Physiographic Divisions and Differential Uplift in the Piedmont and Blue Ridge

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By JOHN T. HACK

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A study of the topography and other surface features of the Piedmont and Blue Ridge, as related to late tectonic history

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PHYSIOGRAPHIC DIVISIONS AND DIFFERENTIAL UPLIFT IN THE PIEDMONT AND BLUE RIDGE

By JOHN T. HACK

ABSTRACT

The Piedmont and Blue Ridge are dynamic landscapes that have undergone substantial change since the orogenies that ended in late Paleozoic or, as some believe, early Mesozoic time. The southern Blue Ridge region south of Roanoke, Va., lies on the crest of a topographic uplift that corresponds to the eastern continental drainage divide. To the north, this uplift and divide cross the Appalachian Valley and form the crest of the Appalachian Plateaus as far north as central Pennsylvania. The northern Blue Ridge Mountains as well as parts of the Piedmont are on the eastern part of the uplift area. The southeastern margin of the uplift corresponds to a line within the Piedmont physiographic province that extends northeastward from the Tallapoosa River at the Fall Zone and crosses the Rappahannock River at the Fall Zone. The differential elevation on either side of this line is sharp in some places, as, for example, northeast of Atlanta, Ga. In other places, the difference in elevation is difficult to detect, and, in effect, the line becomes a broad monoclinal slope.

The region as a whole can be divided into at least six broad subregions that have somewhat different histories in late geologic time.

The Piedmont Lowlands subprovince, southeast of the uplifted area, is dominated by a monotonous topography of low rounded ridges and ravines largely underlain by saprolite on crystalline rocks. Isolated ranges of hills of greater relief are scattered across the region; those investigated are directly related to the presence of erosionally resistant rocks. Stream patterns as well as broad topographic forms indicate that although the southern part of the Piedmont Lowlands was probably once covered by younger sediments, this area has been exposed to erosion for a long time. In North Carolina, the inner part of the Piedmont Lowlands has strongly trellised stream patterns, which suggest that subaerial erosion was active for an even longer time period, perhaps since the latest orogeny. North of the Cape Fear River, the outer part of the Piedmont Lowlands was covered by either fluvial or marine sediments or both, probably during Miocene time. Tectonic activity has affected the Piedmont Lowlands in late geologic time. The Fall Zone that forms the southeast border is, at least in places, controlled by faults active in Tertiary time. Late faults have also been found in the Pine Mountain area of Georgia. Minor differences in relief affecting large regions within the Piedmont Lowlands may be related to different rates of uplift in addition to rock resistance, either past or present.

The Piedmont northeast of the Potomac River (Northeastern Highlands) rises to more than 300 m in altitude. The major streams have convex profiles that steepen as they near the Coastal Plain. Unusually narrow valleys and broad upland surfaces indicate an increased rate of erosion and show that the relief is now or recently has been increasing because of uplift or tilting.

West of the southern end of the Piedmont Lowlands is an area herein called the Southwestern Highlands that in some respects is similar. The area is crossed by two large streams that have convex profiles. The highest mountain ranges in the area rise to altitudes greater than 600 m.

Northwest of the Piedmont Lowlands, the topography and relief are higher, and in some places, the rise is gradual, forming a Foothill zone between the Piedmont Lowlands and the high Blue Ridge. This zone is morphologically more complex than the Piedmont Lowlands. North of the Roanoke River, the foothills are commonly chains of isolated hills and ridges generally underlain by resistant rocks. The hills increase in height near the Blue Ridge, an indication that they owe their height to tectonism of late geologic age. South of the Yadkin River, the hills are believed to be residual, the remnants of a larger highland that has been only partially reduced to the lower relief of the general Piedmont surface.

The Blue Ridge physiographic province has two subdivisions. The Northern Blue Ridge Mountains are underlain by resistant rocks and in effect form the easternmost range of the Valley and Ridge province. Three different rock types form the crest in various places; the differences in their resistance as well as in the width of outcrop belts determine the altitude of the range. North of Chester Gap in northern Virginia, the lower altitude of the range may be related in part to differences in uplift history over a broad region.

The Southern Blue Ridge province is a complex area. Part of it, the New River basin, is essentially a plateau that owes its relative height to resistant rocks crossed by the New River downstream from the Blue Ridge itself. The southern part is underlain by massive and resistant gneiss and metasedimentary rock and has relief exceeding 1,100 m. Rock control has a strong influence on forms locally. Deep trenchlike valleys follow weak-rock outcrop areas like the carbonate rock of the Murphy marble belt as well as fracture zones or shear zones along faults. The Asheville basin may be partly controlled by the outcrop of weak rocks but also by intersecting shear zones.

The Southern Blue Ridge province is bounded on its southeast margin by a bold escarpment that corresponds to the major drainage divide. As shown by a major piracy by the Dan River, at least the northeastern part of this escarpment has retreated toward the northwest. At the Grandfather Mountain window and south of it, the escarpment may be stabilized along a zone of resistant rocks. The origin of the escarpment is probably a monoclinal flexure or a movement along a series of undetected faults as much as 30 km or more southeast of the present position of the escarpment.

The time frame in which the uplift in the region took place can be estimated from data on erosion rates in areas of different relief, as suggested by Ahnert's concept relating rates of relief reduction, rates of uplift, and rates of erosion. His analysis shows that if uplift is not continuous, any area such as the Blue Ridge would be reduced to 10 percent of its initial relief in a period of only about 20 million years. This means either that uplift of the areas northwest of the Piedmont Lowlands has taken place since Miocene time, or, more likely, that it has gone on continuously at different rates, the highest rate being in Miocene time.

INTRODUCTION

Geologists have long thought that the late tectonic history of the Appalachian Mountains involved vertical uplift. It is difficult to escape this conclusion because the mountain system declines in both altitude and relief near its southwestern terminus to become buried beneath a cover of Cretaceous and Tertiary sediments. Many studies of the forms of individual areas within the mountain chain have been based on the assumption that the hilltops were relict features inherited from surfaces having much lower relief, called "peneplains." The degree of dissection of these peneplains was considered a measure of the uplift. The times and average rates of successive periods of uplift could be deciphered if the age of the peneplains could be established. This age was rarely possible to determine by stratigraphic methods in the Appalachians because of the scarcity of fossiliferous postorogenic sediments in the erosional landscape.

The existence of relict forms inherited from older landscapes of the kind postulated by W. M. Davis and his followers is questionable. The ubiquitous pattern of interconnecting valleys separated by branching ridge systems throughout the region is an indication that this topographic form prevails and probably always has. Regional differences in topographic relief are significant. Many can be explained by differences in erosional resistance of the bedrock. Some can be explained only by differences in rates or history of uplift (Hack, 1960). This study attempts to identify different erosional systems within the crystalline rock area of the unglaciated Appalachians and their tectonic relation.

My earlier work in the Appalachian region has convinced me that the present landscape is dynamic and that most features can be explained by a balance of forces involving uplift, erosion, and rock resistance. The landscape chosen for study at that time was mostly in the Ridge and Valley province, where extreme differences in rock resistance are obvious (for example, limestone or shale as compared with quartzite). More subtle tectonic effects were not detected, and this aspect of the problem was not addressed except to acknowledge that it existed.

Ahnert (1970) has critically evaluated the relation of erosion and uplift in areas of differing rainfall, average slope, and relief. He has shown that topographic relief in a variety of areas having complex geology has a close linear relation to rate of denudation as measured by present rates of debris transport in streams. Differences in climate between the areas apparently have little effect on erosion rates. Average slope is related to denudation rates but not as closely as relief. Ahnert pointed out that although erosion rates and relief are probably not evenly balanced so as to achieve a steady state, a tendency toward such a balance exists. His analysis also indicates that, in part because of isostatic compensation, it would require periods on the order of 18 million to 20 million years to reduce a landscape to 10 percent of its initial relief if no tectonism were involved. If his analysis is valid, high relief should be correlated with high uplift rates at present or at some time in the not-too-distant past.

In the discussion of the Piedmont and Blue Ridge that follows, I assume that differences in relief from place to place may be the result of differences in rock resistance, which I refer to as rock control (Yatsu, 1966). On the other hand, differences in relief may be due to different rates or different histories of tectonic uplift. However, it is assumed that all high-relief features must be postorogenic, that is, Mesozoic or Cenozoic. Furthermore, areas having differences in relief must have had different tectonic histories if it can be shown that the differences were not caused by rock control.

Evidence other than relief can be used to analyze tectonic effects and histories. Stream patterns may reflect stages during downwasting. Stream-profile analysis proves to be a powerful tool with which to identify possible tectonic movement, because the steepness of the channel slope for a given discharge is a measure of stream power that may, like relief, be related to either the resistance of the rock through which the stream flows or to local uplift or tilting. Faults, folds, and other local features are, of course, also manifestations of tectonism.

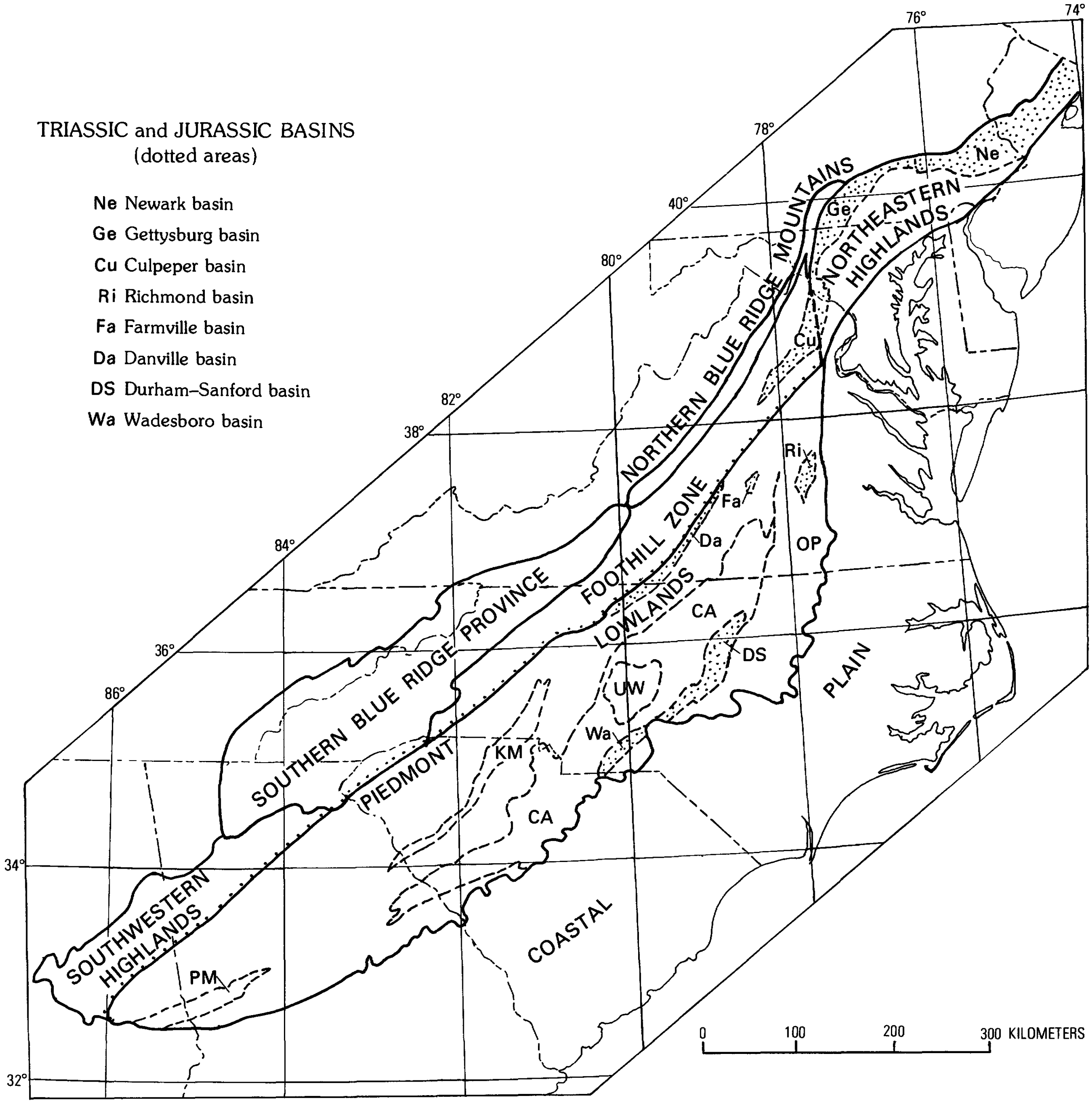
The writer is indebted to many colleagues in the U.S. Geological Survey for advice on the geology of many areas, and especially to Milan Pavich, Helaine Markewich, and Douglas Rankin for their reviews of the manuscript.

PLAN OF REPORT

The Piedmont and Blue Ridge provinces, together, can be divided into subprovinces or areas having distinctly different topographic forms. On the basis of the geology, topography, and stream systems, six major subprovinces (fig. 1) have been identified.

The discussion is illustrated by four maps of the region. Figure 2 is a highly generalized lithologic map adapted from the "Tectonic Lithofacies Map of the Appalachian Orogen," by Williams (1978). This map was chosen because it is probably as close to a lithologic map as can be constructed at so small a scale.

Figure 3 is an averaged topographic map that shows the topographic setting of the Piedmont and Blue Ridge provinces and their relation to the adjoining topography of the folded Appalachians and the Appalachian Plateaus. Contours on the basement rocks beneath the Coastal Plain are also shown. The subaerial altitudes are adapted from an averaged contour map having 50-m contours that show mean altitudes. The map shows the mean altitude within 3-minute rectangles (approximate-



EXPLANATION

TALLAPOOSA-RAPPAHANNOCK LINE

SUBDIVISIONS OF PIEDMONT LOWLANDS

CA Carolina slate belt	**OP** Outer Piedmont of North Carolina and Virginia
UW Uwharrie Mountains	**KM** Kings Mountain belt
	PM Pine Mountain belt

FIGURE 1.—Map of Southeastern United States showing the subprovinces within the Piedmont and Blue Ridge and the Tallapoosa-Rappahannock Line.

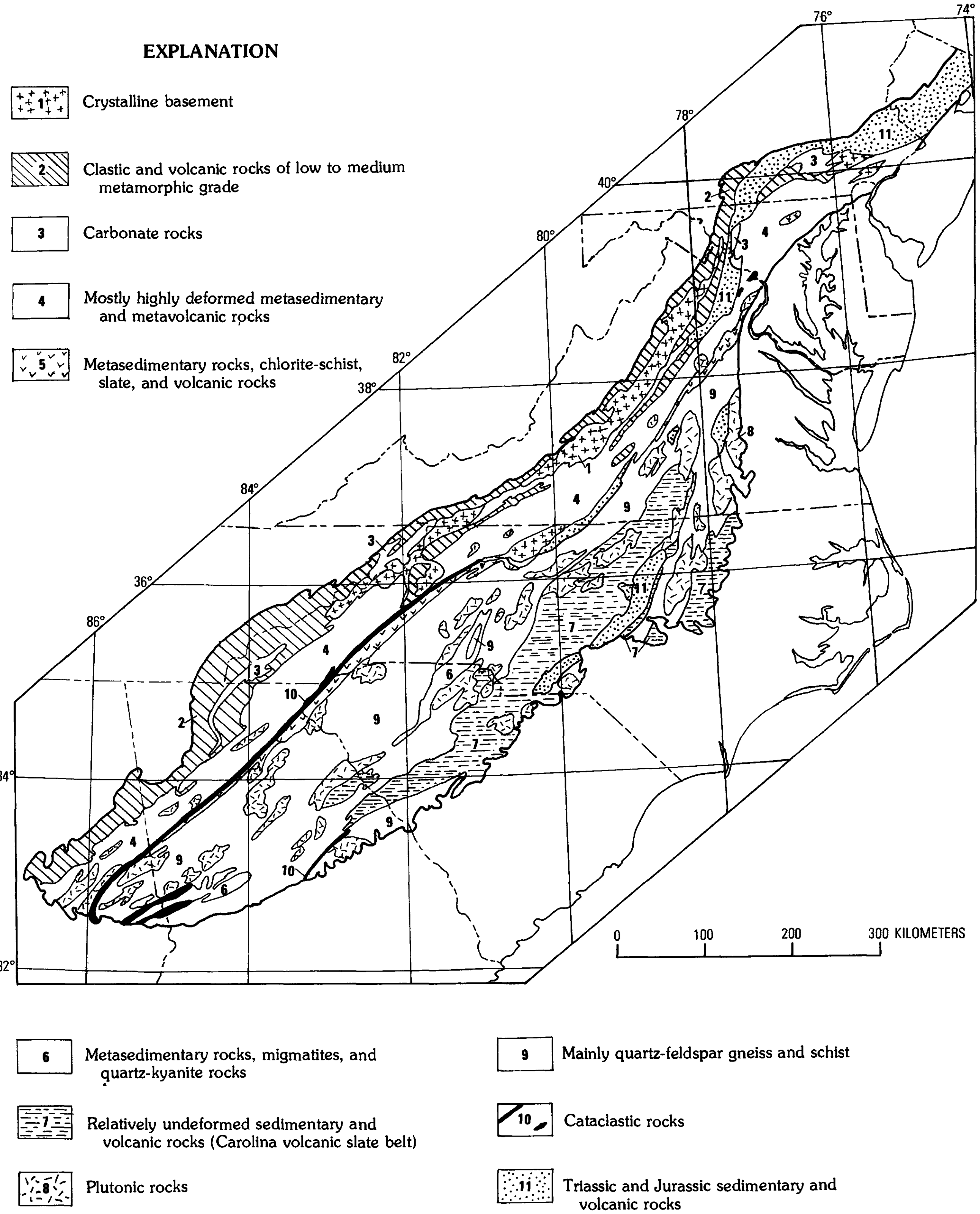


FIGURE 2. – Generalized lithologic map of the Piedmont and Blue Ridge provinces. Data from Williams (1978).

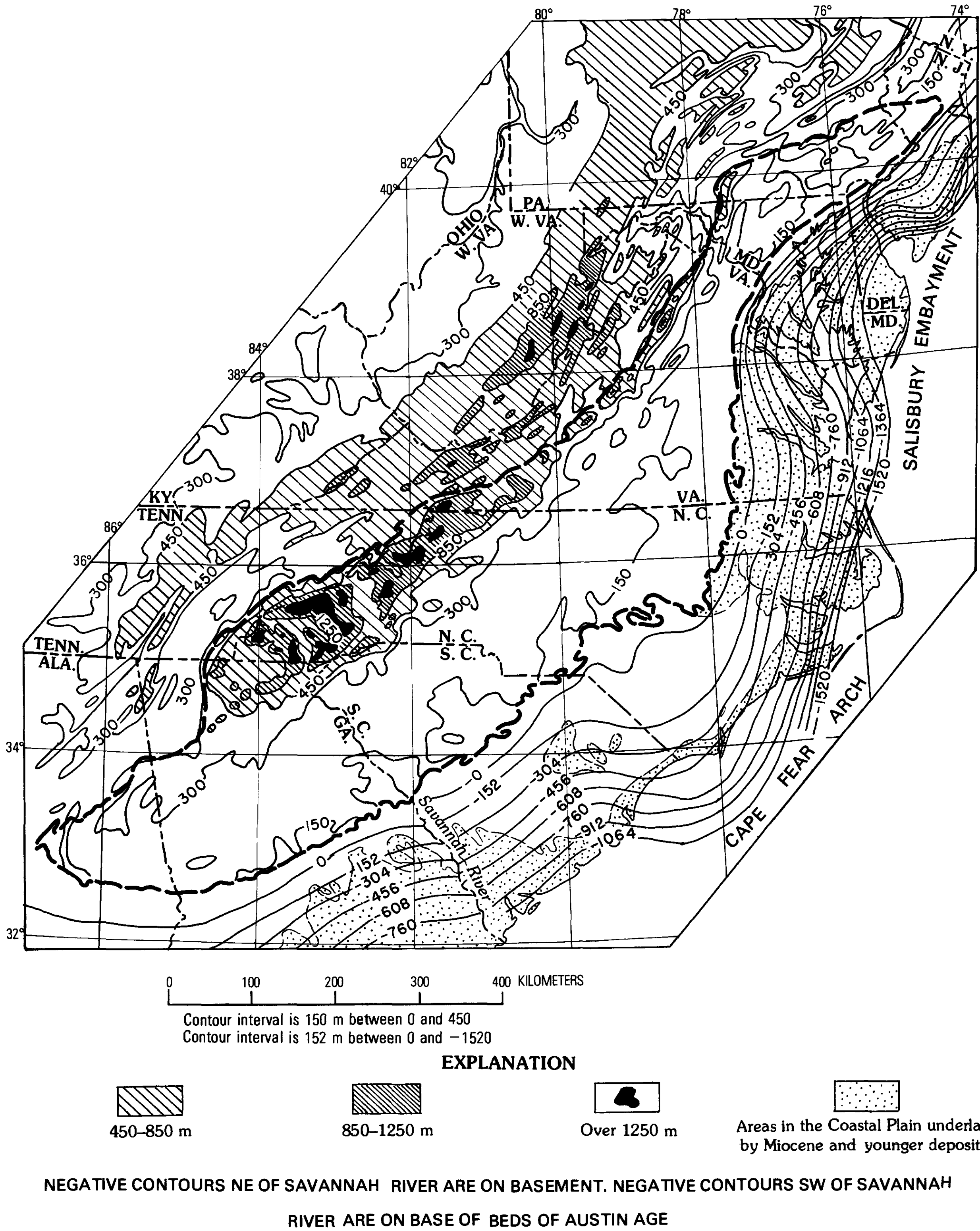


FIGURE 3. – Averaged topographic map of part of the Southeastern United States, showing the crystalline rock area of the Piedmont and Blue Ridge in relation to the Coastal Plain, the Valley and Ridge, and the Appalachian Plateaus. The contours on the land surface show mean altitudes within 3-minute rectangles and were prepared from digitized data (original contour map furnished by M. F. Kane). Generalized contours on the basement are from the "Tectonic map of the United States" (U.S. Geological Survey and American Association of Petroleum Geologists, 1961).

ly 6×5 km in size). The means are computed from thousands of points within each rectangle. This kind of map, of course, does not show the true topography. It leaves out many details that, if shown, would tend to obscure the general topographic form. Such local details commonly are related to rock control of the topography. Thus, a generalized map may be emphasizing forms in part inherited from or caused by tectonic movements.

Figure 4 is a relief map made visually by counting the numbers of 100-ft contours within each 100-km^2 unit of the Universal Transverse Mercator grid, as shown on the 1:250,000-scale 1° to 2° maps of the U.S. Army Map Service and the U.S. Geological Survey. A relief map is quite different from a topographic map in that the values shown vary depending on the definition of the units used. Probably the best method of making a relief map would be to measure or estimate the difference between ridge crests and adjacent valley bottoms. Such measurements could be made by making a series of detailed topographic profiles across an area and subtracting the altitudes of the low points from those of the high points along the profiles (see fig. 6 for an example of such a profile). However, the work involved for so large an area is prohibitive. The method used here gives higher relief values, but the purpose of the map is to distinguish large areas of different relief, particularly within the monotonously low Piedmont province. Relative values can serve this purpose.

Figure 5 is a map of the drainage system of the Piedmont and Blue Ridge that includes all the headwater areas of drainage flowing across the region. The data are from the outline maps of the States (U.S. Geological Survey 1:1,000,000 series).

Figures 6 and 7 are detailed profiles across two areas in the Blue Ridge and Piedmont. These profiles were made using large-scale (1:24,000 and 1:62,500) quadrangle maps by the U.S. Geological Survey. Their locations are shown in figure 4. Other maps of smaller areas are used to illustrate specific areas discussed in the text.

REGIONAL SETTING

The Piedmont and Blue Ridge form an elongate area in the Southeastern United States, 100 to 300 km wide and about 1,375 km long. As the averaged topographic map (fig. 3) shows, the region is asymmetric in the vertical dimension, the high area known as the Blue Ridge being along the northeast margin. A close examination of the map shows that the crest of this range does not correspond to the highest average topography. The topographic high is roughly parallel to the present coastline, whereas the Blue Ridge trends in a more northeasterly direction. Judson (1975) has noted that the general topographic high south of central Pennsylvania may have been formed in early Mesozoic time, as it is related to a low-gravity anomaly and also corresponds to the drainage divide between the Atlantic and Gulf of Mexico drainage. I have suggested that the major Tertiary uplift of the Appalachians also crested along this drainage divide. The details of the forms of individual narrow ranges were determined by rock control (Hack, 1980, p. B14–B15).

