained. Likewise, the seismic evidence for a velocity gradient and layering in the lower crust suggests that a significant portion of this denser material may lie in the lowermost 8-10 km of the crust. The suggestion by Hatcher and Zietz [1980] that thick mafic crust underlies large areas of the southeastern United States would seem to be consistent with these results.

A second model (Figure 10, model 2) was easily constructed whose computed gravity field matched the observed gravity profile to within 10 mGal. This model was identical to the first except for additional masses representing the Catoctin metavolcanic rocks on the east limb of the Blue Ridge, the Petersburg granite, and the mafic rocks beneath the ACP sediments. The shapes of these masses are constrained by surface exposures or drill holes [Johnson, 1973, 1975; Gleason, 1979] and to some extent by the seismic reflection data. The outline of the mafic rocks shown on the model is made under the assumption that they form a thin body which penetrates the entire crust at about the same angle as the east dipping events imaged to the west of them (a variety of other shapes produced equally tractable results). A strong, narrow magnetic anomaly [Zietz et al., 1977] is coincident with the top of this mafic body and is consistent with a relatively thin body extending for about 175 km along strike. The outlines for these bodies are thus relatively unconstrained and largely hypothetical; much more detailed three-dimensional studies will be needed to properly determine their shapes. The significance of this model is simply that it gives an estimate of the minimum amount of anomalous material present in the upper crust. As Keller et al. [1985, 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 some of the metavolcanic rocks. The bulk densities of the other units are apparently similar and therefore cannot be used effectively to constrain the interpretation by gravity methods.

Discussion

Lower Crust

Above and along strike with the Moho reflections on the I-64 data, and summarized earlier in this paper, the Piedmont rocks have undergone extensive Alleghanian deformation and metamorphism as well as Mesozoic rifting. In particular, the western limit of the lower crustal reflections approximately coincides with the westernmost of the exposed rift basins that are indicative of Mesozoic extension. The Moho reflections on the I-64 seismic reflection data thus follow the tentative distribution pattern found in other areas; namely, strong reflections beneath the most recently active regions of the crust, particularly areas of extension [Bamford and Prodehl, 1977; Meissner and Wever, 1986; Smithson and Johnson, 1986]. 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; Bamford and Prodehl, 1977; Braile and Chaing, 1986]. As others have noted, these observations suggest that the lower crustal layering may be a relatively short-lived phenomenon that fades to a more transitional boundary, a second-order discontinuity on reflection data, beneath the older crust. Possible models are summarized by Klemperer et al. [1986] and Matthews and Cheadle [1986].

Several observations can be made about the deep crustal reflections on the I-64 section that constrain some aspects of published models. First, comparison of the refraction and reflection data indicates that mantle velocities (8.15 km/s) 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/s) or gradational between lower crustal and upper mantle. If our interpretation of Chapman's [1979] data beneath the Blue Ridge is correct, however, the lower crust in that region is contaminated by higher-velocity (mantle?) material. These observations suggest that a likely model would accommodate either contamination of the lower crust by upper mantle material or depletion (by melting?) of the lower-velocity material from the base of the crust. An underplating model [Furlong and Fountain, 1986] would need an underplating material that does not have a strictly mantle composition. Likewise, a shearing model for the lower crustal layering must allow for a higher-velocity material in the lowermost crust. Emplacement of mantle material into the lower crust could also produce a lower crust-mantle density contrast, thus satisfying the gravity modeling if the distribution of material is appropriate.

The layering has an indistinct upper surface but a comparatively sharp, well-defined lower boundary on the section. This reflector geometry requires an explanation that tends to concentrate the reflectors in the lowermost portion of the crust. If fluids were responsible for the reflectivity, for instance, they 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.

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 (or latest?) shearing occurring at the lower boundary of the viscous zone where the brightest and most extensive reflections are. The upper mantle material either did not undergo strain to produce the layering, or it has not been permanently recorded. Perhaps this places an upper time limit of about 200 Ma (since the Mesozoic) on the persistence of upper mantle reflectors. The lateral decay in reflection amplitudes would also imply 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 is that the layering and crustal thinning are interrelated, the layering a by-product, or the cause, of thinning. Partial melting of a rising, and thus depressured, lower crust is therefore an appealing model.

The Alleghanian plutons and Mesozoic dikes [Speer et al., 1979; Fullagar and Butler, 1979; Ragland et al., 1983] indicate that lower crustal-upper mantle melting was occurring at that time and lend support to a partial melting model for the layering. The subhorizontal attitude of the reflections could then be attributed to pressure effects determined by the gravity field [Meissner and Wever, 1986]. Finally, if the layering seen on the reflection profile has persisted since the Mesozoic and will eventually fade to a second-order order discontinuity, it would suggest either a long equilibration time for the layering or, more likely, the necessity of another tectonic event to decrease the reflectivity.

Crustal Thickness Variations

The gross geometry interpreted from the geophysical data in Virginia (Figures 5 and 6) consists of a crust that thins eastward until the edge of the ACP sediments and then rethickens, at least locally, to the east. The results from gravity modeling demonstrate that in Virginia the Piedmont gravity high is primarily due to this region of 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.

In Figure 1 the Piedmont gravity high is shown as a dot pattern, and the nine straight lines show the locations of the gravity profiles across the high shown in Figure 3. 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 that crop out in the Piedmont [Williams, 1978] and the variations in crustal thicknesses interpreted from the refraction data [James et al., 1968] are examples of such changes that obviously effect the gravity field by producing significant local anomalies. Note, for example, the localized thin crust beneath station FRV on Figure 4 that corresponds to a small eastward extension of the Piedmont gravity high ("A" on Figure 1). This same perturbation is somewhat offset by the presence of the Petersburg granite that probably causes the elongate negative anomaly encroaching on the gravity high adjacent to the exposed Richmond Mesozoic basin. Likewise, the northernmost profiles in Figure 3 show the effects of a gravity high near 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 3 are the computed gravity values from model 1 in Figure 10 with (dotted line) and without (dashed line) the light material on the eastern edge. The model profile was aligned with each of the observed gravity profiles to produce a good match between the two. The comparison shows that even a simple model including only the Moho profile and the ACP 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 ACP 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 to be preferred.