The line separating the Piedmont Lowlands from the other divisions of the crystalline rock area (fig. 1) appears to be a major dividing line in the region. It is expressed in the topography by a slight steepening of the regional slope. At the southwestern end of the region, the line terminates approximately where the Tallapoosa River leaves the Piedmont. On the northeast, it intersects the Fall Zone where the Rappahannock River leaves the Piedmont. In Georgia, the altitude of this line is about 300 m. It declines to 180 m in southern Virginia and to sea level at its northern end. This line roughly forms the southeastern boundary of regions having high relief, although it becomes a little vague in central Virginia. The Tallapoosa-Rappahannock line also corresponds in places to a major line on Williams' (1978) tectonic lithofacies map of the Appalachian orogen. It is near the south boundary of his units 4 and 5 (unit 4, fig. 2, in this paper), except in southern North Carolina, where it is in Williams' unit 23 (unit 9 in fig. 2).

The break in slope of the basement rocks at the inner edge of the Coastal Plain is a striking feature (fig. 3). These basement rocks are an eastward extension of the rocks of the Piedmont. The sharp increase in slope in places is the same order of magnitude as the increase in slope of the inner Piedmont and the rise to the Blue Ridge crest. Northwest of the deep Salisbury embayment, however, the rise of the subaerial topographic surface is steep only at the edge of the Coastal Plain. A short distance inland, the rise to the Applachian crest in Pennsylvania is gradual, and the general topographic surface is lower than it is to the south. The Salisbury embayment collects the drainage of some of the largest rivers in the Appalachians, including the Delaware, Susquehanna, Potomac, James, and Roanoke. This embayment and its extension to the south in North Carolina received great thicknesses of sediment, particularly in Cretaceous and Miocene time (Owens, 1970). In late Tertiary time, it also received large quantities of sediment, including sediment from the Hudson River system (Owens and Minard, 1979).

The decline in altitude of the Appalachian crest in Maryland and Pennsylvania may be related to the tectonic system that formed the Salisbury embayment. The embayment and a possible troughlike area to landward may have been responsible for gathering a fanlike network of large rivers.

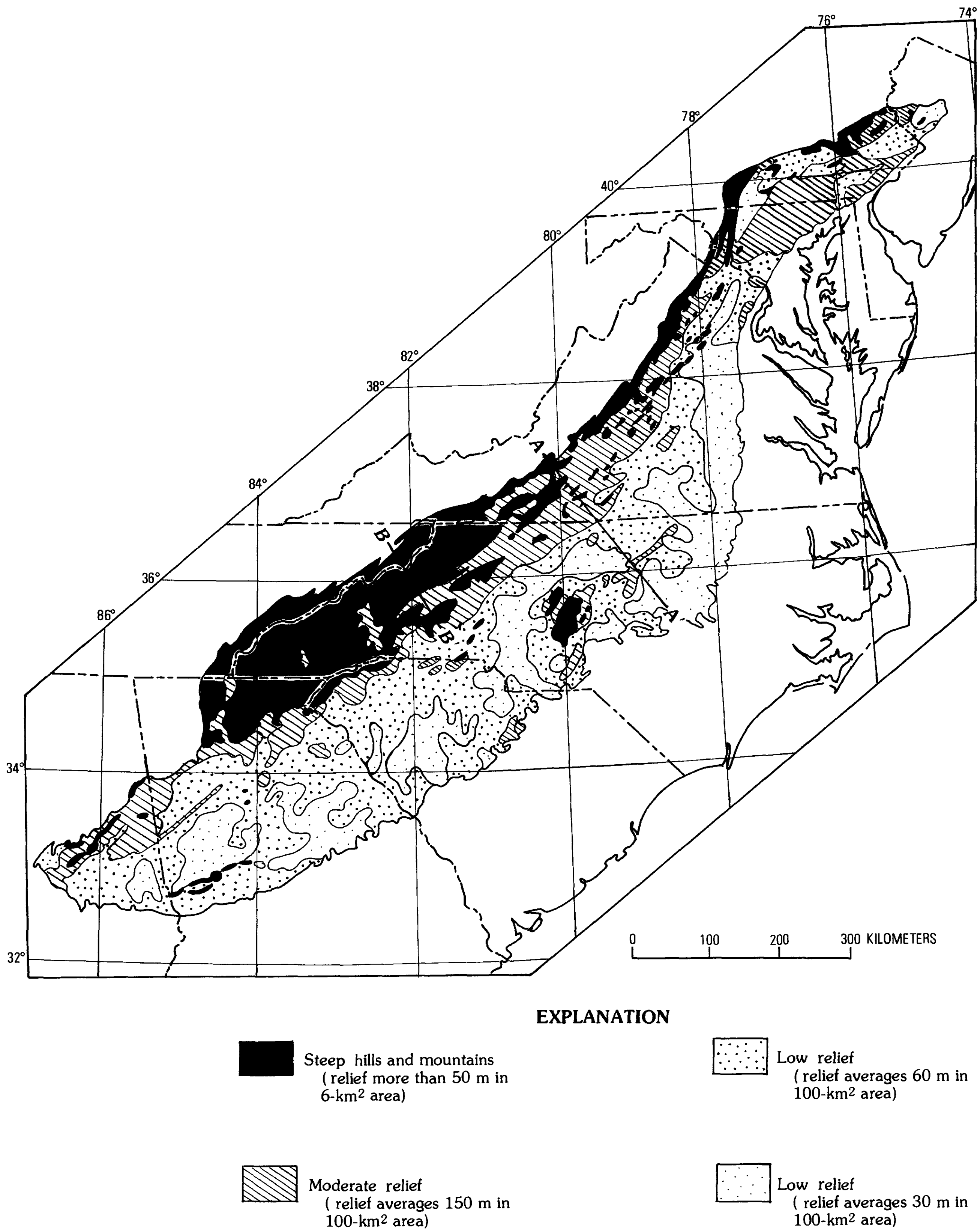
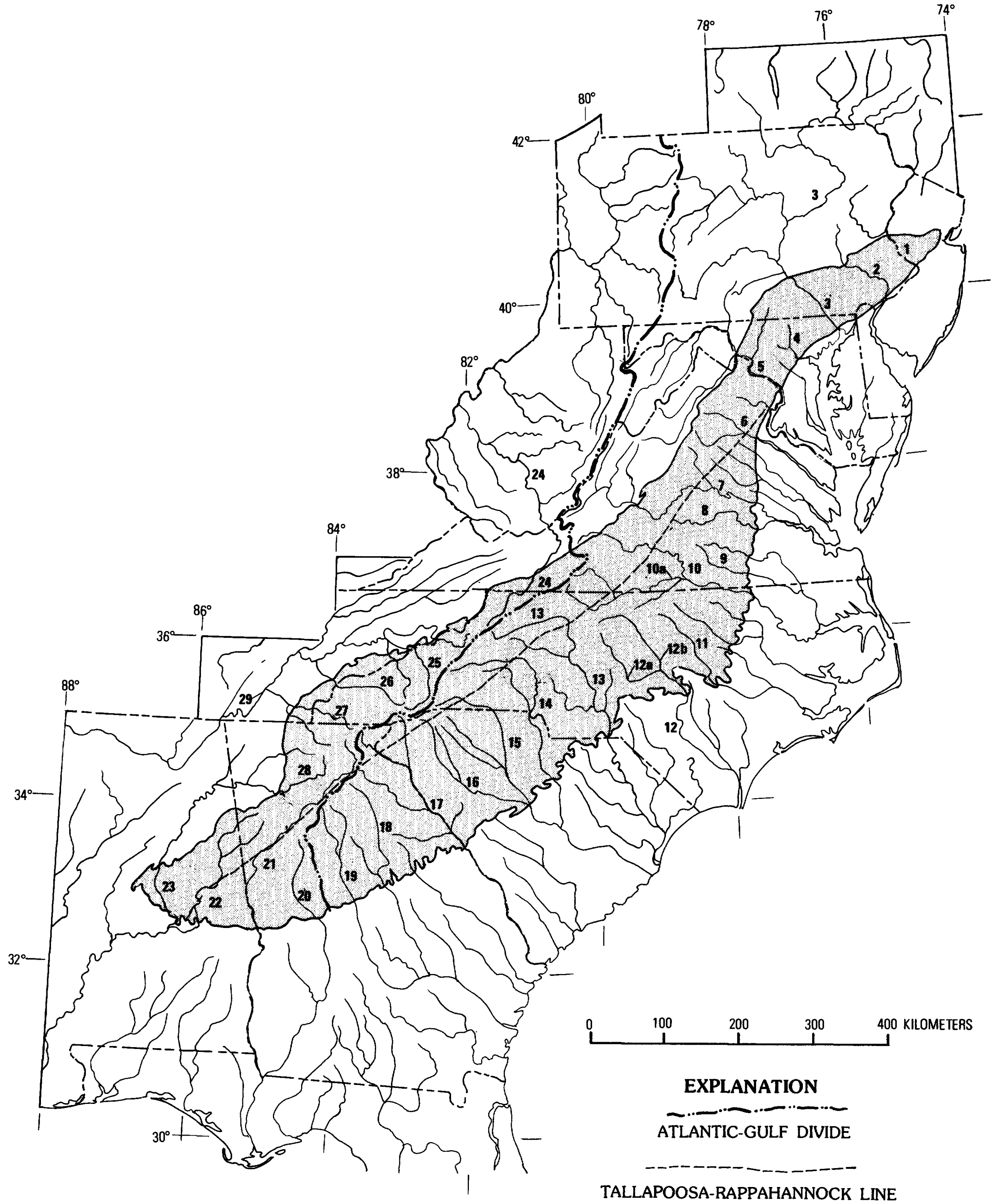


FIGURE 4.—Relief map of the Piedmont and Blue Ridge provinces (based on the number of 100-ft contours within 100-km² areas, as shown on topographic maps at a scale of 1:250,000). Lines A–A' and B–B' refer to topographic profiles of figures 6 and 7.



FIGURES 5.–Drainage map of part of the Appalachian highlands (shaded area), including the Piedmont and Blue Ridge. Rivers are identified by numbers: 1, Delaware; 2, Schuylkill; 3, Susquehanna; 4, Patapsco; 5, Potomac; 6, Rappahannock; 7, James; 8, Appomattox; 9, Meherrin; 10, Roanoke; 10a, Dan; 11, Neuse; 12, Cape Fear; 12a, Haw; 12b, Deep; 13, Yadkin; 14, Catawba; 15, Broad; 16, Saluda; 17, Savannah; 18, Oconee; 19, Ocmulgee; 20, Flint; 21, Chattahoochee; 22, Tallapoosa; 23, Coosa; 24, New; 25, French Broad; 26, Tuckasegee; 27, Hiwassee; 28, Etowah; 29, Tennessee.

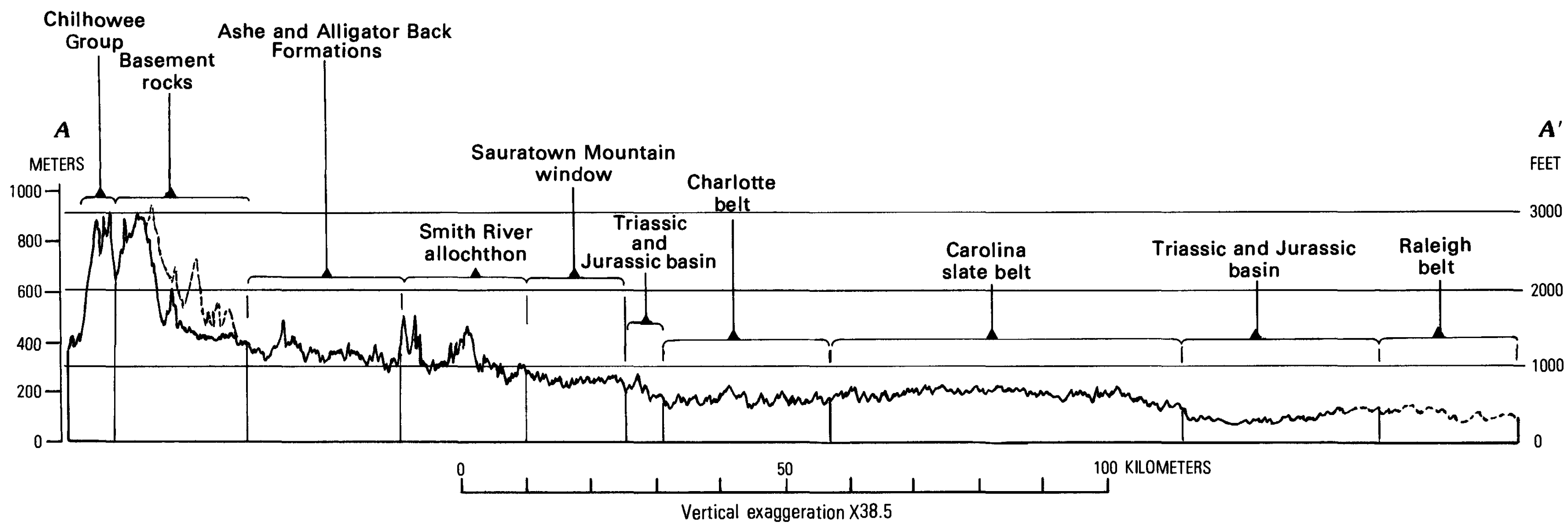


FIGURE 6.—Topographic profile across the Blue Ridge and Piedmont from the Roanoke River near Salem, Va., to the Durham Triassic and Jurassic basin, showing the relation of local relief to rock units. As the profile follows the axis of a valley within the basement rocks, an alternate profile 1.3 km to the northeast is shown by a dashed line in that area. See figure 4 for location (A–A′).

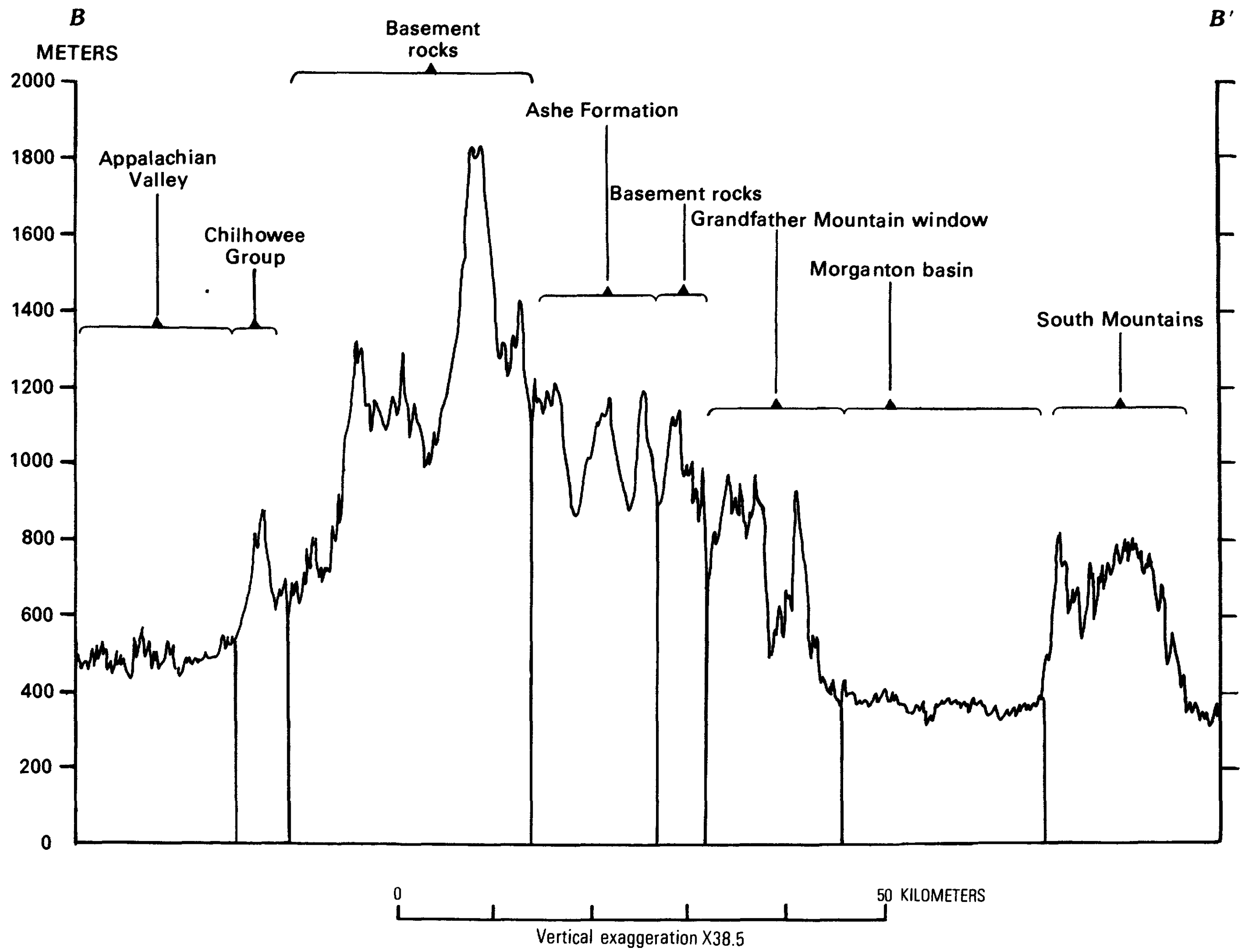


FIGURE 7.–Topographic profile across the Blue Ridge from the Appalachian Valley near Johnson City, Tenn., to South Mountains south of Morganton, N.C. See figure 4 for location (B–B').

A similar but less impressive coalescence of streams occurs landward of the embayment southwest of the Cape Fear arch (fig. 3), where several large Piedmont streams join to form the Peedee and Santee Rivers. These two rivers are less than 10 km apart at the Atlantic coast (see rivers 13 to 16, fig. 5). If the basins of these rivers were at one time gathered together in a trough-like area, that trough is not defined in the present topography and has been destroyed by erosion.

PIEDMONT LOWLANDS

The Piedmont Lowlands comprise most of the area southeast of Williams' (1978) unit 8, which consists of polydeformed and metamorphosed sedimentary rocks that he suggested may have marked the eastern border of the North American continent in early Paleozoic time. The rocks in the Piedmont Lowlands are predominantly feldspathic gneiss and schist intruded by plutons, most of which are granitic, but the area also includes sedimentary and volcanic rocks of lower metamorphic grade. The topographic expression of these rocks is different from that of the gneiss and schist. The largest area of low-grade sedimentary and volcanic rock is the Carolina volcanic slate belt (figs. 1 and 2), which is mostly a low-relief area underlain by argillite, slate, and tuffaceous rocks. In the northern part, feldspathic intrusive and volcanic rocks are more abundant. These rocks are quite resistant and generally underlie hilly topography. The Uwharrie Mountains (fig. 1) exceed 300 m in altitude and have sharp relief. Many hills rise 90 m above their bases. The importance of rock control in shaping hilly areas as well as individual hills is quite evident and is well shown on the geologic map of the Denton quadrangle (Stromquist and others, 1971) and in the field guide by Seiders and Wright (1977).

The Kings Mountain belt is an elongate area 15 to 30 km wide. It consists mostly of metasedimentary rocks, but it also contains granite gneiss, biotite gneiss, metamorphosed quartz diorite, and intrusive granitic bodies. The metasedimentary rocks are primarily nonresistant rocks, such as metasiltstone, phyllite, and marble, but resistant rocks, such as quartzite, kyanite quartz rock, and conglomerate, also occur in the belt. The resistant rocks form chains of low hills. Crowders Mountain near Gastonia, N.C., has a relief of 183 m and rises to an altitude of 460 m. The crest of this mountain, as well as that of Kings Mountain southwest of it, is held up by narrow outcrops of quartzite and conglomerate. Clasts of these rocks cover parts of the lower slopes. As the correlation between local relief and rock type is pronounced in the Kings Mountain belt, I conclude that the unusual topography in this belt is probably controlled by rock resistance.

The Pine Mountain belt of Georgia and Alabama is considered by some geologists as part of the Kings Mountain belt. Although the two are not contiguous, they may be connected by the Towaliga fault (Hatcher, 1972). The rocks in the Pine Mountain belt are similar to those in the Kings Mountain belt as they include micaschist, gneiss, and extensive, although thin, beds of quartzite. An important feature of the area is a downfaulted block containing unconsolidated fine-grained fluvial deposits of Paleocene age on the north border of the belt. These deposits constitute a basin more than 3 km long and 0.8 km wide. They are bounded by a major fault (Towaliga fault) on the north, and by a local fault (Warm Springs fault) on the south. The Paleocene sediments are overlain by sand and gravel washed from Pine Mountain. These deposits are also affected by the Warm Springs fault (White, 1965; Christopher and others, 1980). This occurrence of fine-grained Paleocene sediments shows that part of the Georgia Piedmont 40 km north of the Fall Zone had a cover of Tertiary sediments and also that tectonic deformation took place within the Piedmont in Tertiary time. On the other hand, the high topographic areas within the belt, such as Pine Mountain and Oak Mountain, probably do not owe their altitudes directly to the faults. The high areas are localized along outcrop belts of quartzite and alluvial fans of quartzite clasts. The maximum relief is 150 m, equivalent to the relief at Crowders Mountain in the Kings Mountain belt.

A few small hilly areas other than those in the Carolina slate belt, the Pine Mountain belt, and the Kings Mountain belt also occur, commonly related to small lenticular zones of quartz-kyanite rock. Willis Mountain in central Virginia is perhaps the most spectacular; it consists of a narrow range of hills more than 40 m high that rises to an altitude of 345 m above sea level (Espenshade and Potter, 1960).

The Piedmont Lowlands as defined here (fig. 1) contain several Triassic and Jurassic basins. Except for the Danville basin, the relief of these basins is generally lower than that of the crystalline rock areas surrounding them. The Danville basin, a long narrow basin at the western margin of the area, contains 4,500 m of continental sedimentary rocks, much of which are arkose and graywacke. Because of the arenaceous nature of the rocks, the relief is high, ranging from 45 to 140 m. Whiteoak Mountain, the highest range of hills, is higher than the adjacent Piedmont, but, in general, the relief is comparable with that in the Piedmont (as shown in the cross section, fig. 6). The Farmville and Richmond basins are underlain by shale, thin sandstone, and coal. The relief within areas of 100 km^2 averages about 55 m, in some places reaching 85 m, but differs little from that of the surrounding Piedmont.

The Durham-Sanford basin has generally low relief, somewhat lower than that of the Piedmont to the west (fig. 6.) The rocks range from claystone and shale to fanglomerate. They include carbonaceous clay and mineable coal. The Wadesboro basin is separated from the Sanford basin by overlapping sediments of the Coastal Plain. Its relief is also somewhat lower than than of the adjacent Piedmont and can be accounted for by the lower resistance of the rocks.

STREAM PATTERNS AND EVIDENCE FOR SEDIMENTARY OVERLAP

The river patterns of the Piedmont Lowlands (fig. 5) support the idea that parts of the region were covered by overlapping sediments in late geologic time. In contrast to much of the drainage northwest of the lowlands, the major streams flow directly down the slope of the Piedmont toward the Coastal Plain; two distinct patterns can be clearly distinguished, one southwest of the Cape Fear River (fig. 5, no. 12) and the other north of it.

Southwest of the Cape Fear River, almost the entire Piedmont Lowlands area is drained by large streams that head at the Atlantic–Gulf of Mexico divide. The major trunk streams of highest order flow in rather straight courses down the regional slope; no structural control is apparent. Many of the lower order streams join them at fairly wide angles, and first- and second-order streams commonly are parallel to the prevailing northeast structural trend. This kind of adjustment is especially common in the Slate belt. Only major streams drain into the Coastal Plain, for the inner edge of the plain is a cuesta that is higher than the Piedmont northwest of it. The cuesta is generally penetrated only by the larger streams, and its northwest slope drains into long tributaries of the major Piedmont streams.

Staheli (1976) has suggested that the drainage of the outer Piedmont of Georgia was superimposed from a Coastal Plain cover and extended inland about 190 km, almost to the Blue Ridge front, a concept that explains the lack of structural control of the master streams. The work of Christopher and others (1980) at Warm Springs supports Staheli's concept, because, as they state (citing Cramer, 1979), updip projections of the Paleocene Nanafalia Formation in Georgia suggest that a blanket of sediment did exist in this area. Furthermore, the crossbeds in the Paleocene fluviatile sediments north of Pine Mountain indicate a flow to the south and the absence of a topographic high at that time.

The superposition hypothesis can probably be extended through South Carolina because the outer Piedmont has a drainage pattern similar to that found in Georgia northward as far as the Cape Fear River. The major streams of this drainage system, as will be shown, have basins that are broader than those of most consequent streams of similar length. This width probably is related to the fact that the drainage is perpendicular to the structural trend of the rocks across which they flow. If deep erosion, several hundred meters or more, took place after the original drainage was superimposed on the Piedmont rocks, there would be a strong tendency for tributaries to form parallel to rock structure and at large angles to the main stream at the expense of streams that were perpendicular to the structure. The process would begin as soon as the hard rocks were penetrated by the trunk streams.

A different drainage pattern exists northeast of the Cape Fear River, at least as far as the Rappahannock River (fig. 5, no. 6). In this region, the belt of northeast-trending streams along the inner Piedmont (fig. 5) widens. However, the major rivers that head at the Atlantic–Gulf of Mexico divide, such as the Roanoke and the James Rivers, flow across the outer Piedmont in rather direct courses. Between these streams are many smaller streams that head in the Piedmont but that flow directly into the Coastal Plain. The Piedmont–Coastal Plain boundary is closer to the ocean in this region than it is farther south, and the Coastal Plain sediments overlap the Piedmont without any apparent cuesta. The protrusion of the Piedmont margin, or Fall Zone, to the east corresponds to the inner edge of the Cape Fear arch (fig. 3.) North of the protrusion, patches of fine-grained sediments and river gravels are found some distance west of the Coastal Plain margin (for example, Goodwin, 1970). These features suggest a Coastal Plain overlap of the outer part of the Piedmont that is much younger than the overlap to the south, as explained below. A narrower overlap in the Washington, D.C., area, that is probably of the same age (described by Darton, 1951, and discussed by Hack, 1975), is believed to be of late Tertiary age.

QUANTITATIVE ANALYSIS OF DRAINAGE PATTERNS

The differences in the two drainage patterns described above can be further defined by an analysis of river-basin shapes. One way to compare the shape of drainage basins is simply to measure their lengths along the main stream and then to calculate the average width by dividing the drainage area by the measured length. The ratio of width to length would be a measure of basin shape. These calculations have been done for 11 basins of the outer Piedmont in Virginia and North Carolina (table 1) as well as for 9 of the larger basins north of the Cape Fear River. As expected, the average width of the southern basins is greater than that of the northern basins. The average width-length ratio is 0.28 south of and including the Cape Fear basin and 0.19 north of the Cape Fear basin. This difference is sizable, but, as the data show, the scatter between individual basins is large. One reason for the wide range is the fact that basin shape is partly a function of basin size. Considerable research has shown that, as basins enlarge downstream, they tend to develop lower width-length ratios. In other words, large basins are commonly more elongate than small ones. If there are no geologic restrictions on basin shape, basins 100 km^2 in area on the average have a width-length ratio of 0.33, whereas basins 10,000 km^2 in area on the average have a width-length ratio of only 0.16.

This characteristic of drainage basins was studied by Hack (1957) using data on drainage areas and stream lengths as measured along the stream, including bends and meanders. The data included 92 measurements in the Shenandoah Valley as well as data obtained by Langbein (1947) for 400 localities in the Northeastern United States that included streams whose basins were more than 2,000 mi^2 (5,180 km^2) in area; stream length was found on the average to approximate the function $L = 1.4\,A^{0.6}$, where L is length along the stream in miles as measured from the head of the longest stream in the basin and A is the drainage area of the basin in square miles. As this equation is not dimensionally balanced, the constant of proportionality (in this case 1.4) varies with the units of measurement. In kilometers and square kilometers, the equation is $L = 1.27\,A^{0.6}$. Leopold and Langbein (1962) showed that drainage systems simulated by constructing random walks become more elongate as they enlarge downstream. Their example suggested that stream length was proportional to the 0.64 power of the basin area. Hack (1965), in a study of a stream system in grooved glacial and lacustrine deposits 9,000 years old and younger, showed that restraints on

TABLE 1.—*Mensuration data for drainage basins of the Piedmont and three selected basins in the Coastal Plain*

[Piedmont basins are assumed to terminate at the Fall Zone]

Basin	Length (km)	Area (km^2)	Average width (km)	Predicted length-law width	Difference (percent)	Width-length ratio
Piedmont south of Cape Fear River						
Deep–Cape Fear, N.C.	136	9,680	71	40	+77	0.52
Yadkin, N.C.	317	18,663	58	53	+9	.19
Catawba, N.C.–S.C.	279	11,691	42	43	−2	.15
Broad, S.C.	241	13,263	55	46	+19	.23
Saluda, S.C.	226	6,612	29	34	−14	.13
Savannah, Ga.	233	18,277	78	53	+47	.54
Oconee, Ga.	178	7,337	41	36	+14	.23
Ocmulgee, Ga.	142	6,128	43	33	+30	.30
Flint, Ga.	137	5,239	38	31	+23	.28
Outer Piedmont north of Cape Fear River						
Po River, Va.	36	297	8.2	9.2	−11	0.23
North Anna, Va.	71	1,052	15	16	−6	.21
Little River, Va.	38	281	7.3	8.9	−18	.19
South Anna, Va.	90	1,113	12	16	−25	.13
Appomattox, Va.	162	3,387	21	26	−19	.13
Nottoway, Va.	70	1,193	17	16	+6	.24
Meherrin, Va.	86	1,797	21	20	+5	.24
Fishing Creek, N.C.	64	1,161	7.6	11	−30	.12
Swift Creek, N.C.	61	464	18	16	+12	.28
Tar, N.C.	114	1,793	16	20	−20	.14
Neuse, N.C.	142	4,430	31	29	+7	.22
Coastal Plain						
Edisto, S.C.	211	7,457	35	36	−2	0.17
Salkahatchie, S.C.	129	3,129	24	25	−4	.19
Ogeechee, Ga.	267	7,684	29	36	−19	.11

the lateral development of stream systems changed the value of the constant of proportionality and, to a smaller extent, changed the value of the exponent. Also many large basins, especially very large ones, tend to enlarge in area relative to length as they grow, and thus may have exponents less than 0.5 (Mueller, 1972, 1973; Mosely and Parker, 1973). Very large basins are more likely to be restricted by tectonic, geologic, and other factors than are small basins, but, of course, even small basins can be influenced by various geologic factors. Recently, a study of 155 drainage basins in Hokkaido, Japan (Shimano, 1975), showed that an empirical equation relating stream length and basin area $L=1.413\,A^{0.61}$, expressed in kilometers and square kilometers, fit the data closely. Shimano's work, however, indicates that errors in measuring meanders of different sizes do affect the relationship by increasing the value of the exponent slightly. Shimano also constructed an equation for basin length and area $L=1.224\,A^{0.576}$ in kilometers and square kilometers. This equation avoids the problems involved in measuring meandering streams and is used in the analysis that follows.

In table 1, the average width of the drainage basins south of and including the Cape Fear River basin and 11 of the basins north of the Cape Fear River are compared with the basin width that could be predicted if their dimensions followed the length law as determined by Shimano (1975). Three large rivers of the Coastal Plain are also included in the table. The equation for normal or predictable basin width can be determined directly from the basin area by using Shimano's length-width equation and is $W=0.82A^{0.424}$.

As the table shows, the streams south of and including the Cape Fear basin are mostly much wider than the predictable width based on the length law as determined by Shimano. This is interpreted to mean that the basins have a tendency to form long tributaries because of some structural control of all but the major streams. Two exceptions are the Saluda River basin (14 percent narrower than the length-law width) and the Catawba River basin (2 percent narrower than the length-law width). The shape of the Saluda River basin is readily explained, for that river flows in and parallel to a swarm of diabase dikes of Triassic age shown on the geologic map of the United States (King and Beikman, 1974). These dikes can be expected to have strongly restricted the lateral growth of the basin. The Catawba River basin is exceptional also in that it has a long reach upstream in the inner Piedmont in which the river course is strongly controlled by rock structure. This region is presumably upstream and northwest of the area of superposition, so that the main stream itself is structurally controlled.

The 11 smaller river basins north of the Cape Fear River that originate on the Piedmont are considerably narrower. They average about 8 percent narrower than their length-law widths, and they resemble closely the three rivers of the Coastal Plain that are also included in the table. Presumably, these streams have patterns similar to the normal patterns of consequent streams whose basins have formed unrestricted by geologic or tectonic factors. This assumption is consistent with the concept of a late overlap of the outer Piedmont north of the Cape Fear River by either marine or fluvial deposits. As the drainage patterns have not been much affected by the trends of the rocks in the Piedmont, I assume that the overlap could probably have taken place in late Tertiary time, as studies in the Washington, D.C., area suggest (Darton, 1951). The area of overlap by upper Tertiary Coastal Plain sediments corresponds to a steep gradient of the submerged and buried crystalline basement and is associated with the deep Salisbury embayment (fig. 3).

The analysis of basin width in the Piedmont supports the idea that the large southern drainage basins have been affected by geologic restrictions imposed during a long period of erosion occupying most of the Cenozoic. The shorter streams of the outer Piedmont to the north are probably much younger, and their original pattern, developed on a sedimentary cover, has not been substantially altered.

REGIONAL PHYSIOGRAPHIC FEATURES RELATED TO THE PINE MOUNTAIN BELT

The Chattahoochee and Flint Rivers (20 and 21, fig. 5) have anomalous profiles in the outer part of the Piedmont. Both rivers descend abruptly about 90 m between the northern part of the Pine Mountain belt and the Fall Zone, a distance of about 40 km. The average gradients in these rocks are 0.0024 for the Chattahoochee River and 0.0021 for the Flint River, greater than the gradients upstream by factors of 5.3 and 4.2, respectively. Rock control is a possible explanation, as the two streams must cross several outcropping layers of resistant quartz rock within these steep reaches that would increase the caliber and resistance of the bedload. On the other hand, the Chattahoochee River is much larger than the Flint, having a drainage area of 11,450 km^2 at the Fall Zone, compared with 5,240 km^2 for the Flint. One would expect that the gradient of the Chattahoochee River would be gentler than than of the Flint if both rivers had adjusted their profiles. Upstream for a distance of 96 km from the Pine Mountain Belt, the average gradient of the Chattahoochee is 0.00045, whereas that of the Flint is only slightly steeper, 0.00050.

When the profiles of these two rivers are compared with those of the Ocmulgee and Oconee, the neighboring streams to the east, one sees little similarity. The Ocmulgee and Oconee show a concave profile as they descend the Piedmont, whereas the Chattahoochee and Flint Rivers have a sharp change in gradient just above the Pine Mountain belt that makes their overall gradient convex. The low gradient of the Flint River upstream is reflected in the low relief of the region north of the Pine Mountain belt (fig. 4), an indication that the Pine Mountain belt has acted as a barrier to slow the erosion of the river basin upstream relative to the bordering areas. The steep reaches in and below the Pine Mountain belt may be related entirely to quartzite in the bedload of these rivers. However, one can speculate that the initial convex profiles of these two rivers were formed while faulting was in progress, that both streams were then affected to the same degree, and that time has not been sufficient since faulting ceased for the profiles to become adjusted in relation to the relative sizes of the rivers. On the Chattahoochee River, the sharp break in slope that presumably should locate the most recent fault is below the Goat Rock fault at or close to the inner edge of the Coastal Plain. On the Flint River, the sharp break in slope is within the Piedmont at or near the Goat Rock fault.

DIFFERENTIAL MOVEMENT ALONG THE FALL ZONE

South of the Roanoke River, most of the streams of the Piedmont enter the Coastal Plain more or less at grade, that is, without a sharp irregularity in the profile (fig. 8). The Chattahoochee and Flint Rivers are exceptions, as explained above. The Savannah River has a short steep reach at the Fall Zone in Augusta, Ga., in which the river drops 11.7 m in 5.5 km. This slope is about 0.002, compared with 0.00056 in a more typical 16-km reach upstream (data from U.S. Army Corps of Engineers). The steep reach crosses the Belair fault believed to be active in the Tertiary (Prowell and O'Connor, 1978).

The Cape Fear River (fig. 8) has a prominent steep reach at the Fall Zone. The steep reach is just above North Carolina Highway 217 near the town of Erwin. The river drops 12 m in 8 km, a slope of 0.0015, compared with 0.0003 in a 24-km reach upstream (data from U.S. Army Corps of Engineers). No late fault has been reported in this area. Other more pronounced steep reaches occur at the Fall Zone on the Roanoke, Appomattox, James, and Rappahannock Rivers. On the James River, a fault zone of Tertiary age is reported just downstream from the head of tidewater at Hopewell, Va. (Dischinger, 1979). The Rappahannock River

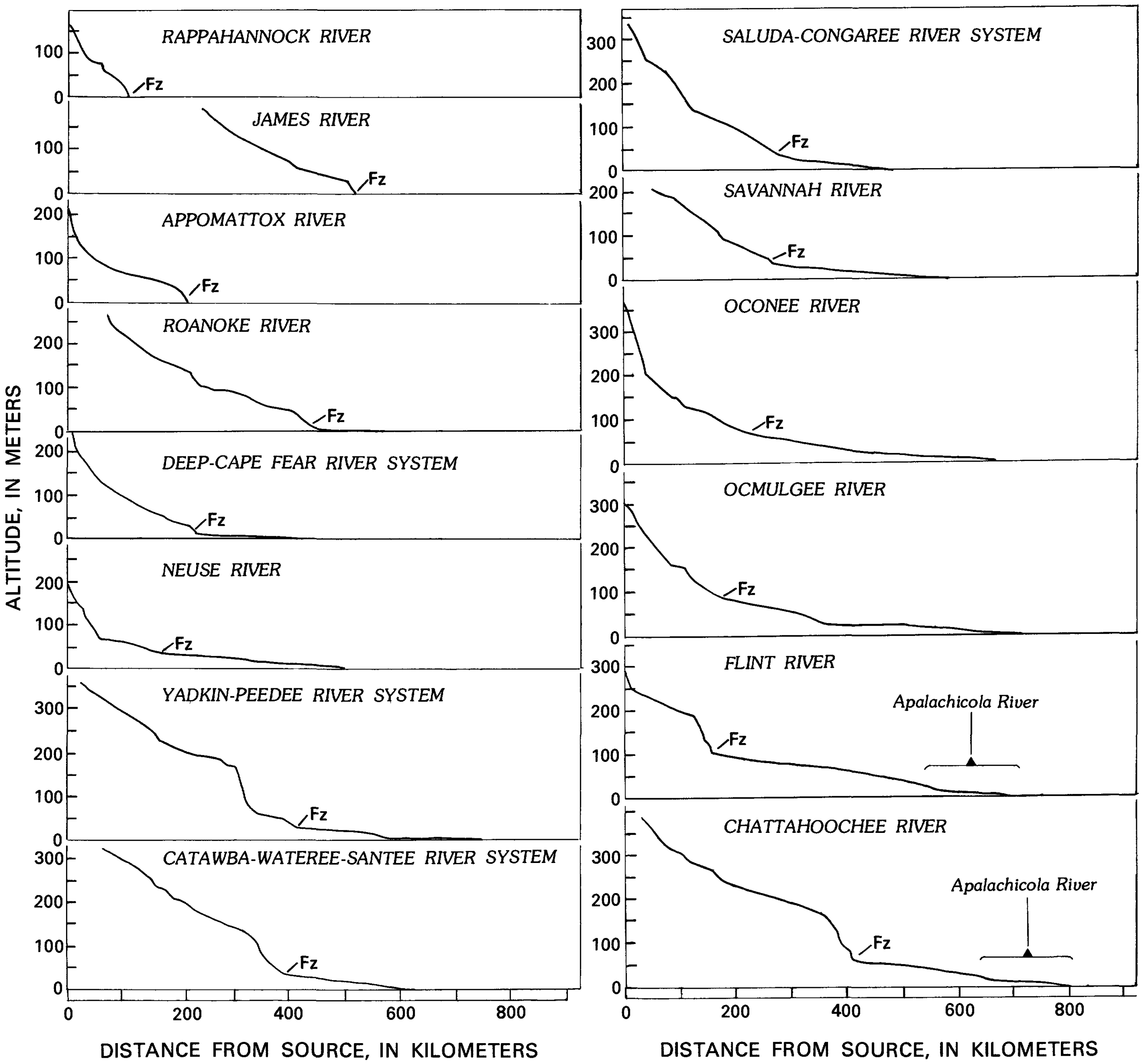


FIGURE 8.—Profiles of 14 streams in the Piedmont Lowlands. Only those parts of the profiles within the Piedmont are shown. Distances along the profiles are measured from the stream heads. Fz, Fall Zone.

crosses a fault zone of Tertiary age at Fredericksburg, Va., in the Fall Zone (Mixon and Newell, 1977).

NORTHEASTERN HIGHLANDS

The Northeastern Highlands is an area of substantially higher relief than the Piedmont Lowlands. It extends from northern Virginia to the Delaware River and is bounded on the southeast by the Fall Zone and on the west and north by the Blue Ridge Mountains and the Appalachian Valley. The highlands include a strip of interconnected Triassic and Jurassic basins on the northwest margin, in general lower than the highlands proper. The area southeast of the Triassic and Jurassic basins in Maryland and southeastern Pennsylvania is comparatively high. It was recognized by Campbell (1929, 1933) as distinctly different in relief from the surrounding Piedmont; he referred to it as the Westminster anticline, an upwarped peneplain. Cleaves and Costa (1979) have cited evidence indicating late geologic uplift of the Maryland Piedmont. I described the area briefly in an earlier paper (Hack, 1980) as an example of an area uplifted in late geologic time relative to other parts of the Piedmont. The evidence includes (1) higher relief, (2) a steep gradient in the basement beneath the Coastal Plain, (3) differences in the gradients of sheets of river gravel of Cretaceous and Miocene age, and (4) the Stafford fault zone at the Coastal Plain boundary described by Mixon and Newell (1977). In this paper, additional evidence based on stream profiles and valley forms is discussed.

The relief of the Northeastern Highlands is in itself a distinctive feature suggesting an uplift rate higher than that in other Piedmont areas. In the center of the crystalline rock area, the relief exceeds 125 m within many areas of 100 km², whereas, in much of the Piedmont, relief is less than 90 m in areas of that size. It is also of interest that, although the Triassic and Jurassic basins included in this area (the Gettysburg and Newark basins) contain large areas of low relief, generally of shaly rocks, the more resistant rocks within them generally have relief comparable with that in the adjacent highlands. For example, in the Frenchtown quadrangle of the Newark basin (Drake and others, 1961) a 100-km² area of fanglomerates has a total relief of 220 m, equivalent to the relief in the crystalline rocks of the adjacent Reading Prong to the north. The Reading Prong is a part of the New England physiographic province, not included in this report. Areas of similar and even higher relief of as much as 270 m occur on diabase sills and quartzose fanglomerates in the Gettysburg basin. None of the Triassic and Jurassic basins in the Piedmont Lowlands has comparably high relief, indicating that the areas of Triassic and Jurassic rock within the Northeastern Highlands probably are in a different tectonic setting and are within the uplift area.

Stream profiles and valley forms, not previously analyzed, are useful features for analysis. As the topographic map (fig. 3) shows, the highlands are cut by two deep and narrow valleys, those of the Potomac and the Susquehanna Rivers. These long streams originate in the Appalachian Plateaus. They are referred to here as exterior streams because they originate outside the Piedmont and Blue Ridge. No similar deep valleys exist in the Piedmont Lowlands. Figure 9 shows the longitudinal profiles of these rivers and other rivers of the subprovince. Distances on the horizontal scale are measured from the source of the longest tributary, but, except for the Patapsco River, the profiles include only the lower reaches. Consider first the Susquehanna River. This stream is the largest crossing the Piedmont in terms of drainage area and discharge, yet its profile is strongly convex. The convexity begins near Selinsgrove,

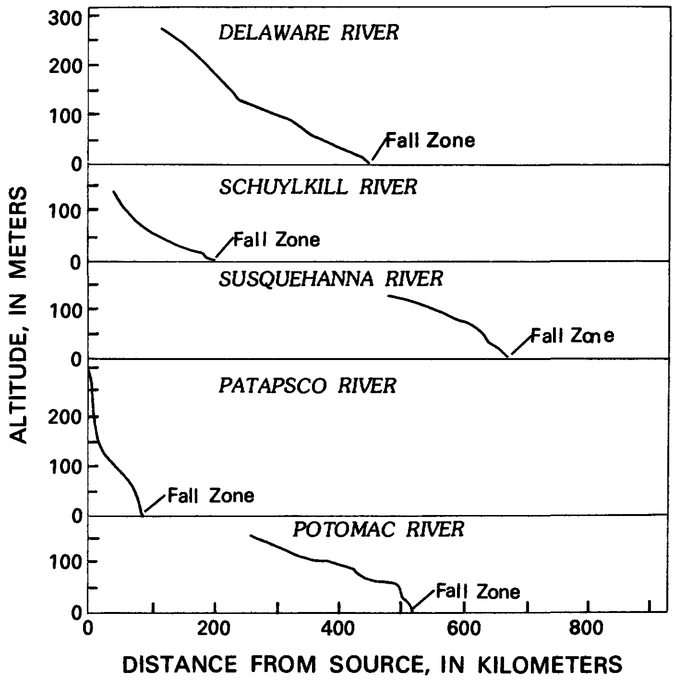


FIGURE 9.—Profiles of five streams in the Northeastern Highlands. In all the streams except the Patapsco River, the upper reaches are not shown, but distances are measured from the stream heads (along the longest tributary). All except the Patapsco head in the Valley and Ridge or Appalachian Plateaus provinces.

Pa., 78 km upstream from the Piedmont. The channel slope continues to steepen through the Triassic and Jurassic belt and in the crystalline rocks. Although it is not shown in the figure, the steep slope continues below sea level, and at the river mouth, the drowned Pleistocene channel is bordered by steep walls now covered by sediment. Its altitude at the point where it enters Chesapeake Bay is −45 m.

The Potomac River is similar, but the convexity is not so regular. The river has a steep concave profile in the Valley and Ridge province. At Harpers Ferry, W.Va., it drops more steeply as it passes through the Blue Ridge. In the Triassic and Jurassic rocks, it again assumes a concave profile. However, 29 km above tidewater, it enters the crystalline rock area and plunges through a deep gorge to the Coastal Plain. The average slope in the gorge is 0.009. Below the head of tidewater, the Pleistocene channel is buried in sediments, but 19 km downstream at Alexandria, Va., borings show the bedrock channel at −30 m. This is an average slope of about 0.0016, much less than the slope in the Piedmont.

The Delaware River (fig. 9) has a more gentle slope in the lower reaches, averaging about 0.0006 in the Piedmont. At the Fall Zone at Trenton, N.J., the river turns sharply south. Sixteen kilometers downstream, the elevation of the bottom of the buried Pleistocene channel is -15 m. The average slope in this reach is 0.00095, somewhat steeper than the slope upstream. The opposite situation occurs in the Potomac, where the buried channel is much gentler than the subaerial channel in the Piedmont. The Schuylkill River (fig. 9) has a concave profile in the Piedmont, but it is notched at the Fall Zone.

The Patapsco River is the only one of the group shown in figure 9 that originates within the highlands. As can be clearly seen, it steepens as it approaches the Coastal Plain. Note that its profile is similar to that of the Rappahannock River (fig. 8). These data seem to indicate that the Susquehanna, Patapsco, Potomac, and Rappahannock Rivers were all affected by differential movement at the Fall Zone in late geologic time when the Piedmont was uplifted with respect to the Coastal Plain. The characteristics of the Schuylkill and Delaware River profiles are different. These rivers may be northeast of the zone of latest disturbance.