Regional seismic reflection profiles are indicated as heavy lines on Figure 1 with the adjacent dots denoting the major zones of crustal penetrating eastward dipping reflections seen on those sections. The spatial relationship between the gravity high and the dipping events along the entire length of the Appalachians is striking and, and noted by other authors [Ando et al., 1984; Hutchinson et al., 1986] supports a relationship between the two features. The geometry suggested here is that, as in Virginia, there is significant rethickening, at least locally, and possibly lighter crust east of the Piedmont gravity high and the adjacent zones of dipping events. Determining the relative effects of rethickening versus lighter crustal material is contingent on mapping the Moho depths beneath the Atlantic Coastal Plain Province.

Control on the Moho depths beneath the Piedmont and Coastal Plain from refraction data is sparse outside of Virginia, and the available data are inconclusive [see Hutchinson et al., 1983, Figure 2]. Equally frustrating is the lack of Moho reflections on the multichannel data [Cook et al., 1979, 1981; Nelson et al., 1985a], but slightly longer travel times to the Moho reflectors may be indicated beneath the highly reflective zones of dipping events on both southern Appalachian COCORP profiles (Figure 2) [Cook et al., 1981; Nelson et al., 1985a]. Cook et al. [1981] show that the apparently thicker crust on the East Georgia lines coincides with a gravity low [Cook et al., 1981, Figure 3] (see also discussion by Iverson and Smithson [1983]), creating a geometry similar to that in Virginia.

Coincidence of Geologic and Geophysical Features

The exposed Carolina Slate Belt and related metavolcanic rocks, possibly a Cambrian island arc terrane accreted to North America during the Taconic event [Bland and Blackburn, 1979; Pavlides, 1981], outcrop in a belt coincident with the gravity high. These rocks may contribute to the gravity high, but in Virginia, at least, their contribution is apparently small. The easternmost exposures of Grenville basement rocks in the Piedmont, the Goochland-Raleigh belts in Virginia and North Carolina [Farrar, 1984, 1985], and the Pine Mountain Belt in Georgia [Schamel et al., 1980; Nelson et al., 1985a] also lie close to the gravity high. Fault systems in the south central Appalachians likewise exhibit continuity along most of the Piedmont gravity high. 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 in relatively thin thrust sheets [Cook et al., 1979, 1981; Secor et al., 1986b]. The eastern Piedmont fault system [Hatcher et al., 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 et al., 1981; Nelson et al., 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 (some of which are part of the Eastern Piedmont fault system) and their associated sedimentary basins (Figure 1). The Mesozoic basins persist beneath the Atlantic Coastal Plain and offshore regions, but their western limit is strikingly coincident with the gravity high.

These features, all seemingly interrelated, can be interpreted as marking a fundamental zone of crustal weakness that, because of the range of ages of these features (Ordovician to Triassic), may have persisted since the Precambrian. As mentioned earlier, an interpretation whereby the dipping events on the seismic section and the apparently related gravity high mark the position of the late Precambrian-early Paleozoic continental margin, and thus the eastern limit of

North American Grenville basement, is favored here. The continental edge, 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. The variation in distance between the Blue-Green-Long axis, which may be the exhumed 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. Perhaps during initial Triassic rifting, a region of thinner crust formed which now contributes substantially to the gravity high. An intriguing possibility consistent with the geometry along the I-64 line is that rifting initially began in the Piedmont region in a simple-shear fashion [Wernicke, 1985], with the shear planes occurring within the already existing crustal weakness marked by the zone of east dipping events. Rifting then ceased in the Piedmont when the Atlantic ocean basin opened farther to the east.

Conclusions

Allochthonous crystalline rocks beneath the Blue Ridge in Virginia are relatively thin (~ 3 km) and possibly underlain by other thrust sheets. Autochthonous Grenville basement is at a depth of about 9 km (3 s) beneath the Blue Ridge. The associated thrust structures turn down on the east side of the Blue Ridge anticlinorium where east dipping reflections penetrate to midcrustal levels. 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 base of this synform is marked by a highly reflective band that is interpreted to originate from Catoctin metavolcanic rocks and decollements.

A major conclusion of this study is that the Goochland terrane is interpreted to be a nappe structure that overrode the Chopawamsic rocks from the east. East of the Goochland terrane, a zone of prominent east dipping reflections appears to penetrate to lower crustal depths. This broad zone of east dipping events may be coincident with the Precambrian continental edge.

Strong reflections in the 9- to 12-s range on the section are interpreted as lower crustal layering about 5-10 km in thickness. They have a well-defined base that almost exactly coincides with the Mohorovicic discontinuity derived from earlier refraction work. 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 (~ 35 km) beneath the Goochland terrane, and then rethickens to the east. Results of gravity modeling suggest that this crustal 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. Major zones of anomalous crustal rocks are not necessary to satisfy the gravity modeling, except possibly beneath the Atlantic Coastal Plain where significant amounts of lighter crustal material may be present. The thinner crust on the Moho profile is bounded on the east by the pronounced zone of east dipping reflections on the seismic section. 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. These features are interpreted as delineating a zone of crustal weakness that was the continental edge in Precambrian times and has persisted as a zone of tectonic activity during subsequent events.

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