The conclusion that uplift took place in late geologic time within the Northeastern Highlands can be further supported by evidence of an evolutionary development of the topographic forms. The Patapsco River basin (fig. 10) is in the central part of the highlands on the east side

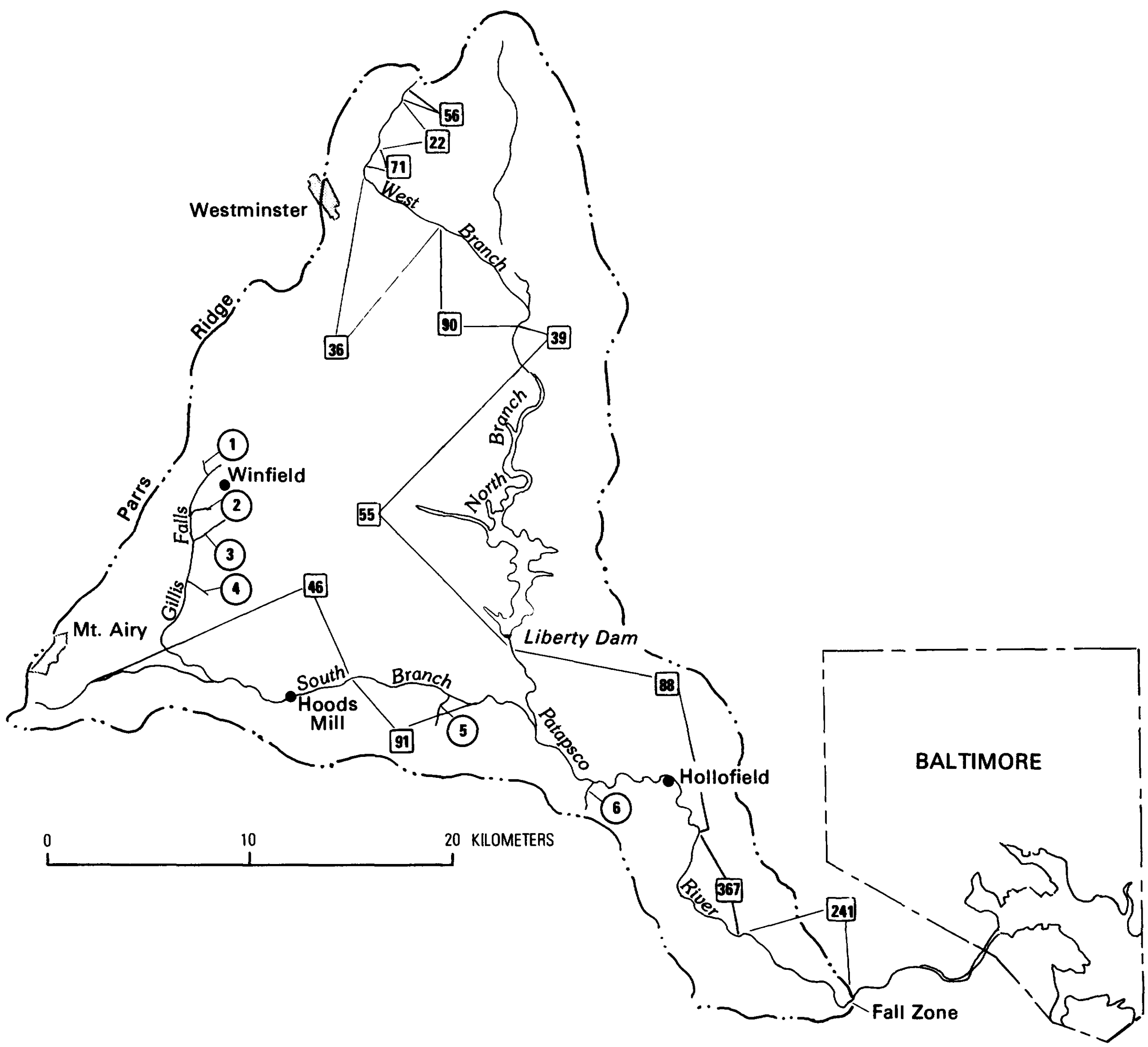


FIGURE 10. – Map of the Patapsco River basin, Maryland. Numbers in circles identify first-order tributaries shown in figure 14. Numbers in rectangles are gradient index values, in meters, of the reaches shown. The gradient index is essentially the slope of the logarithmic profile along a river reach and is a crude index of stream power. It is measured by obtaining the ratio of the fall within the reach, in meters, to the difference between the natural logarithms of the river lengths at each end of the reach, as measured from the source of the stream.

of the high divide, known as Parrs Ridge, that separates the drainage toward Chesapeake Bay from that toward the Triassic and Jurassic basins. In the lower eastern part of the basin, slopes are steeper and power is greater than they are in streams near the divide, as might be expected in a landscape that was uplifted relative to the Coastal Plain in late geologic time. The process that has produced these features resembles the classic concept of landscape rejuvenation by headwater migration of steepened slopes.

Consider first the profile of the Patapsco River itself. The profile is plotted on a logarithmic scale (fig. 11), starting at the stream head farthest from the river mouth, on this river, the Cranberry Branch north of Westminster, Md. Streams having normal profiles generally plot on a straight line on such a graph. A normal stream in this context is one in which discharge and slope are inversely related so as to conserve the power of the stream to transport a load of a certain caliber at approximately the same rate along its course. Figure 12 is the logarithmic profile of the Appomattox River as an example for comparison. The Appomattox River crosses a part of the Piedmont covered by thick saprolite and weak sediments of a Triassic and Jurassic basin. It has a smooth profile and flows through banks of silty material. The river bars are composed of sand and fine gravel. Except for the first 4 km below the source and the 30 km above the Fall Zone, the Appomattox River has a remarkably smooth profile that follows fairly closely the same parabolic curve.

In contrast, the Patapsco River (fig. 11) has a logarithmic profile that becomes increasingly steep as it approaches the river mouth at Baltimore Harbor. In general, this profile seems not to be related to the rocks in the basin. Locally, however, changes in slope in places correspond to geologic contacts. Such a change is most obvious in the lower part of the river where it cuts through the gabbro of the Baltimore Complex.

The bed material in the Patapsco River basin becomes coarser in a downstream direction. Estimates of the median size of bed material were made at several localities, and data for Gillis Falls (fig. 10), a headwater tributary of the South Branch, have been published (Hack, 1957). Median size of bed material in this tributary increases from 7 mm at a distance of 1 km from the source to 40 mm at a distance of 13 km. At Hoods Mill on the South Branch, the median size has increased to 80 mm, and at Hollofield on the main stem of the Patapsco River, the median size is 170 mm. This locality is in the Baltimore Gneiss. At extreme low water, the riverbed resembles a

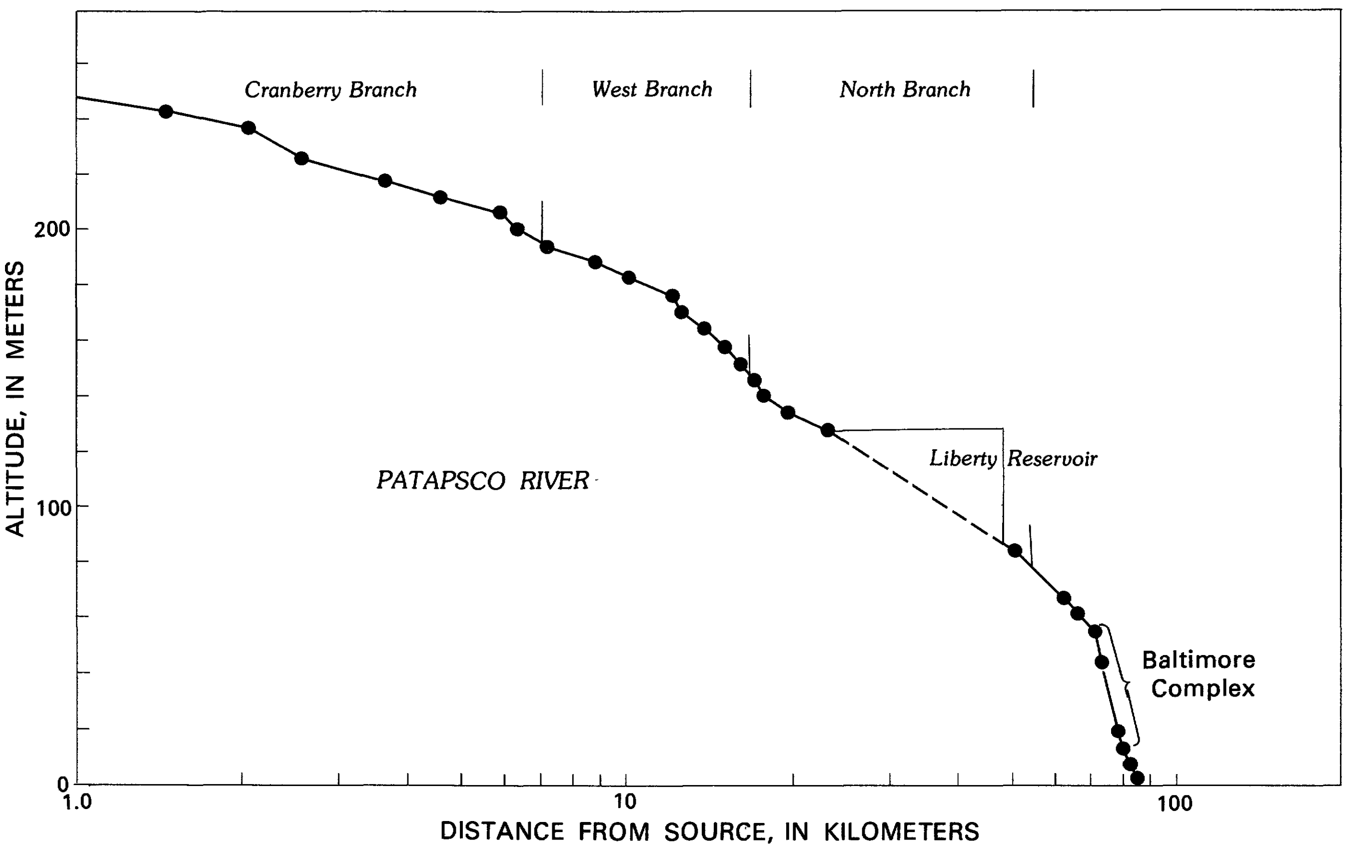


FIGURE 11.–Logarithmic profile of the Patapsco River, Maryland.

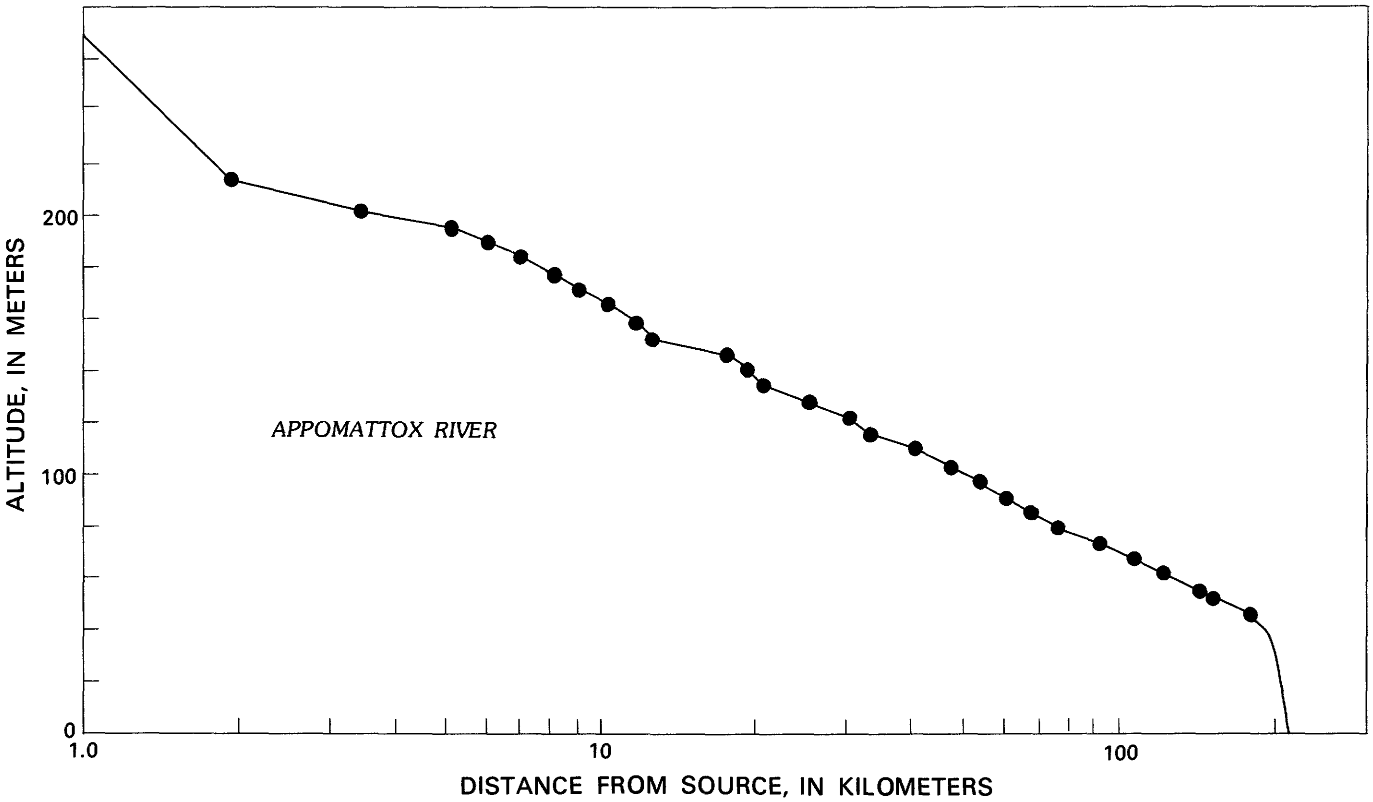


FIGURE 12.—Logarithmic profile of the Appomattox River, Virginia.

boulder field. The channel at this locality is 50 m wide and 1.6 m in average depth. These data are dramatic evidence that the power of this river system increases substantially toward the mouth.

The valley forms also change downstream; the side slopes become steeper and higher (fig. 13), and tributary streams increase in slope in the lower part of the basin (fig. 14). The streams shown in figure 14 were selected because they were approximately the same length from source to junction with the master stream.

Hypsometric analysis of quadrangles in this area shows that the topography is plateaulike in the sense that the largest part is upland. For example, a 100-km^2 area of the Winfield Quadrangle that lies on the drainage divide at the crest of Parrs Ridge has an elevation-relief ratio (Pike and Wilson, 1971) of 0.6. In other words, the mean relief is 60 percent of the total relief. A normal ridge-and-ravine landscape characteristic of most of the Piedmont where isolated high hills are lacking would have a ratio of approximately 0.5. Hammond (1964), in his study of the landforms of the United States, classified the entire crystalline rock area of the Northeastern Highlands as tablelands having moderate relief. In this terrain category of Hammond's, 50 to 80 percent of the area is gently sloping, local relief is 300 to 500 ft (within areas 6 mi across), and 50 to 75 percent of the gentle slope is on the upland.

The topographic features of the crystalline rock area in the Northeastern Highlands seem to point to uplift in late geologic time as the explanation for the systematic change toward the interior of the area and away from the Coastal Plain. They suggest uplift at the Coastal Plain boundary or possibly upward tilting toward the northwest. The limestone valley and Triassic and Jurassic basins behind the crystalline rock, on the other hand, have large areas at lower altitudes, particularly in the southern part of the Gettysburg basin, probably because this area is underlain almost entirely by nonresistant rocks such as limestone, siltstone, and shale. Where crystalline rocks such as diabase and resistant sedimentary rock such as fanglomerate do occur, relief is equivalent to that of the highlands on the southeast.

Probably the uplifted area extends far to the northwest, and the Northeastern Highlands are simply the southeastern margin of the Appalachian Mountain chain. They are not an anticlinal area, as Campbell (1929) believed, but a monoclinal area.

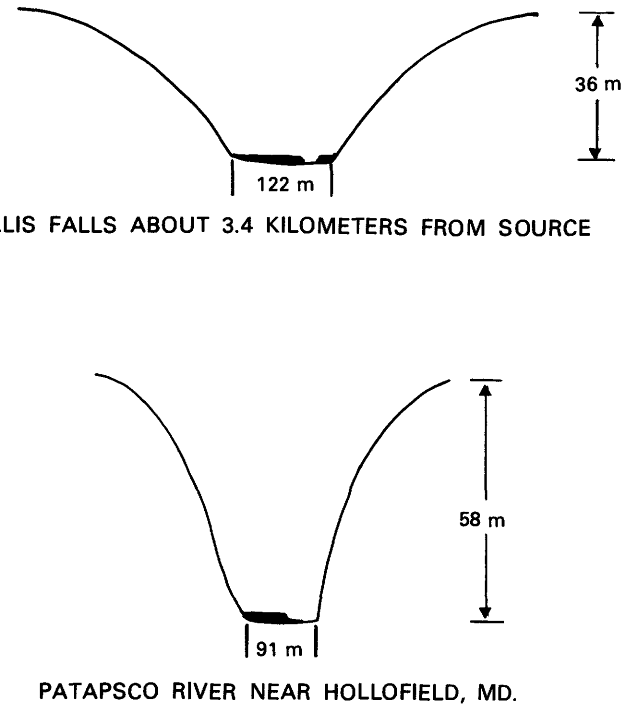


FIGURE 13.–Transverse profiles of two valleys in the Patapsco River basin.

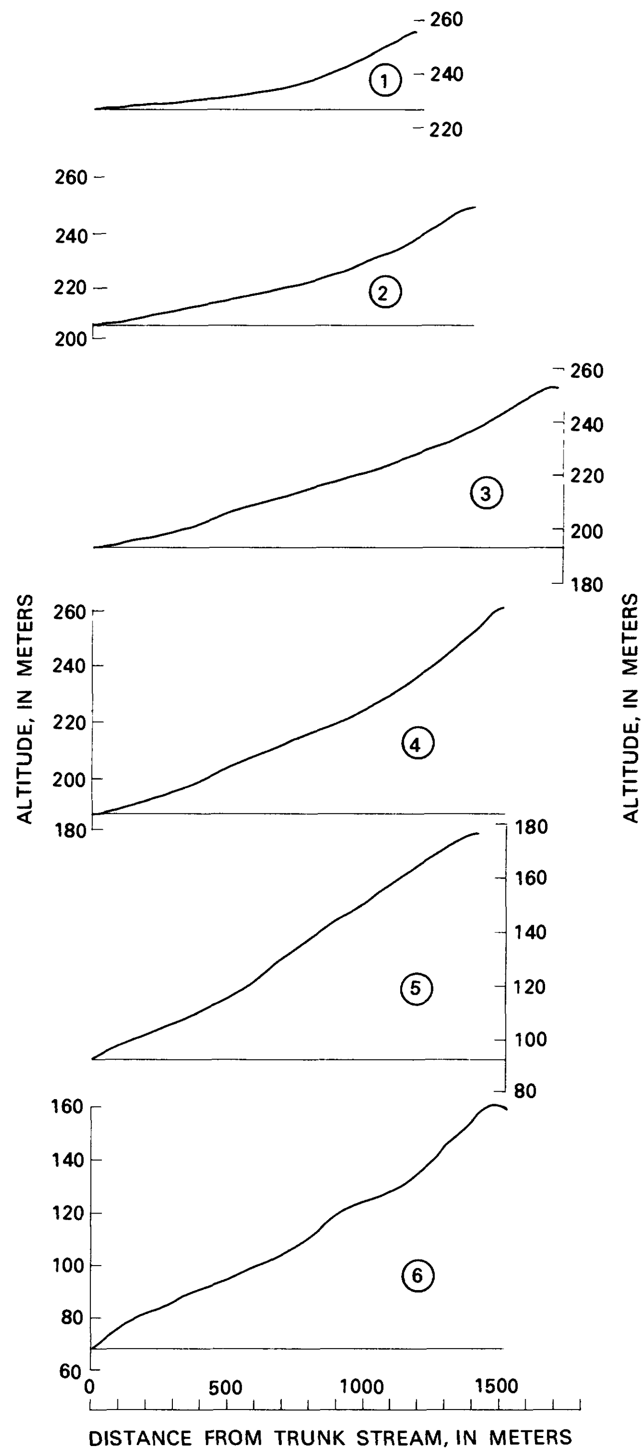


FIGURE 14.–Profiles of some first-order tributaries in the Patapsco River basin. For locations, see figure 10.

SOUTHWESTERN HIGHLANDS

The Southwestern Highlands is similar in some ways to the Northeastern Highlands. Both areas are underlain in large part by polydeformed metasedimentary and volcanic rocks of Williams' (1978) map units 8 and 2. In the Southwestern Highlands, these rocks are bordered on the northwest by a band of clastic rocks of lower metamorphic grade, many of them fine grained and interbedded with quartzite. As in the Northeastern Highlands, the relief is moderate, although in places, it is much higher than that in the Piedmont Lowlands. The highest parts of the area are along the northwest margin, where Cheaha Mountain (730 m) and Talladega Mountain form a curving ridge held up by quartzite in the phyllitic rocks of the Talladega Slate. Figure 15 is an averaged topographic map showing the setting of the Southwestern Highlands in relation to the Blue Ridge and Piedmont. Figure 16 shows generalized topography in relation to rock types.

The physiography of the area has been described as a series of plateaus. The northern, or Georgia, part was divided into three areas by La Forge and others (1925).

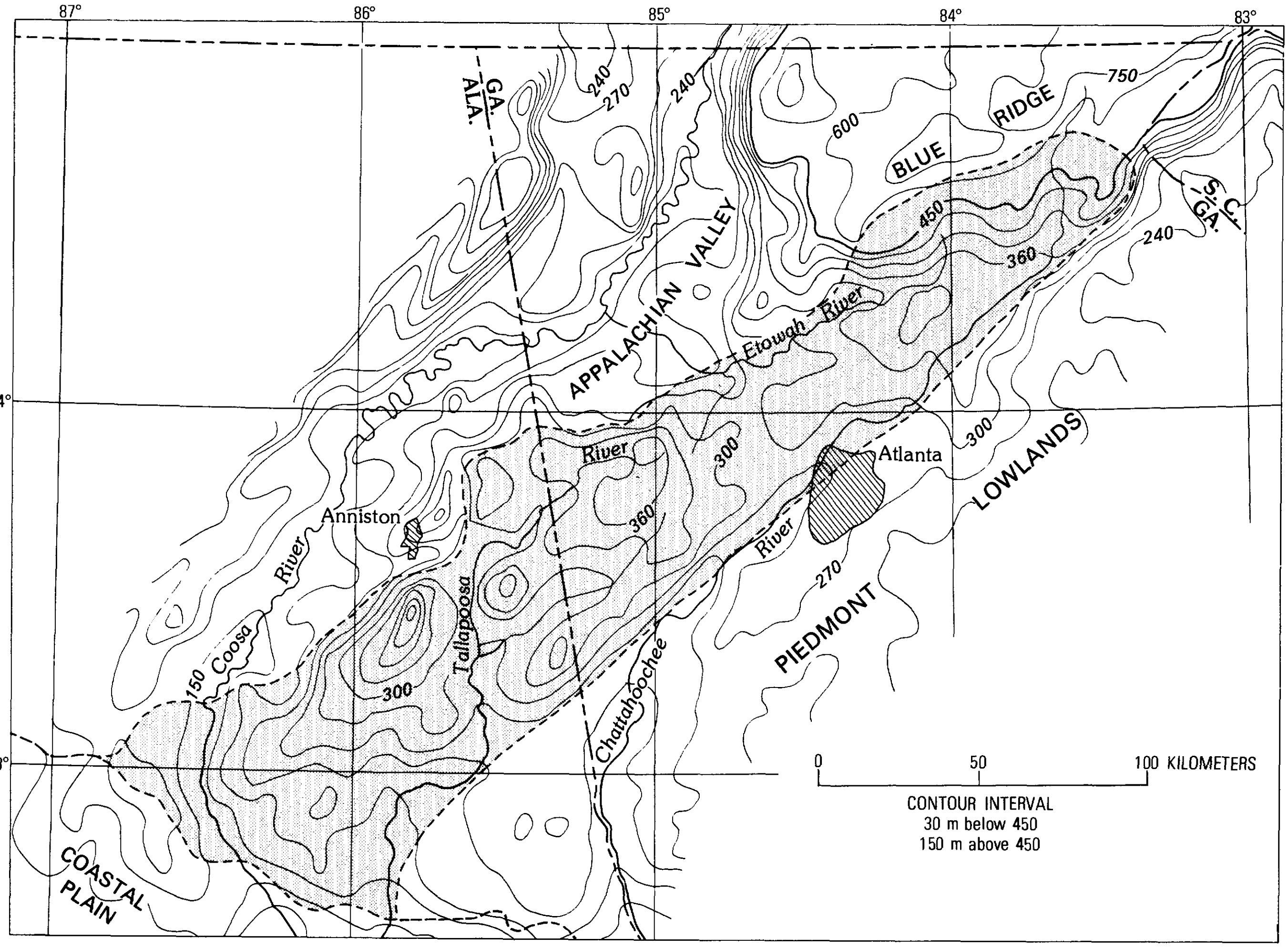


FIGURE 15. – Averaged topographic map of the Southwestern Highlands and surrounding area (shaded). Altitudes are based on the mean altitude of points within 3-minute rectangles.

The Dahlonega Plateau at the north is a dissected upland. The lower Atlanta Plateau extends westward into the area, and the Tallapoosa Plateau constitutes the higher area to the southwest. The Alabama section was called the Ashland Plateau by Johnston (1930). These areas are not true plateaus, as the landscape forms are little different from those of the rest of the Piedmont but are higher in altitude and relief and have steeper slopes.

A prominent feature of the area is an escarpment more than 30 m high that separates the Tallapoosa Plateau or upland from the Atlanta Plateau to the northeast (A, fig. 16). This escarpment corresponds to the drainage divide separating the headwaters of the Tallapoosa River from small tributaries of the Chattahoochee River on the east and from the Etowah River on the northeast. The Etowah River drains into the Coosa system in the limestone and shale valley to the northwest at a much lower level. The Tallapoosa River system is higher than the other two systems, probably because of rock control. Steep reaches occur in the volcaniclastic rocks as well as in the feldspathic rocks above the junction with the Little Tallapoosa River.

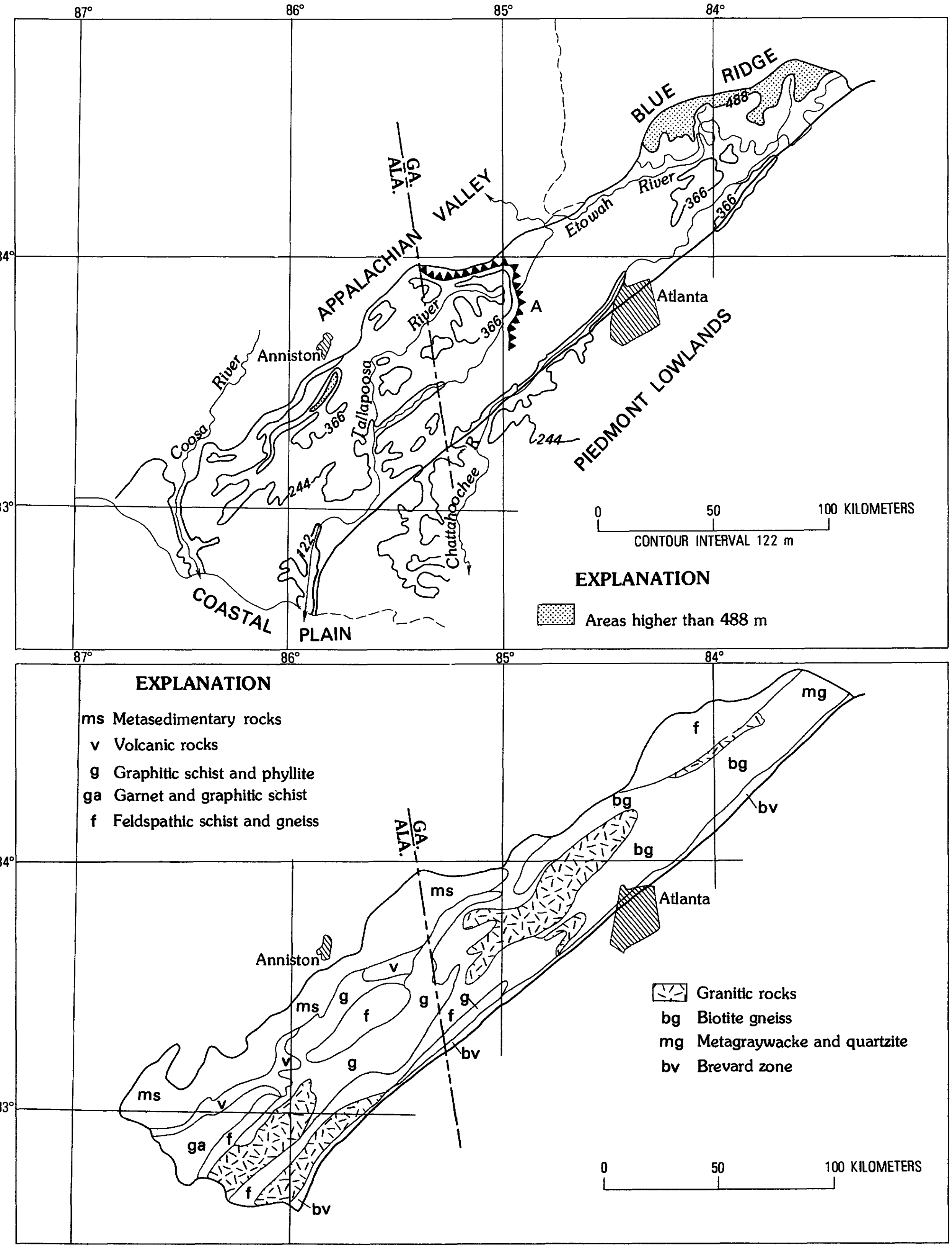


FIGURE 16. – Simplified topographic and geologic maps of the Southwestern Highlands. Geologic data from Neathery and others (1976).

Many local features of the topography of this region appear to be related to rock control. On the other hand, the topography and relief on all the rock types become lower as the Coastal Plain is approached on the southwest. The general altitude of the landscape also declines to the southeast, and relief becomes distinctly lower near the Brevard zone (fig. 15).

The profiles of the Coosa and Tallapoosa Rivers (fig. 17) suggest substantial uplift of the area relative to the Coastal Plain in late geologic time. The Coosa River passes through the area in a broad low valley, but its gradient is steep (0.00095) for a river of such large size. As river profiles prepared by the U.S. Army Corps of Engineers show, in the lower 60 km of the Piedmont, the river was intrenched as deeply as 130 m before impoundment by a series of dams. The area of the drainage basin of this river is about 2,600 km^2. In terms of average discharge, it is the second largest river crossing the Piedmont, exceeded only by the Susquehanna River. Figure 17 shows the extreme convexity of the profile, beginning in the Appalachian Valley.

The Tallapoosa River, like the Patapsco River in the Northeastern Highlands, is not an exterior stream. Its source lies within the crystalline rocks. Its profile is much like that of the Patapsco, and its power increases downstream toward the Fall Zone (fig. 18). A 32-km reach immediately above the Fall Zone has an average slope of 1.9 m/km or 0.0019. The gradient index (Hack, 1973a) is 490 gradient m (1,600 gradient ft), equivalent to that of many powerful streams in the Blue Ridge, such as the French Broad River below Asheville, N.C. Unfortunately, this steep lower reach of the Tallapoosa River is now drowned by the waters of Martin Lake so that it cannot be seen. However, at Martin Dam, the river was intrenched into the upland approximately 84 m before impoundment (U.S. Army Corps of Engineers data). Upstream from the lake, the average gradient in a long reach is 165 gradient meters (870 gradient ft). The river in this reach has an average discharge of 225 cms (2,426 cfs) and is a powerful river, though it flows through a topography of low relief.

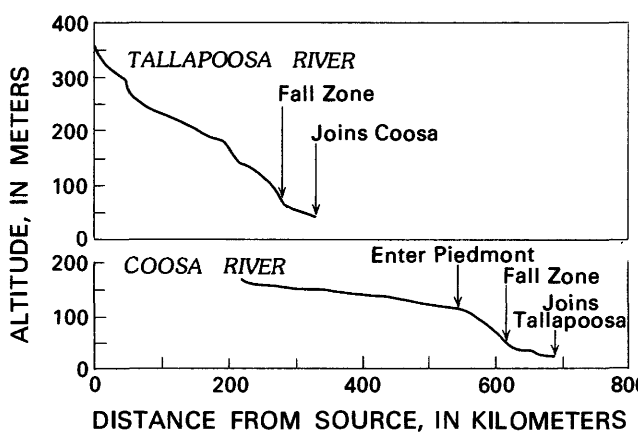


FIGURE 17.–Longitudinal profiles of the Tallapoosa and Coosa Rivers. Distances are measured from the heads of the streams.

The general configuration of the topography of the Southwestern Highlands of Alabama indicates a sharp uplift at the Fall Zone and a more gradual monoclinal rise from the Brevard zone northwest toward the Appalachian Valley (fig. 15). The topography in the Appalachian Valley is strongly controlled by the bedrock. Nevertheless, the resistant formations underlie plateaus and mountains 300 to 500 m in altitude, comparable with the general level of the Alabama part of the highlands, though not with the highest ridges such as Cheaha Mountain (730 m). In the extreme northeastern part of the highlands, the monoclinal rise to the Blue Ridge province is steeper.

FOOTHILL ZONE

The Foothill zone, as defined here, extends from the Potomac River southwestward through the inner Piedmont to near the South Carolina–North Carolina boundary. Its boundary on the northwest is defined by the southeast margin of the Blue Ridge physiographic province (Fenneman and Johnson, 1946). The southeastern boundary is defined approximately by the outer limit of hilly areas and isolated hills within the inner Piedmont and by a slight increase in slope (fig. 3).

The Foothill zone can be divided into three areas. The part north of the Roanoke River consists mostly of small ranges of hills and mountains underlain by resistant rocks and separated by extensive lowlands having much lower but varied relief. The Catoctin Formation, consisting of mafic lava flows, forms many ranges of high hills and must be one of the most resistant rock types, though the relief varies within local areas. The metasedimentary rocks of the Evington Group exposed in the long northwesterly reach of the James River contain resistant beds that underlie ranges of hills. These rocks are commonly quartzites of the Mount Athos Formation and of the Candler Formation, which also contains phyllite, muscovite schist, and locally marble (Espenshade, 1954, p. 15). Ranges of steep but low hills occur on the Lynchburg Formation; in Madison County, Va., hills are apparently on beds of conglomerate exposed in a syncline (Allen, 1963, map cross section A–A′). A broad belt of coarse-grained granitoid rocks of Proterozoic Y age forms the core of the Blue Ridge anticlinorium just southeast of the Blue Ridge Mountains. The rocks have considerable variety; some are sheared

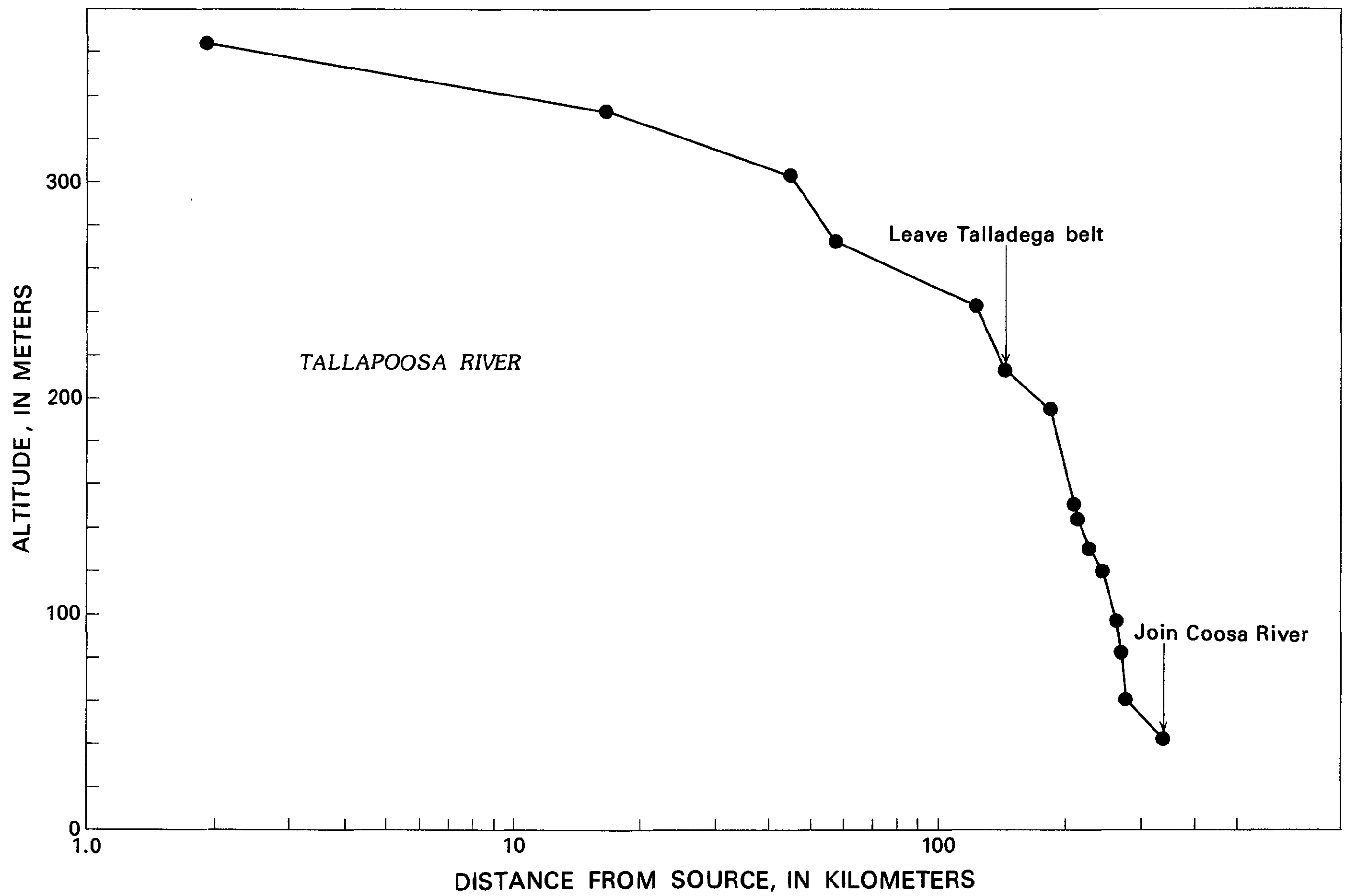


FIGURE 18.–Logarithmic profile of the Tallapoosa River.

and fragmented by cataclasis. Within this belt are many ranges of hills. The nature of the rock in places locally controls the topography. For example, Big Cobbler Mountain in the Orlean Quadrangle, Virginia, is within the local outcrop area of the magnetite-rich granite of the Robertson River Formation of Allen (1963) (James Clarke, oral commun., 1978). Thus, in this area, the effect of rock control is evident, and no proof exists that the local relief is related to uplift or tilting of the area. Nevertheless, the height of the hills is generally greater to the northwest close to the Blue Ridge Mountains.

Between the Roanoke River and the Yadkin River, the Foothill zone is more clearly defined and takes the form of an upland that has a relief uniformly greater than that of the Piedmont to the east. In much of this area, the zone is defined by two geologic provinces, the Smith River allochthon and the Sauratown Mountain window. Locally, quartzite forms mountains higher than their surroundings (for example, see Espenshade and others, 1975), but in general, the relief of the Foothill zone is markedly greater than that of the Piedmont Lowlands to the east.

Between the Yadkin River and the South Mountains, where the Foothill zone ends, the boundary is drawn across the lowlands of the inner Piedmont, that is, the lowlands between the Brevard zone and the Kings Mountain belt. The boundary shown on figure 1 follows a faint lineament that can be seen clearly on topographic maps that show contours only (U.S. Geological Survey maps of Winston-Salem, Charlotte, and Knoxville Quadrangles, scale 1:250,000). Whether this boundary itself has real meaning is uncertain, but it does serve as well as or better than an arbitrary boundary because most hilly areas of the inner Piedmont are northeast of the line. Figure 6 shows the contrast in relief. In this part of the Foothill zone, as is generally true, rock type does control the topography to some extent, although to a lesser degree than it does in the Virginia part of the Foothill zone. For example, the Brushy Mountains and South Mountains are both areas that have many peaks with altitudes of 750 to 900 m. These mountains are underlain by the more resistant rocks of the region, as shown on the geologic map of the Charlotte 2° Quadrangle (Goldsmith and others, 1978). The rocks include (1) a biotite gneiss unit containing hornblende gneiss, garnetiferous gneiss, and felsic volcanic rocks; (2) an augen gneiss, known as the Henderson Gneiss; (3) muscovite and sillimanite schist and minor quartzite and sandstone; and (4) a biotite gneiss unit that contains layered quartzofeldspathic gneiss, metasandstone, and siltstone.

Although the conspicuously high areas do contain what appear to be resistant units, all these resistant units also occur in low relief areas, where a variety of other rock units include granite gneiss, migmatite, hornblende gneiss with amphibolite, and schist. Even the sillimanite-rich rocks are not restricted to the mountains. This relation between topographic forms and rock types suggests that both tectonic forces and rock control have operated to produce the present forms, tectonic processes playing a major role. Figure 7 shows that the local contrasts in relief are impressive. The South Mountains rise 450 m above the Morganton Basin. The local relief within the mountains averages about 100 m, compared with less than 30 m in the basin. Many of these hills and small mountain ranges may be residuals formed by slope retreat as described by Kesel (1974).

Four large rivers cross the outer border of the Foothill zone. Three of them have a reach that has a steep gradient at the crossing. However, the steep reach on each river corresponds to what is probably a resistant rock type. The gradient index or logarithmic slope (see fig. 10 for definition) can be used to compare the profiles of rivers having different discharges or lengths. Gradient index values greater than 300 gradient m are characteristic of mountain streams, whereas values in the Piedmont are commonly less than 150 gradient m.

The James River has a gradient index of 365 gradient m in a reach 20 km long in the area crossed by the outer boundary of the Foothill zone. Downstream and upstream values average about 150 gradient m. The steep reach occurs in rocks of the Evington Group and in metavolcanic rocks of the James River syncline, both of which are probably resistant. The Roanoke River has a steep reach 15 km long just below the Danville Triassic and Jurassic basin, which forms the boundary of the Foothill zone. Values range from 275 to 560 gradient m. The rock units are the "Melrose granite" and "Shelton granite gneiss." Immediately below the steep reach, values drop to 75 gradient m.

In North Carolina, the Yadkin River crosses the boundary of the Foothill zone without a break in its profile. Twenty kilometers upstream is a short reach of 290 gradient m in a resistant plutonic rock, the Cranberry Gneiss (Espenshade and others, 1975). No reason exists to believe that the Foothill zone boundary has any effect on the profile of this river. The fourth river, the Catawba, crosses the Foothill zone boundary at the Lookout Shoals Reservoir site. The gradient index of the channel here is 300 gradient m, much higher than values for some distance above or below. The rock is mapped as part of a large area of biotite schist, which would not be expected to steepen the profile.

The evidence for an active uplift localized along the boundary of the Foothill zone is not clearly supported by the stream-profile data, at least on the large streams. The Yadkin River shows no steep reach along the line. The Roanoke has a steep reach a few miles downstream at the border of a Triassic and Jurassic basin. The other two rivers, the James and the Catawba, do have steep reaches on the boundary, but the reach on the James could be related to rock resistance rather than to a tectonic break. These facts may mean either that the outer boundary of the Foothill zone is a broad zone of uplift that is not faulted at its border or that the large river profiles have become adjusted through time to the resistance of the rocks that they cross. Nevertheless, the Foothill zone must be an area that has at some time been tilted upward, and even though many of the hill areas correspond to resistant rock, the hills are generally higher near the Blue Ridge Mountains than they are farther out in the Piedmont.

NORTHERN BLUE RIDGE MOUNTAINS

The term "Blue Ridge" has acquired several meanings. As a geographic feature recognized by nongeologists, it is simply the range that forms the southeastern border of the Appalachian Mountains between northern Georgia and Pennsylvania. South of Roanoke, it includes only the escarpment and mountains behind it that rise directly above the Piedmont. This common definition excludes, for example, the Great Smoky Mountains as well as some other ranges southeast of the Appalachian Valley. Physiographers have long used the term "Blue Ridge" as a name for a physiographic province. The boundaries are shown on the map of the "Physical Divisions of the United States," published by the U.S. Geological Survey (Fenneman and Johnson, 1946). The province includes all the high mountains underlain by crystalline rocks from northwest Georgia to Pennsylvania. Structural geologists now use the term "Blue Ridge belt" to describe a geologic province (King, 1955, p. 358). In their definition, the Blue Ridge belt is an anticlinorium cored typically by crystalline rocks of Precambrian age and flanked by stratified rock of probable Precambrian and Early Cambrian age. It includes all the area northwest of the Brevard zone in Alabama and Georgia. It thus includes the Southwestern Highlands of this report, an area that is part of the Piedmont of the physiographers. The geologic definition broadens the Blue Ridge in northern Virginia to include all the rocks northwest of the Culpeper Triassic and Jurassic basin, that is, the Blue Ridge–Catoctin Mountain anticlinorium. However, from the north end of the Brevard zone near Winston-Salem to Charlottesville, Va., the southeastern boundary is not clearly defined.

In this paper, I have followed the usage of the physiographers but have divided the Blue Ridge physiographic province of Fenneman and Johnson into two parts. The part north of the Roanoke River, which is essentially a narrow range of high mountains, is referred to as the Northern Blue Ridge Mountains. The area south of Roanoke is referred to as the Southern Blue Ridge province. I have separated the two because I believe that the origins of the topography are not the same.

The Northern Blue Ridge Mountains are essentially the frontal mountain range of the Valley and Ridge province. They are underlain by a sequence of Precambrian and Cambrian rocks that form the northwest limb of the Blue Ridge anticlinorium of the geologists (Espenshade, 1970). The sequence is composed of rocks that are resistant to erosion.

Only three rock units support the crest of the range along its entire length, a distance of 395 km (fig. 19): (1) a complex assemblage of medium to coarse granitoid rocks, some of which contain large crystals of potassic feldspar and include hypersthene granodiorite; (2) a volcanic unit known as the Catoctin Formation; and (3) a thick series of quartzite, arkose, and phyllite called the Chilhowee Group that forms the base of the Paleozoic sedimentary column exposed in the Appalachian Valley.

The hypersthene granodiorite unit extends most of the length of the Northern Blue Ridge Mountains and is the northwestern rock unit within the large area of Precambrian granitoid rocks that make up the core of the Blue Ridge anticlinorium and that are referred to as the Virginia Blue Ridge Complex of Brown (1958). The hypersthene granodiorite is medium to coarse grained and massive. It contains andesine or calcic oligoclase, potassic feldspar, pyroxene (both hypersthene and augite), biotite, and blue quartz. In many places, the feldspar crystals are more than 8 cm long. In places, the hypersthene granodiorite is altered to unakite, a rock composed of feldspar, epidote, and blue quartz. The two types grade into each other (Jonas, 1935). As the hypersthene granodiorite is locally variable in texture and composition, it has been commonly referred to as the "Pedlar Formation" (Bloomer and Werner, 1955; Allen, 1963). The "Pedlar Formation" is generally more resistant than other rock units in Brown's Virginia Blue

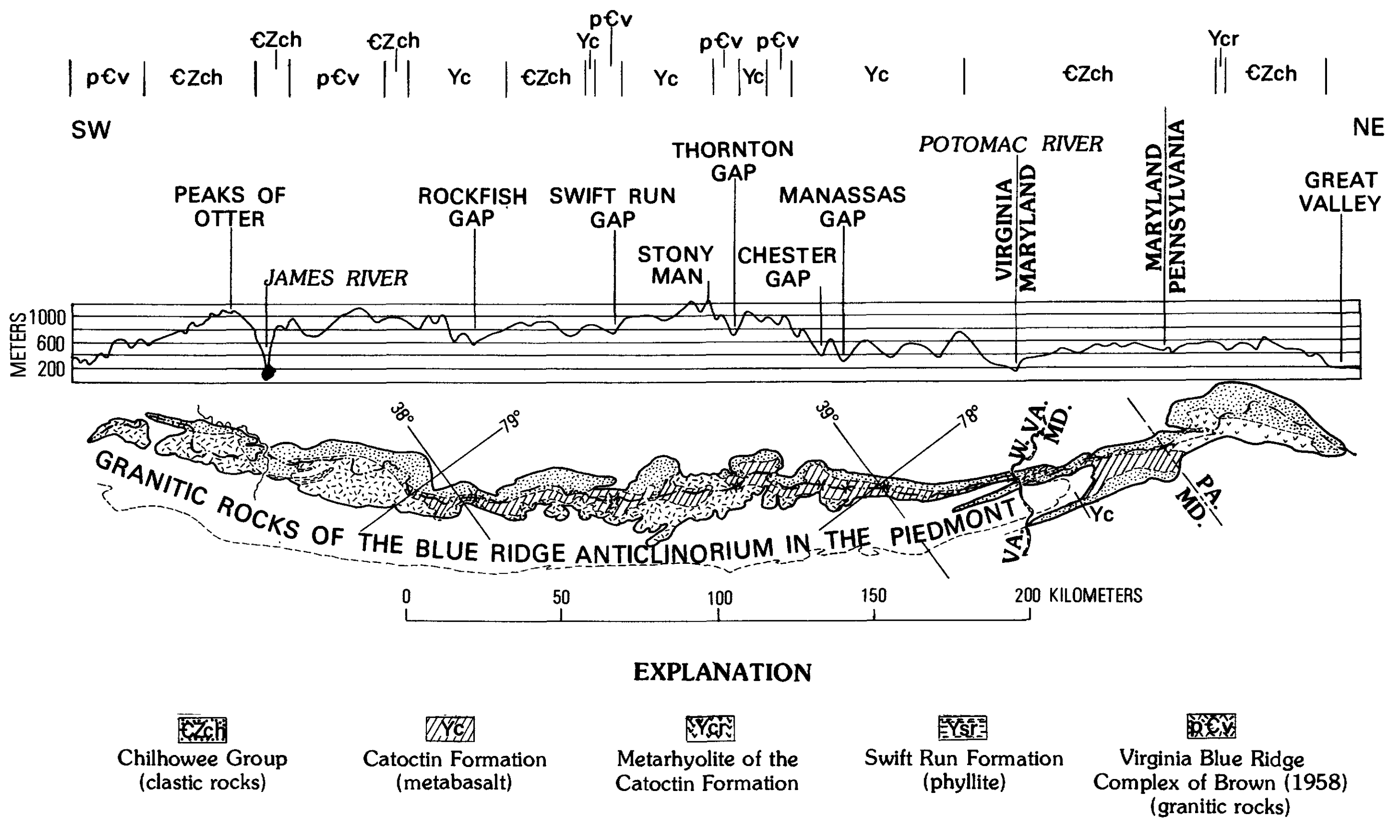


FIGURE 19.—Profile along the crest, and geologic map, of the Northern Blue Ridge from the Roanoke River to southern Pennsylvania. On the geologic map, the crest of the range is shown by a line of dashes and dots.

Ridge Complex. This resistance can be demonstrated in the area mapped by Bloomer and Werner (1955), which includes 1,600 km^2 in the Blue Ridge and Foothill zone between the James River and Rockfish Gap (fig. 17). Comparison of the mapped geologic units with topographic maps shows that the "Pedlar Formation" occupies the highest summits in both the Blue Ridge and the Foothill zone. The highest peak in the area is the Priest (1,236 m), about 11 km southeast of the range crest. In the Foothill zone, these rocks underlie the Tobacco Row Mountains that contain a peak 891 m in altitude. Other rocks of Brown's Virginia Blue Ridge Complex in the Foothill zone occupy areas of lower altitude and relief. The relief of the Lovingston Granite Gneiss, for example, is somewhat less than that of the Tobacco Row Mountains. The most extensive rock unit, the "Marshall Formation," is a medium-grained granitoid rock that occupies valleys and low hilly areas. The Roseland anorthosite body occupies a pedimentlike area of about 17 km^2 in the Foothill zone that has a total relief of only 75 mm. Figure 19 shows that the "Pedlar Formation" underlies some of the highest areas along the Blue Ridge crest.

The Catoctin Formation, another unit having considerable resistance to erosion, contains a variety of rock types but consists mostly of massive green metamorphosed basalt composed of albite, chlorite, epidote, and actinolite. Agglomerate and other pyroclastic rocks occur between the lava flows. The thickness of the formation is variable and reaches 550 m south of Thornton Gap (Reed, 1955). In this area, the Catoctin lavas are massive and have many features that are seen in recent lava flows, such as columnar jointing and amygdules filled with secondary minerals. Seven lava flows, having a total thickness of 460 m, can be identified. The massive character of these rocks and the gentle dips to the east account for the fact that this area is the highest part of the Northern Blue Ridge Mountains (fig. 19.) The Catoctin Formation is not everywhere so massive and in places is sheared and schistose. Nevertheless, as the differential topographic relief in the Foothill zone shows, it is a resistant rock and a ridge maker along much of its outcrop. For most of the extent of the Blue Ridge anticlinorium north of the Potomac River, the Catoctin Formation forms the core, flanked on both sides by the Chilhowee Group. In Pennsylvania, the principal rock type is metarhyolite. In that region, however, the divide that forms the crest of the range follows the Catoctin Formation for only a short distance (fig. 19).

The Chilhowee Group is a clastic sequence of Proterozoic Z and Early Cambrian age. The basal part is generally a metamorphosed sandstone known in the north as the Weverton Formation, and in the south as the Unicoi Formation. In some places, a thin red slate (former Loudoun Formation, now abandoned and replaced by Catoctin) forms the base of the group. The Weverton or Unicoi, however, is 150 to 500 m thick and is mostly arkose but also contains shale and quartzite. It is overlain by the Harpers Formation (Hampton Formation in the south), a sequence of metamorphosed shale and siltstone and minor sandstone and quartzite; thickness ranges from 300 to 600 m. The topmost formation in the group is the Antietam Formation (Erwin Formation in the south), composed of siliceous sandstone and vitreous quartzite, 125 to 250 m thick. These formations generally dip steeply to the northwest and form the northwest flank of the Northern Blue Ridge Mountains. The Weverton Formation underlies the crest of the range at many places for about 45 percent of the crest length, a combined distance of 176 km. Over most of this distance the divide is lower than it is on the other rocks. The entire Chilhowee Group forms a wide belt of mountains on the northwest margin of the Blue Ridge that is generally lower in altitude than the crest of the range and that is also lower than many hills southeast of the crest. Thus, rocks of the Chilhowee Group, including the Weverton Formation, are probably somewhat less resistant than those of the Catoctin and "Pedlar" Formations.

Rock control does not explain the entire profile of the range. Fiqure 19 shows that from Chester Gap to the northern end in Pennsylvania, the range is lower than it is to the south. North of the Potomac River, the range does not exceed 620 m in altitude, and its average altitude is little more than half that of the southeastern part. On the other hand, the outcrop of resistant rocks in places is as wide or wider than it is to the southeast. This difference is perhaps best explained as the result of tectonic causes.

The Northern Blue Ridge Mountains appear to be part of an erosional system in an uplifted area that includes the mountain ranges of the Valley and Ridge province that rise to approximately the same altitudes. The forms and relief in the Valley and Ridge province are controlled largely by rock resistance and width of outcrop. The decline in altitude north of the Potomac River may be due simply to a lesser vertical component of uplift in that region either currently or at some time in the past.

SOUTHERN BLUE RIDGE PROVINCE

The Southern Blue Ridge province extends from the Roanoke River southwestward into north Georgia, where the mountain chain on the crystalline rocks is interrupted by lowlands along the lower Etowah River. As defined here (fig. 1), the Blue Ridge province ends at Pine Log Mountain north of Cartersville, Ga. The northwest border of the Southern Blue Ridge province follows the outcrop of the uppermost quartzite beds in the Chilhowee Group of Early Cambrian age. The southeast border is an erosional escarpment related only locally to the underlying rocks. From the Georgia–South Carolina border northward, the crest of the escarpment forms the drainage divide between streams draining to the Atlantic Ocean and those draining to the Gulf of Mexico. Viewed as a whole, the escarpment cuts diagonally across the entire Blue Ridge anticlinorium.

The Southern Blue Ridge province can be divided into several areas on the basis of topography. Some of these areas correspond to distinctive geologic areas, but some do not (fig. 20). Rock control, however, is responsible for many landscape features.

CHILHOWEE-WALDEN CREEK BELT

The landforms on the northwestern margin of the province are closely related to the geologic pattern. They form a zone of elongate mountain ridges and valleys 5 to 50 km wide, underlain by sedimentary and metasedimentary rocks of low metamorphic grade. Contrasts in rock resistance are great, as the rocks range in composition from carbonate and shale to massive quartzite. Long, steep-sided ridges separated by parallel valleys are characteristic of the belt. Northeast of the French Broad River, the quartzite of the Chilhowee Group forms most of the ranges, whereas the valleys are underlain by infolded or faulted carbonate and shale of Cambrian age typical of the Appalachian Valley, such as the Shady and Honaker Dolomites and the Rome Formation. Southwest of the French Broad River, metasedimentary rocks of Proterozoic Z age, referred to as the Walden Creek Group, are involved, forming a belt of intermediate relief between the Chilhowee ridges and the mountain highlands (Swingle and others, 1966; Hadley and Nelson, 1971).

The mountains in the Chilhowee–Walden Creek belt are not as high as many of the peaks of the mountain highlands. Nevertheless, some ridges underlain by quartzite of the Chilhowee Group exceed 1,200 m in altitude, especially in the central part of the area. Local relief is also high, in places more than 750 m. The many longitudinal valleys, especially where they are underlain by carbonate rocks, are bordered by large fans and floored by residuum.

MOUNTAIN HIGHLANDS

The largest part of the Southern Blue Ridge province consists of ranges of high mountains, many of which exceed 1,500 m in altitude. Mount Mitchell, the highest mountain in the Eastern United States, is 2,040 m in altitude. Local relief is generally high, exceeding 900 m. Along the northwest and southeast borders of the Mountain highlands the ranges tend to be roughly parallel to the northeast regional strike of the rocks. In the center, however, the ranges have various trends that seem to be determined by the complex pattern of deep basins and valleys that cut through the mountains in different directions.

Some rock control of topography is evident. For example, the Great Smoky Mountains have many peaks higher than 1,800 m. The highest part of the range is underlain by massive conglomerate and sandstone of the Thunderhead Sandstone. The trends of many valleys in the Mountain highlands parallel the strike of the rocks, and some deep basins are related to large outcrop areas of readily weathered rock types, for example, the low areas of the Spruce Pine district in which feldspar-rich rocks are extensive. On the other hand, the general im-

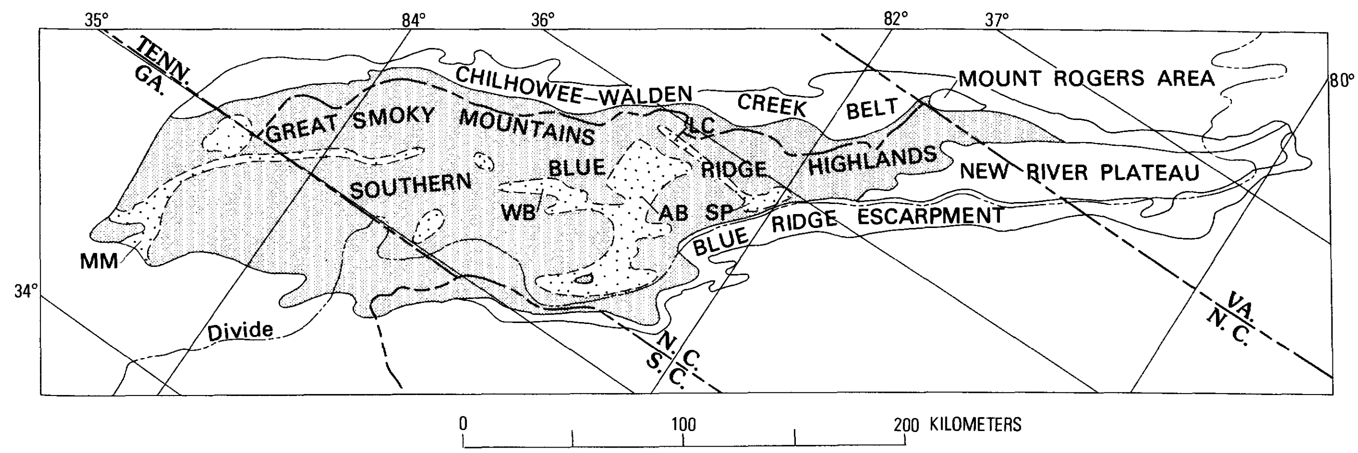


FIGURE 20.–Map of the Southern Blue Ridge province, showing the subdivisions discussed in the text. Ruled line pattern represents high-relief areas in the Southern Blue Ridge province. Dotted areas are the principal basin and trench areas: AB, Asheville basin; SP, Spruce Pine area; WB, Waynesville-Canton basin; MM, Murphy Marble belt; LC, Laurel Creek trench.

pression given by the region does not suggest rock control like that seen in the Ridge and Valley province or in the Chilhowee–Walden Creek belt.

The intermontane basins are irregular, have angular boundaries, and are often connected with other basins by deep valleys that look like trenches. This appearance is due to the fact that the valleys commonly are straight and cut across drainage divides at high angles. They are referred to herein as trench valleys. The intersecting pattern of linear valleys and interconnected basins suggests an origin due in part to control by nonresistant rocks and in part to tectonic influences. The tectonic influences could have involved faulting in the late geologic time or, more likely, differential erosion along brittle fracture zones associated with older faults.

Several trenches and basins show interesting correlations with other features that shed light on their origin. The most impressive of these is the long deep trench that corresponds to the outcrop area of the Murphy Marble and Andrews Schist of Cambrian(?) age; these formations are the youngest in a great syncline folded into the rocks of the southwestern part of the Mountain highlands (MM, fig. 20). The trench is approximately 150 km long and extends north through Georgia into North Carolina. The beds that floor the trench in the North Carolina part are marble, phyllite, and mica schist (Hadley and Nelson, 1971; Kish and others, 1975). In Georgia, where the marble occurs only locally along the belt, the relief is fairly high. In North Carolina, however, the marble has a continuous outcrop and forms a spectacular trench far below the mountains on either side. The trench culminates in the Nantahala Gorge, whose floor is 450 to 900 m lower than the mountaintops on either side (fig. 21), and is in places less than half a kilometer wide. The gorge ends at the termination of the marble outcrops.

At no place does the trench contain more than a headwater stream, although it crosses the large Hiwassee River and its northern end is near the Little Tennessee River. The trench cuts directly across the main westward-flowing drainage system with little change in altitude. This trench is significant because it demonstrates conclusively that where contrasting rock types occur, rock control of topography is important in the Blue Ridge. The immediate local relief along the trench matches in magnitude the relief in the Appalachian Valley, where carbonate rocks are adjacent to wide outcrop belts of rocks such as quartzite and massive sandstone. The morphology of the trench on the Murphy Marble is well shown on the Knoxville, Chattanooga, and Rome sheets of the 1:250,000-scale topographic-map series distributed by the U.S. Geological Survey.

Another spectacular trench is north of Asheville, N.C., and trends approximately east (LC, fig. 20). It is referred to in this report as the Laurel Creek trench after a stream that occupies the trench near its western end. The profile of part of this trench is shown in figure 21. The trench crosses several drainage divides. The Cane River is a large stream that in crossing the trench, occupies it for about 14 km. The South Toe River also crosses the trench, occupying it for about 2.5 km. The Laurel Creek trench is about 750 m lower than the mountain ridges north of it (fig. 21). The trench crosses the Spruce Pine basin at its eastern end and terminates against the Grandfather Mountain window. The geologic map of the Knoxville Quadrangle (Hadley and Nelson, 1971) shows that the trench is occupied for much of its length by a right-lateral fault having a horizontal displacement of about 2.5 km near the western end. The Laurel Creek trench crosses the strike of the rocks at a 50° angle, so that it is not related to weak-rock strata, as is the trench on the Murphy Marble. On the other hand, the presence of the fault suggests that the trench may owe its origin to a shear zone that has the same erosional effect as a narrow belt of weak rock. A sheared zone, if it contains fractures that allow the passage of ground water, would be a zone of low resistance to weathering and erosion.

The outlines of intermontane trenches, valleys, and basins can be seen clearly on map separates showing the topography only, at 1:250,000 scale. On such maps, the differences in topographic relief between mountains and low-relief areas are shown by differences in spacing of contours, and areas of low relief are much lighter in tone. Because the mountainous relief is generally quite high, as much as 1,000 m within a 100-km^2 area, the contrast with basins having relief of less than about 200 m is great. Figure 22 is an outline map of part of the Mountain highlands made by tracing the boundaries between sparse and dense contours. The map also shows the Brevard zone (Hadley and Nelson, 1971; Goldsmith, Milton, and Wilson, 1978; Nelson and Clarke, 1978) and the outcrop of one map unit referred to as gneiss and migmatite on the map of Hadley and Nelson (1971). This unit seems to be a preferred rock type for the formation of intermontane basins, although it is not known to be particularly nonresistant. It is outlined here simply to show the general structural trend of the rocks in a central part of the area. Several large trench valleys that crosscut the structural pattern of the strata are clearly shown. A long trench valley shown in part on figure 22 contains a fracture zone known as the War Woman fracture (or shear) zone (Pickering and Higgins, 1979; Georgia Geological Survey, 1976). Some other trench valleys are identified by names coined by the writer.

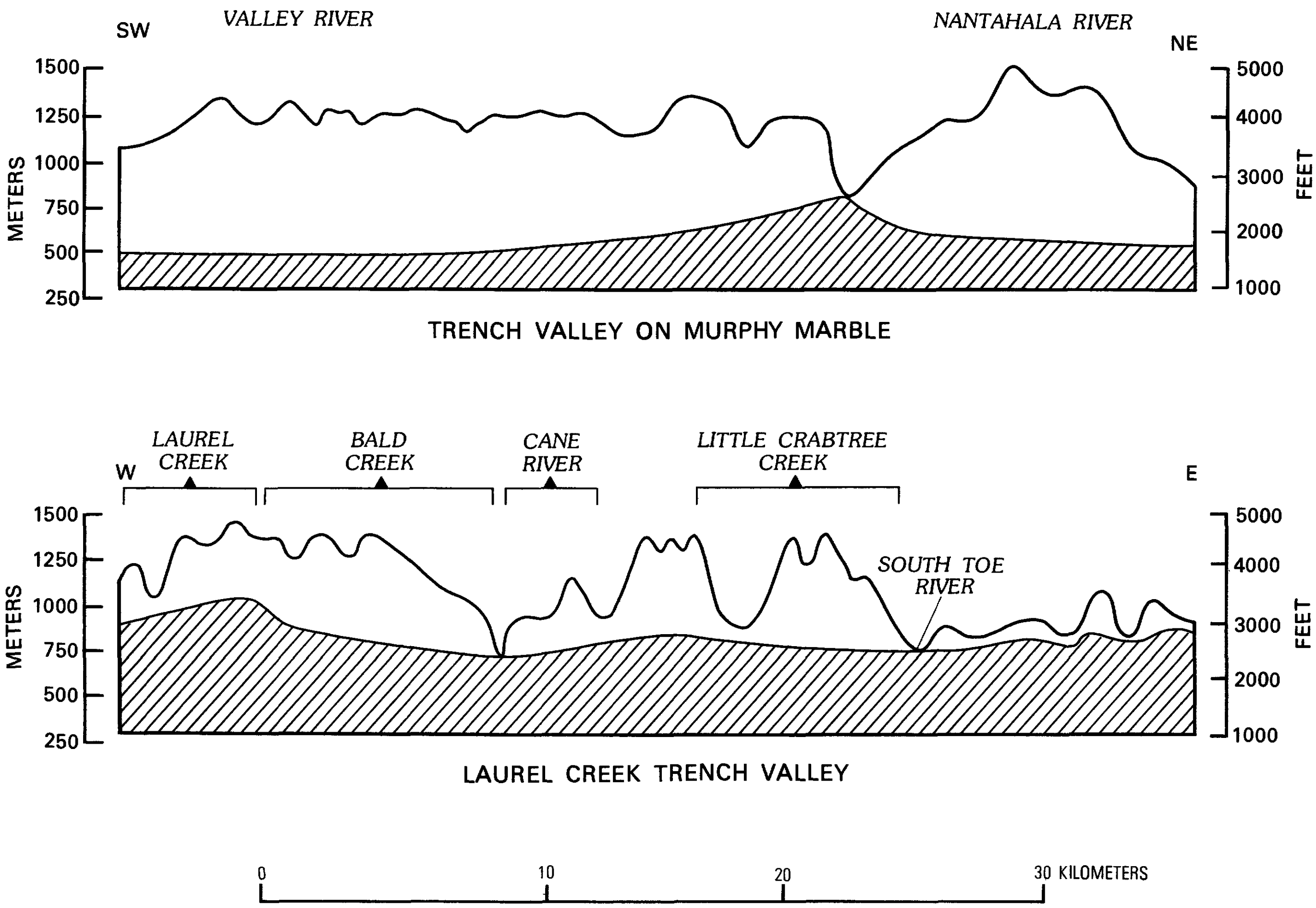


FIGURE 21.–Profiles along two trench valleys in the Mountain highlands. The wavy heavy lines represent the profile along ridge crests immediately on the north or northwest flanks of the valleys. The lines on the shaded areas represent the profiles of the valley floors.

Silicified breccia zones mapped by Hadley and Nelson (1971) are shown by lines of triangles.

As the map shows, many valleys have distinct linear trends. Some valleys clearly parallel the strike of the rocks or the trends of thrust faults that parallel the strike. Examples are the trench valley that trends southwest from Waynesville (W, fig. 22) to the Tuckasegee trench and, of course, the Brevard zone. Other valleys clearly crosscut the structural grain, as do major basins such as the Little Tennessee and the Asheville basins.

Acker and Hatcher (1970) have identified rectilinear systems of valleys controlled by joint systems at right angles to the structural grain. Spectacular valleys as deep as 100 m are found in the drainage areas of the Chauga, Tugaloo, and Chattooga Rivers in northeast Georgia and northwest South Carolina. The spacing of the valleys is controlled by the drainage pattern, but the flow directions and linearity are commonly controlled by joint directions. Some of the examples illustrated resemble the trench valleys and can clearly be seen on topographic maps at 1:250,000 scale.

BREVARD ZONE AND HENDERSONVILLE BULGE

The Brevard zone is a broad zone of faults and fault slices that extends from Virginia to Alabama. It has had a long history involving several periods of activity possibly extending as far back as the Precambrian (Rankin, 1975). The youngest tectonic features exposed at the surface are brittle-fracture phenomena (Hatcher, 1978). The Brevard zone is itself a major topographic feature generally forming a narrow zone of low relief. At its northern end near Winston-Salem, N.C., the fault zone is about 35 km south of the Blue Ridge escarpment. In the Morganton basin, it is at the foot of the Blue Ridge escarpment. For about 122 km southeast, it rises to cross the Mountain highlands for a distance of about 120 km. In this region, it forms a prominent topographic boundary. The mountain region southeast of the Brevard zone is distinctly lower than the area northwest

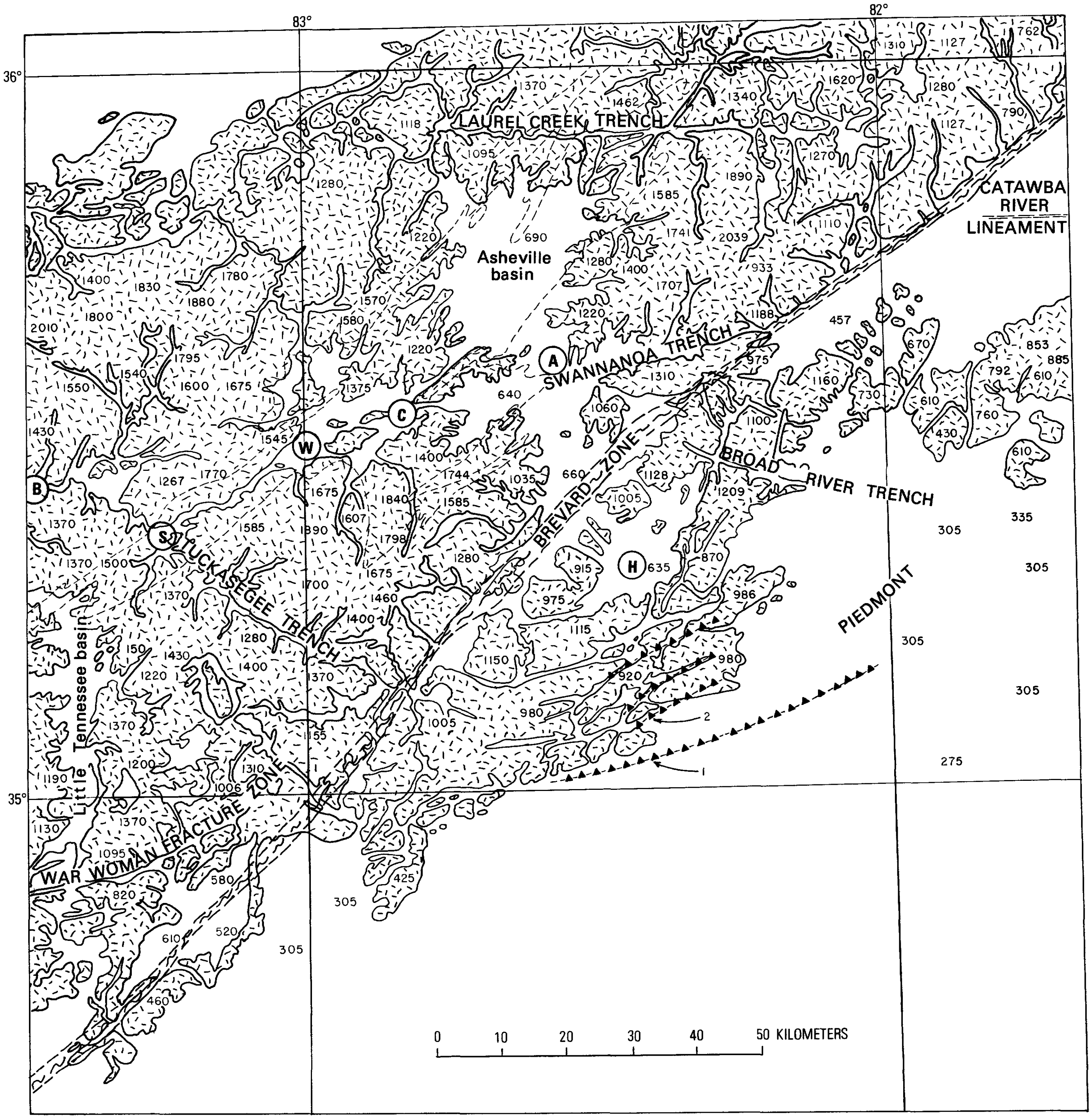


FIGURE 22.–Map of a part of the Mountain highlands, showing the larger trench valleys and intermontane basins. Patterned areas are rugged terrain having generally high relief. Unpatterned areas are low-relief areas. Lines of triangles represent silicified breccia zones (from Hadley and Nelson, 1971). Fine dashed line represents boundary of gneiss and migmatite unit of Hadley and Nelson (1971). Letters in circles represent Asheville (A), Canton (C), Hendersonville (H), Waynesville (W), Sylva (S), and Bryson City (B).

of it and has a different structural grain (fig. 22). The mountains and trench valleys within this lower region are arranged in linear and arcuate patterns. The arcuate features are related to the strike of the underlying rocks. The area is like a huge berm, bulging outward into the Piedmont from the main Mountain highlands. For convenience of discussion, it is here called the "Hendersonville bulge."

The topographic evolution of the Hendersonville bulge is probably related in part to the drainage system and in part to rock resistance. The drainage of the bulge is mostly northwest of the eastern continental drainage divide to the Tennessee River via the French Broad River. The fact that its general level is higher than that of the Piedmont or southeast drainage probably explains the difference in altitude between the bulge and the Piedmont (Hack, 1973b).

However, the difference in relief between the Hendersonville bulge and the Mountain highlands on the northwest side of the Brevard zone must have another cause. The rocks of the bulge are considered part of the Piedmont province by geologists. The most extensive formation is the Henderson Gneiss, described by Hadley and Nelson (1971) as a medium- to coarse-grained foliated biotite-microcline augen gneiss. The degree of foliation and shear increases and becomes intense near the Brevard zone. In my judgement, such a rock might be expected to be much less resistant than many of the rocks in the Great Smoky Group to the northwest. Thus, the large outcrop area of Henderson gneiss may explain wholly, or in part, the difference in relief.

The middle reaches of the French Broad River (fig. 23) are in the Henderson Gneiss just southeast of the Brevard zone. The river has a low gradient in this area (0.00053 over a distance of 34 km). Where the river leaves the Brevard zone to enter the Asheville basin, the slope increases to 0.00097 for 19 km. The next reach downstream passes through Asheville and has a slope of 0.0013. The gradient continues to increase to a maximum of 0.005 in the Longarm quartzite. This kind of convex profile is possible in a system that has achieved a state close to dynamic equilibrium, where the rocks become increasingly more resistant downstream. On the other hand, an increased rate of uplift northwest of the Brevard zone, relative to the Hendersonville bulge, could also result in a strongly convex-upward stream profile if the stream were uplifted relative to the Appalachian Valley. Thus, the explanation for the different relief in the two areas may involve one or both of two causes.

Several lineaments and conspicuous trench valleys that may be structurally related to the Brevard zone trend away from it at low angles. The Catawba River lineament in the Piedmont is one of these features. This lineament is followed by the river for a distance of 50 km. Only the western end is shown in figure 22. Surface geologic mapping has disclosed no fault (Bryant and Reed, 1970; Goldsmith, Milton, and Wilson, 1978). The Laurel Creek trench west of the Brevard zone trends east, in the same direction as the Catawba River lineament. This feature, as mentioned above, does contain a fault having an offset of 2.5 km. The Swannanoa trench intersects the Brevard zone at an angle of about 20°. No offset along this zone has been identified, but the trench is a prominent lineament that connects the Morganton, Asheville, and Canton-Waynesboro basins. It may continue westward as the Bryson City trench.

The Tuckasegee trench trends northwestward from the Brevard zone. This crosscutting trench is noteworthy because of its straightness and the fact that it separates mountain masses having different general altitudes. The mountain peaks south of the trench are about 300 m lower than those north of it.

The silicified breccia zones in the southeastern part of the area shown in figure 22 are of interest because they show that zones of brittle fracture have formed that do not displace the geologic contacts to a large degree. The breccia zone at locality 1 (fig. 22) is crossed by a Triassic dike that is not displaced within the limits of mapping (Hadley and Nelson, 1971). However, Snipes (1979) reported that the breccia zone at locality 2 offsets a vertical Mesozoic dike at least 175 m. The amount of vertical offset, if any, cannot be determined. This evidence indicates that fracture zones do exist in the area that involved offsets in post-Paleozoic time. Silicification of the breccia zones in the Piedmont is evidence that solutions have percolated through them. Brittle fracture zones may have localized rapid erosion in many places in the Mountain highlands and were perhaps the loci of small differential movements in post-Paleozoic time that affected the gross morphology of the terrain.

INTERMONTANE BASINS

The origin of the basins within the Mountain highlands is another difficult problem. Most geomorphologists who have worked in the area have considered these basins to be straths formed when the rivers that drained them were flowing at a lower grade, generally correlative with the Harrisburg surface of the north (see Thornbury, 1965, p. 107). Wright (1931) thought that rock resistance had a strong influence on the position and altitude of these basins. The idea that a sill of resistant rock slowed relative erosion rates along the streams so that a low-relief area could migrate up the tributaries is appealing, and some such mechanism may have been a part of the process of basin formation. On the other hand, the basins themselves have patterns and outlines

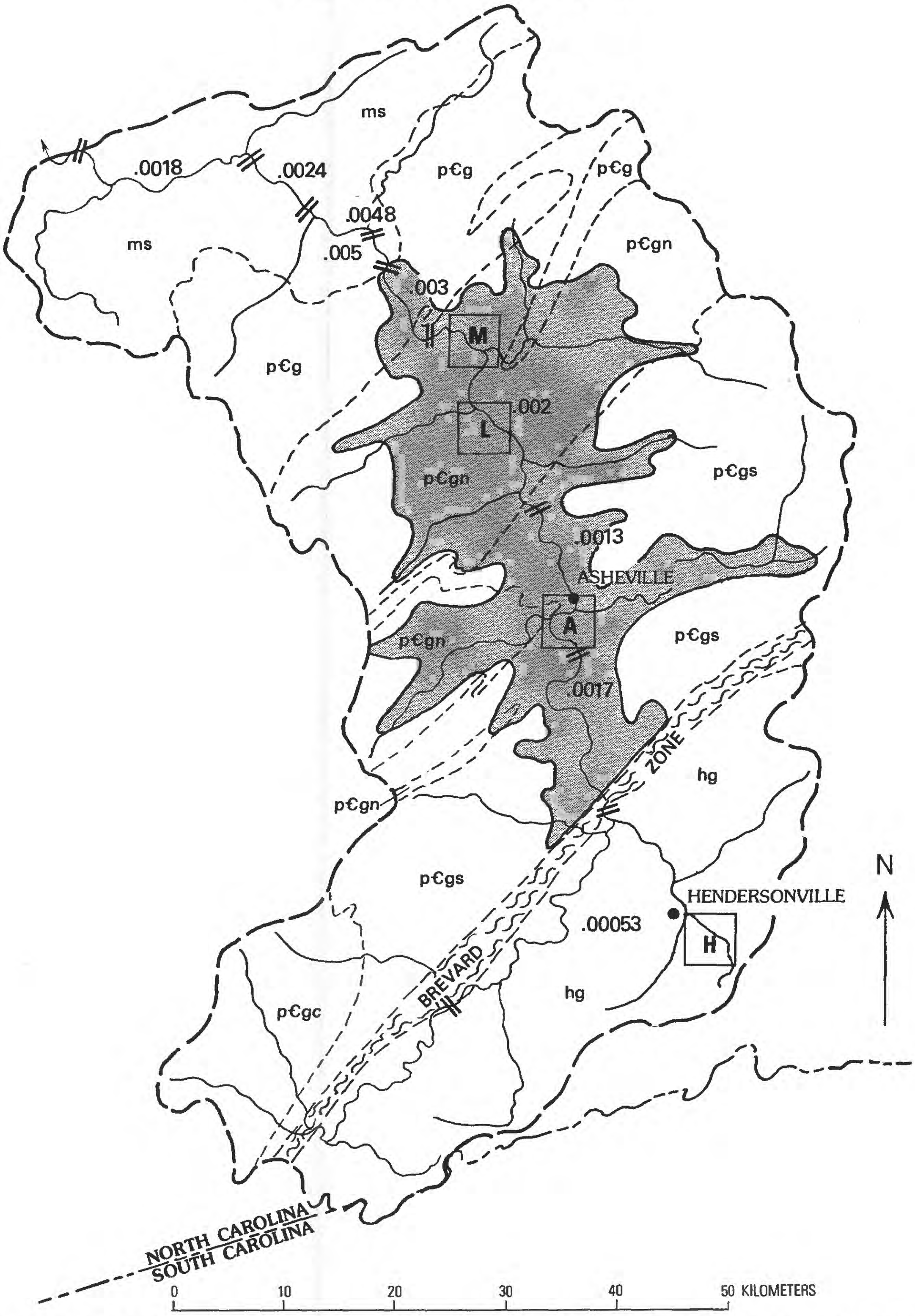


FIGURE 23. – Geologic map of the basin of the French Broad River (Asheville basin shaded). Numbers represent gradients of reaches of the French Broad River between double bars. Letters in squares are the first letters of the quadrangle names listed in table 2: H, Henderson; A, Asheville; L, Leicester; M, Marshall. Geologic units shown are: ms, metasedimentary rocks of Proterozoic and Cambrian age that include quartzite, sandstone, graywacke, phyllite, and shale; p€g, Max Patch Granite and Cranberry Gneiss and migmatite; p€gn, layered gneiss and migmatite; p€gs, Great Smoky Group; hg, Henderson Gneiss and related rocks; p€gc, biotite schist and gneiss. (From Hadley and Nelson, 1971.)

related to the rock structure and to the trench valleys that suggest a more complex origin. Several examples can be cited. The complex basin at Waynesboro and Canton (W and C, respectively, fig. 22) is at the intersection of at least two trench valleys and within the outcrop area of the layered gneiss and migmatite, which is a rock unit quite variable in composition. The broadest part of the Asheville basin is within the outcrop area of the same gneiss and migmatite unit. The complex basin at Spruce Pine (SP, fig. 20) is on highly feldspathic rocks as well as a fault zone. The Henderson basin (H, fig. 22) and the low-relief area to the west along the Brevard zone are in the Henderson Gneiss, a highly foliated rock. Nevertheless, the evidence is not sufficient to indicate that rock control by itself is responsible for all the basins.

The Asheville basin is the largest (fig. 23), and the problem of its origin is probably typical of other basins. It forms a large low-relief area having an average altitude of about 600 m. The floor of this area is almost the same altitude. Table 2 shows hypsometric data for four rectangular areas in this basin, within the Hendersonville, Asheville, Leicester, and Marshall Quadrangles (fig. 23). As the table shows, the altitudes of the highest points in these areas do not decline systematically downstream. The highest point of all is in the Marshall Quadrangle, the farthest downstream. However, the mean altitude declines 41 m, whereas the river declines 128 m, reflecting the systematic increase in total and mean relief.

The elevation-relief ratio is a measure of the symmetry of the topography. It is the ratio of the mean relief to the total relief. It expresses the proportionate amount of the land surface lower than the mean relief and thus reflects the forms. As the ratio in this area is close to 0.5 and does not change systematically downstream, there is no indication that the landscape has relict features caused by dissection of an older surface. The data seem to indicate only that the relief increases downstream and that the local topography is symmetrical and adjusted to the present river. The increase in relief may be due to the overall increase in rock resistance downstream. The possiblity exists, however, that tectonic uplift or tilting could produce the same effect.

The question of origin still remains. The relief in the Asheville basin is significantly lower than that of the mountains around it, and the broader outlines of the area, at least, are not controlled entirely by rock types. A possible explanation is that the area is affected by intensive faulting and shearing along many zones. The absolute amount of movement in these zones may have been small, but the zones provided a network of fractures along which erosion could proceed. The altitudes of the floor of the basin might have been controlled by a sill of massive and resistant rocks that crosses the French Broad River at the lower end of the basin. Possibly the course of the French Broad itself was determined by a tectonic line or zone of weakness transverse to the structure. Also belts and areas of weak rocks may occur in this and other basins that cannot be identified in the available reconnaissance maps. Detailed fieldwork may be required to correlate the bedrock patterns with topographic patterns in order to resolve this problem.

MOUNT ROGERS AREA

Mount Rogers (1,750 m) and White Top Mountain (1,630 m) just north of the North Carolina–Virginia boundary rise conspicuously higher than their surroundings. They are underlain by porphyritic welded rhyolite tuff unique to this area. The surrounding lowlands are underlain by rocks of the same formation (Mount Rogers Formation) but consist of metasedimentary rocks that include tillite, graywacke, siltstone, shale, and tuff (Rankin and others, 1972). North of this local area, the Iron Mountains, underlain by the Chilhowee Group, rise to a lesser height (1,410 m). Thus, in the Mount Rogers area, rock resistance is closely correlated with differences in relief.

NEW RIVER PLATEAU

The northeastern part of the Southern Blue Ridge province is a plateau region having comparatively low relief, relieved in a few places by low mountains. Most of the area is graded to the New River, which flows out of

TABLE 2.—*Comparison of topography at four locations in the Asheville basin*

[Based on altitudes of 64 points in a rectangular grid 20 km² in area. At all three localities, the grid is centered on or close to the channel of the main stream. Quadrangles at 1:24,000 scale; 20-ft contours]

Quadrangle	Highest point (m)	Lowest point (river surface) (m)	Total relief[1] (m)	Mean relief (m)	Mean altitude (m)	Elevation-relief ratio[2]
Hendersonville	664	633.5	30.5	12.8	646	0.42
Asheville	670	600	70	30.5	631	.43
Leicester	670	524	146	84.0	608	.57
Marshall	701	505	195	100	605	.51

[1] Difference in altitude between highest and lowest points.
[2] Ratio of mean relief to total relief; essentially the same as the hypsometric integral (see Pike and Wilson, 1971).

the area and into the Appalachian Valley at an altitude of 610 m. The New River plateau is mostly less than 900 m in altitude. Where there are no mountains, the average relief in a 100-km^2 area is about 200 m. In such places, the relief of the area is higher than that of the Piedmont but much less than that of other parts of the Blue Ridge. The streams cross the New River plateau in sinuous courses, and form exaggerated meanders that have long straight reaches trending northwest at right angles to the strike of the bedrock, much like the meanders that are common in Martinsburg shale in the Appalachian Valley (Hack and Young, 1959).

The New River plateau as a whole appears to owe its distinct character largely to rock control. Most of the area is underlain by thinly layered schist and gneiss of the Ashe Formation. The southeastern margin is underlain by a thinly laminated schist and gneiss of the Alligator Back Formation (Rankin and others, 1972; Espenshade and others, 1975). These laminated and thinly bedded fine-grained rocks contrast sharply with the quartzite in the Chilhowee–Walden Creek belt, the massive gneiss of the "Elk Park plutonic group," and the large bodies of massive amphibolite in the Ashe Formation that border the plateau in the Mountain Highlands to the west. The southeastern boundary of the plateau, however, except for one small area, is coincident with the drainage divide separating the New River drainage from streams that flow to the Atlantic. This divide is 610 m higher than the Piedmont surface at the southwestern end of the boundary but only 450 m higher at the northeastern end. The Piedmont surface as well as the escarpment rising to the plateau is underlain by the laminated schists of the Alligator Back Formation, the same rocks that underlie the adjacent plateau. The origin of the escarpment is treated in some detail in the following section of this report, but the higher altitude of the New River plateau is believed to be related to the extensive outcrops of resistant sandstones and quartzite cut by the New River in its lower reaches that form a high base level for the entire river basin (Hack, 1973b).

BLUE RIDGE ESCARPMENT

The Blue Ridge escarpment, as defined here, is the narrow strip of steep land that drains southeastward to the Piedmont and Atlantic Ocean. It extends southward from the Roanoke River and ends near Rossman, N.C., on the headwaters of the Keowee River, a major tributary of the Savannah River. Most geologists who have written about the escarpment have been concerned only with the steep and spectacular part northeast of the Catawba River headwaters. In that region, the scarp rises abruptly above the Piedmont surface. The crest corresponds generally to the drainage divide and can be seen from many places at the foot. Southwest of the Catawba River, the crest is separated from the low country by one or more narrow ranges of mountains. Nevertheless, the steep rise from the Piedmont is impressive. Southwest of the Keowee River, the southeast slope of the Blue Ridge physiographic province is a little broader, the drainage is mostly toward the Gulf of Mexico, and the topography is too complex to be called an escarpment. At the northeastern end of the Southern Blue Ridge province, the drainage divide crosses the Appalachian Valley near Blacksburg, Va., forming an irregular but impressive escarpment in the valley between the Blue Ridge and the Appalachian Plateau. This part of the escarpment has been retreating southwest by headwater extension of valleys in weak rocks, primarily limestone. It is not considered in this report.

Four hypotheses explaining the Blue Ridge escarpment have been discussed in the past (see Thornbury, 1965, p. 104–106). The first hypothesis was that of Hayes and Campbell (1894), who believed that the escarpment is essentially a monoclinal flexure of a peneplain that was later dissected. According to Campbell (1896), as uplift takes place on the up or northwest side of the monoclinal flexure, stream erosion toward the southeast is accelerated and works headward, producing a scarp.

A second hypothesis was that of Davis (1903), who suggested that the scarp is a topographic discontinuity separating a set of two dissected peneplains of the same age but on opposite sides of the major drainage divide between streams flowing to the Atlantic Ocean and those flowing to the Gulf of Mexico. He thought that streams flowing to the Atlantic had an advantage over streams flowing to the Gulf of Mexico by way of the Mississippi River system because they had a shorter distance to travel to reach the sea. Davis also introduced, as part of his hypothesis, the idea that the existence of barriers to erosion in the form of resistant bedrock on the northwest side of the divide would increase the advantage of the eastern streams. There is a serious problem with the idea that distance to the sea, the ultimate base level, determined the escarpment, however, because the altitude of the base of the Blue Ridge escarpment on the southeast in some places is higher than that of the base of the mountain front on the opposite side of the Blue Ridge province. For example, the Catawba River enters the Piedmont at an altitude of about 1,500 feet. The French Broad River, however, which shares a drainage divide with the Catawba River, enters the Appalachian Valley on the Mississippi River side of the divide at an altitude of only 1,200 feet.

Wright (1931) emphasized Davis' idea that the greater height of the northwestern drainage of the Blue Ridge is related to the greater resistance of the rocks on the northwest side of the divide.

White (1950) introduced the hypothesis that the scarp was caused by faulting on a series of en echelon faults at the foot of the scarp, and he extended this concept northward along the entire Blue Ridge province to Maryland. In his hypothesis, the Asheville basin and New River plateau were once at the level of the Piedmont and are now upfaulted peneplain surfaces. Stose and Stose (1951) criticized the faulting hypothesis, pointing out that the Brevard zone is the youngest large fault zone identified in the region and is a zone of thrusting, the Piedmont being on the upthrown side. Furthermore, the zone coincides with the escarpment for only 50 to 60 km. It departs from the escarpment both to the northeast, where it is farther east in the Piedmont, and to the southwest, where it is within the Blue Ridge Mountains.

Another idea is that, although faulting may have been a factor, the faults were some distance from the present scarp. The discontinuity in the topography produced by faulting was maintained as the divide slowly shifted northwest (Thornbury, 1965, p. 105). This idea or the older idea of Hayes and Campbell (1894) seems close to a satisfactory explanation. The peneplain is not a necessary part of the explanation. All that is needed are a discontinuity between the average altitudes of the drainage systems on opposite sides of the divide and a way to maintain that discontinuity. Either a fault or a monoclinal uplift could produce an escarpment maintained by resistant rocks in the lower reaches of the main streams northwest of the divide (fig. 24). The divide may be near its original position at the time of uplift, or it may have retreated some distance as downwasting took place (Hack, 1969, 1973b).

When the escarpment is viewed as a whole, it is interesting that the average altitude of the escarpment base does not change over its entire length of 350 km. The base is approximately 450 m above sea level and generally varies less than 60 m above or below. This consistency of altitude is related to the general low slope and low relief of the inner Piedmont. It is also related to the sharply defined rise of the escarpment from the foot. The outlying ranges of mountains southeast of the escarpment have basal margins 60 to 100 m lower, depending on their distance from the escarpment. Figure 3 shows the regular configuration of the Piedmont surface along its inner margin.

The width of the escarpment and the height of the escarpment crest above its base, on the other hand, do change regionally. These dimensions are related to the regional geology and the configuration of the drainage system (see fig. 25). Near the Roanoke River, at the northeastern end, the height of the escarpment is 275 m. In this area, the escarpment is steep, and its width, that

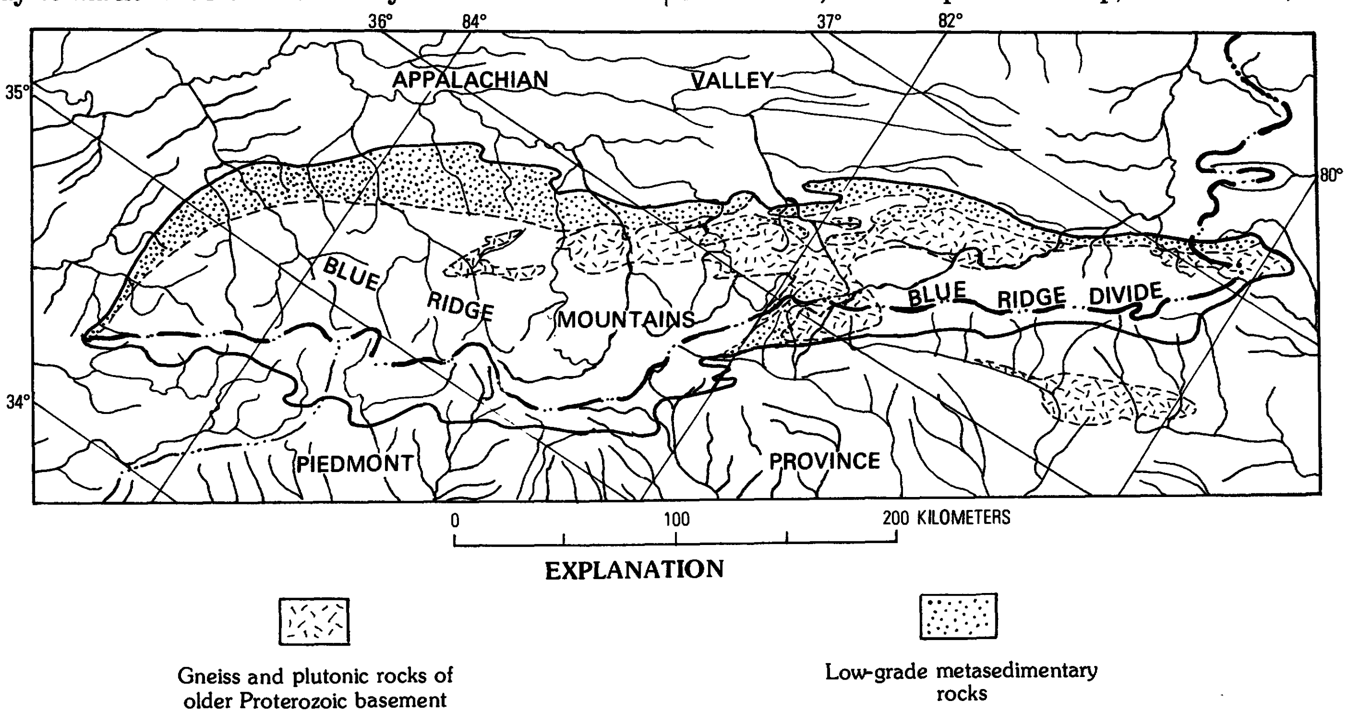


FIGURE 24. – Map of the drainage system of the Southern Blue Ridge physiographic province, showing its asymmetry. The patterned areas represent two major geologic units containing many resistant rocks. Dotted areas are clastic sequences of Proterozoic Z and Cambrian age, including graywacke-shale sequences and many quartzites. Hatched areas represent the crystalline basement rocks of Proterozoic Z age, mostly gneiss and plutonic rock. Geology from Williams (1978).

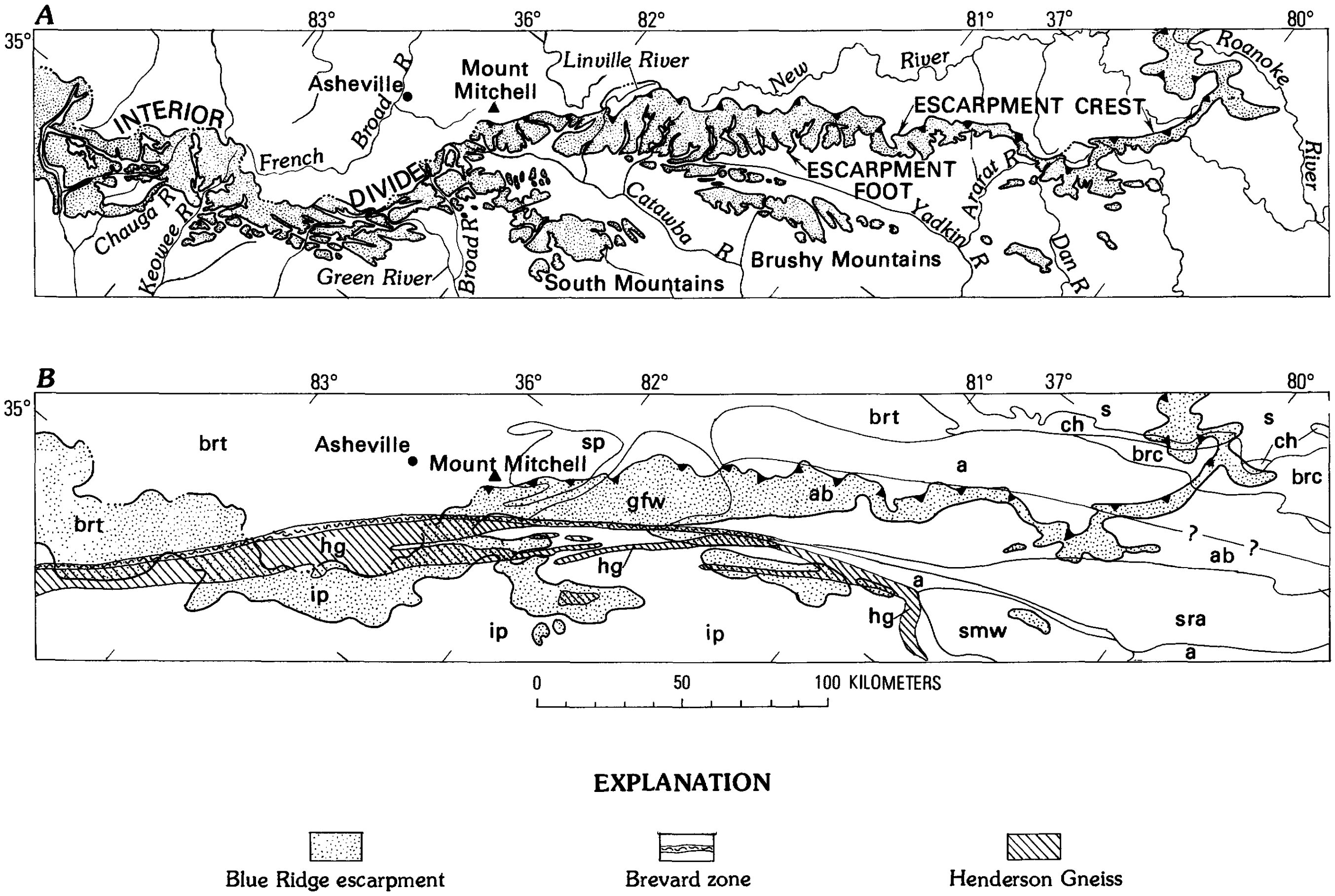


FIGURE 25. – Strip maps showing the southeastern margin of the Southern Blue Ridge and the foothills in the adjacent Piedmont. *A*, Shaded area represents the area of the Blue Ridge southeast of the drainage divide as well as the slopes of the outlying foothills in the Piedmont. *B*, Geologic units are adapted from maps by Hadley and Nelson (1971), Rankin and others (1972), Espenshade and others (1975), Goldsmith, Milton, and Wilson (1978), and Milici and others (1963). brt, gneiss, schist, and plutonic rocks of Blue Ridge thrust sheets, undifferentiated; sp, rocks of Spruce Pine area (mostly coarse-grained feldspathic gneiss); gfw, rocks of Grandfather Mountain window; a, Ashe Formation, mostly thinly layered biotite schist and gneiss; ab, Alligator Back Formation, finely laminated gneiss; sra, gneiss and schist of Smith River allochthon; smw, mostly gneiss and plutonic rock in Sauratown Mountain window; hg, Henderson gneiss; ip, rocks of Inner Piedmont, includes biotite and amphibolite gneiss, sillimanite-mica schist, granite, migmatite, and quartz monzonite; ch, Chilhowee group; brc, Blue Ridge complex; and s, carbonate rocks and shale of early Paleozoic age.

is, the horizontal distance from foot to crest, is 2 to 3 km. From this area southwest to Stone Mountain, the height and width change very little. The entire escarpment is within the Alligator Back Formation, a weak rock consisting of thinly laminated micaceous schist and gneiss (Espenshade and others, 1975). The stream profiles on the escarpment itself are highly concave, a characteristic of headwater profiles in weak rocks. Thus, the aspect of the scarp is impressively steep. On the Blue Ridge, or northwest side of the crest, the topographic relief is low, and the area drains northward to the New River, a meandering stream that flows northeast almost parallel to the scarp toward the Appalachian Valley.

In the headwater area of the Ararat River (fig. 25*A*), the crest of the escarpment is farther from the base. From this place to the northeastern margin of the Grandfather Mountain window, the escarpment has an average width of about 15 km. The crest rises to a height of 460 m above the base, much higher than the escarpment farther north. The streams drain the face of the broad escarpment through deep valleys separated by ridges. The different aspect of the escarpment when it is compared with the area to the northeast is related to two factors: First, although the country rock on the escarpment face is still the weak Alligator Back Formation, many dikes and sills of granitic rocks are intruded into the formation in this region (Rankin and others, 1972). Second, on the upland to the northwest, the New River flows through a phase of the Ashe Formation that contains amphibolite and hornblende gneiss. The local relief is higher than the relief downstream, and the gradient is steeper; thus, the small tributaries that drain toward the New River have their heads at altitudes higher than those of streams farther northeast along the escarpment crest.

In the Grandfather Mountain window and southwest of it, the local geology is comparatively complex and changes along the escarpment. The escarpment is generally broad but varies in height above its base. In the window, the width reaches 23 km, and the maximum height above the base in the Piedmont is 1,000 m. This high point is at Grandfather Mountain, where the escarpment crest is underlain by an arkosic quartzite conglomerate. The lowest point within the window is in the Blowing Rock Gneiss, a feldspathic augen gneiss.

Between the Grandfather Mountain window and the head of the Morganton basin on the Catawba River, the escarpment narrows to 11 km. The height of the crest above the base ranges from 400 m near Spruce Pine to 1,000 m at a spur of Mount Mitchell immediately northwest of the escarpment. Because of the great height and narrow width of the escarpment, the Morganton basin is a spectacular place from which to view it. The northwest margin of the Morganton basin is underlain by sheared migmatite and sheared rocks of the Brevard zone. To the northwest, the rocks underlying Mount Mitchell are coarse metagraywacke containing quartzite mapped by Hadley and Nelson (1971) as part of the Great Smoky Group. It is the juxtaposition of contrasting rock belts that gives the escarpment its grandeur at this point.

Southwest of the Catawba River, the character of the escarpment changes. It becomes a less regular mountain front that lies southeast of the drainage divide. This is the area of the Hendersonville bulge already described, a half-moon-shaped basin drained by headwaters of the French Broad River. The basin is bordered on the southeast by mountains, many of which rise to altitudes greater than 900 m. The drainage divide is close to the inner margin of these mountains. Near Hendersonville, the divide is on the basin floor, and three headwater streams drain through the mountains to the Piedmont by way of the Green and Broad Rivers. As already described, many of the streams have probably worked headward in the mountains along fracture zones and other places of weakness. This area is different from the areas of the escarpment previously described, in that the crest is not on the drainage divide. However, this difference may be an accident of the local geology. The streams flowing toward the Atlantic have penetrated relatively resistant rocks such as massive quartz monzonite, granodiorite, and paragneiss along weak but narrow zones.

Important questions related to the history and origin of the escarpment remain. Geologic evidence reinforced by recent geologic mapping makes it clear that faulting is not the direct cause of the escarpment. At present, the escarpment is maintained by a difference in stream profiles on rocks of different resistance on either side of the divide. Can evidence be found that the escarpment has migrated from some position southeast of the present one? If so, has it migrated far? One way to appraise this problem is to look for recent stream captures. Presumably, if a southeast-flowing stream captured a stream on the Blue Ridge upland as it worked headward, it would take some time to destroy the low relief in the captured part of the upland. If the Blue Ridge escarpment is migrating rapidly toward the northwest, one might expect to find areas on the upland surface that now drain toward the Piedmont, areas in which gorges that are working headward have formed. Only two areas of significant size were found in which streams draining to the Piedmont are northwest of the escarpment crest. One is the head of the Dan River, which, upstream from the crest, has a basin 92 km^2 in area. The evidence that this upland basin has been captured by the Dan River is unequivocal. The Linville River has a headwater basin 108 km^2 in area northwest of the escarpment crest, and this occurrence has often been cited as an example of stream capture. However, I believe that this river inherited its course down the escarpment by superposition on resistant rocks. Both occurrences can provide clues to the origin of the escarpment.

DAN RIVER CAPTURE

The Dan River has its source on the Blue Ridge uplands about 15 km upstream from the escarpment crest. As Dietrich (1970) discussed, the river has worked headward from the escarpment and has captured the upper basin of Reed Island Creek. The geology of the headwater area is relatively simple, and as the entire headwater area of the Dan River is in the same geologic unit (fig. 25), the anomalous profile is probably not related to differences in the rocks. The unit in which the capture took place is the Alligator Back Formation (Espenshade and others, 1975), typically a finely laminated gneiss composed of quartzofeldspathic layers a few millimeters thick separated by thin micaceous partings. It underlies areas of low relief on the upland and can be regarded as a weak rock. The drainage pattern is shown in figure 26, a subenvelope map of the upper Dan River basin. The contours are drawn on the bottom of the stream channels and show in a general way the vertical configuration of the drainage.

The drainage pattern and its relation to the crest of the escarpment suggest that a capture took place some distance downstream from locality B (fig. 26). The Dan River, cutting northward into the upland, captured the former headwaters of Reed Island Creek, which, at the

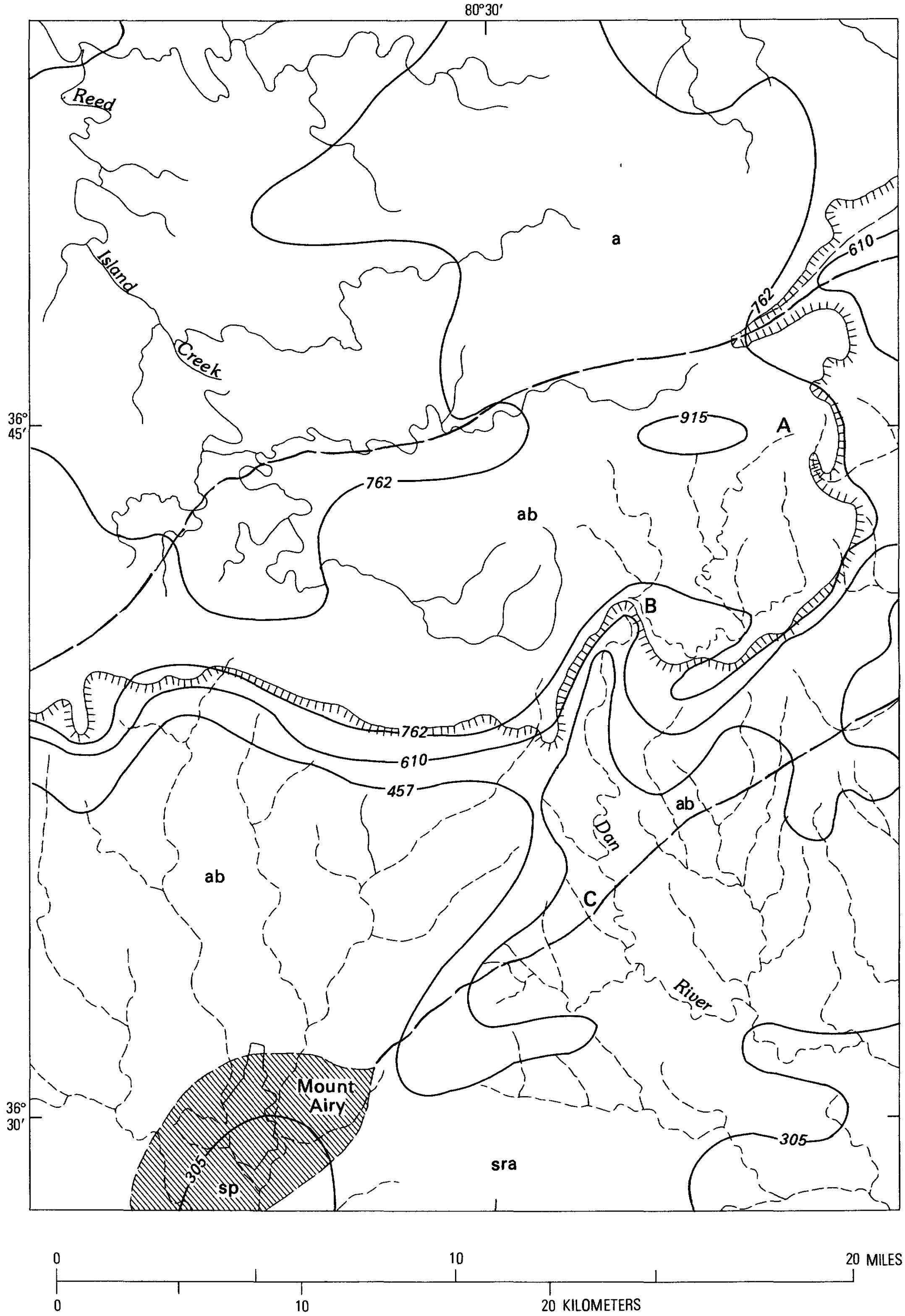


FIGURE 26.—Map of the headwaters of the Dan River, showing the area of stream capture. Streams shown by dashed lines flow toward Atlantic Ocean. Subenvelope contours are drawn on altitudes of the stream channels. Major geologic boundaries are shown by hachured lines. Geologic units modified from Espenshade and others (1975). **sra**, gneiss, schist, and gneissic granite of Smith River allochthon; **a**, Ashe Formation; **ab**, Alligator Back Formation; **sp**, granitic rock of "Spruce Pine plutonic group" (Mount Airy Granite of Stuckey and Conrad, 1958).

time, was close to the crest of the escarpment and flowed parallel to it. This idea is strongly supported by the modern profile of the Dan River (fig. 27). A striking feature is the sharp break shown near locality B on the profile. The profile downstream from the break is sharply concave. The concavity is inherited in part from the original profile of the headwater stream before the time of capture. At that time, the stream had reduced discharge, less power, and steep slopes. As a result of the capture and the increased discharge and power, the steeply concave section or knickpoint worked headward.

Figure 27 shows that the profile above the knickpoint is slightly convex. This is related to the downward component of erosion at the knickpoint. The figure also shows that the stream is intrenched for about 22 km above the knickpoint, as measured along the channel. At about 16 km from the streamhead, the valley is only 60 m deep, normal for valleys on the upland.

The observed conditions show that the river has had a complex response upstream from the capture, as would be expected from the experiments of Schumm (1973, p. 306–307). The knickpoint is decreasing in height as it works headward because of the capture and the introduction of greatly increased discharge. In the Dan River, the interpretation of the facts does not seem to be complicated by differences in bedrock lithologies, as is common.

Differences in stream power along the channel can be expressed in a semiquantitative way by use of the gradient index, which is a function of the slope or rate of fall of a reach along its logarithmic profile (see fig. 10 for definition). This index is expressed here in meters. On the upland in the headwater area, the gradient index is only 14 gradient m. The value graually increases in the convex section, reaching 510 gradient m above the knickpoint, where the index rises sharply to 1,700 gradient m. At the steepest reach, the value is 2,900 gradient m. It then drops gradually to 240 gradient m at the base of the escarpment. Because of its high discharge, the power of this river is much higher than that of streams that head on the escarpment. Hookers Creek, for example, immediately east of the Dan River, has an index of 100 gradient m, 1.22 km from its head on the escarpment crest. This index increases to 460 gradient m at 1.9 km, then declines to 140 gradient m at 5.6 km near the base of the escarpment.

The upper Dan River basin is bordered on both sides by small upland areas that drain to the Piedmont. Two small tributaries of the Smith River have worked headward past the escarpment crest to capture 5 km^2 of the upland terrain. The Ararat River just west of the Dan River has captured about 2.5 km^2 of the upland. There is no evidence, however, that any flowing streams were captured by the headwaters of these streams. Close examination of the details of the Blue Ridge escarpment will show many places where steep headwater streams have penetrated small areas on the escarpment crest. The simple geometry of the escarpment in this region, where the rocks are essentially the same on both slopes, dictates that, as downwasting takes place, the steeper slope must migrate toward the gentler slope. The scarp retreat is the horizontal component of downward erosion.

LINVILLE RIVER

The Linville River has often been cited as an example of capture of the headwaters of a stream northwest of the drainage divide by a southeast-flowing stream. The river makes its descent to the Piedmont entirely within the Grandfather Mountain window, an area of about 55×25 km, on the southeastern margin of the Blue Ridge near Morganton, N.C. In the window, rocks of Proterozoic Y and Z age and rocks of Cambrian age are exposed beneath great thrust sheets (see Bryant and Reed, 1970). The Blue Ridge escarpment crest runs diagonally across the window. The Linville River drains

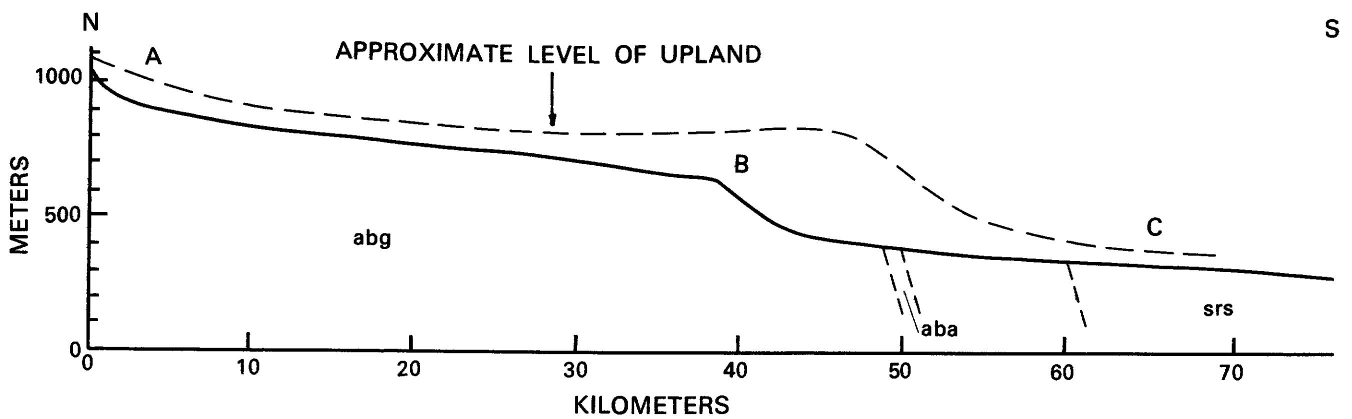


FIGURE 27.–Profile of headwaters of the Dan River. (Geologic units from Espenshade and others, 1975.) abg, typical finely laminated gneiss of the Alligator Back Formation; aba, amphibolite, greenstone, and muscovite gneiss; srs, biotite quartz schist of Smith River allochthon.

a low-relief area on the upland above the escarpment but south of the divide.

Figure 28 is a drainage map showing the Linville River and its relation to adjoining streams in the Blue Ridge. Subenvelope contours having an interval of approximately 152.5 m connect points on the streambeds. The position of the Blue Ridge escarpment is shown by the closely spaced subenvelope contours between 457 and 914 m that cross the area from southwest to northeast. As the subenvelope shows, the upper Linville River basin is higher than the basins of rivers on either side.

The geology of the area (Byrant, 1965; Reed, 1964) shows that the difference in altitude of the rivers is due to rock control. The North Toe River flows in softer rocks consisting of feldspathic schist and gneiss of the Spruce Pine area (Hadley and Nelson, 1971; Rankin and others, 1972). The Linville River, on the other hand, owes its higher altitude to an extensive outcrop area of quartzite in the Chilhowee Group exposed in the Grandfather Mountain window. The geology along the river is shown on the exaggerated profile of figure 29. In its upper course (north of lat 36°), the river is in the Grandfather Mountain Formation of Proterozoic Z age. The rocks of this formation, which range from phyllite and siltstone to felsitic volcanic rocks and arkosic conglomerate, vary widely in their resistance. The river tends to flow in the phyllite and siltstone or in alluvium deposited on them. North of locality A in figure 28, the river crosses the north tip of the lower quartzite unit of the Chilhowee Group and moves over to the Cranberry Gneiss, which it follows as far as Linville Falls, which is at the 915-m contour and at the fault contact between the Cranberry Gneiss and the Chilhowee Group.

At locality C in figure 28, the river winds steeply through the quartzite unit in a series of angular meanders. Although Wright (1928) thought that the meanders were inherited from a peneplain at a higher level, it is much more likely that the river channel is controlled by structures in the quartzite, especially by minor folds that trend northeast (see the geologic map, *in* Reed, 1964). Where the river enters Wilson Creek Gneiss, a massive rock, its profile steepens sharply and is highly concave as far downstream as the faults that bound the Grandfather Mountain window on the south. The lower relief in the drainage basin of its type locality in the Wilson Creek basin indicates that this gneiss unit is somewhat less resistant than is the quartzite of the Chilhowee Group. The profile of the Linville River in the gneiss is extraordinarily steep, however, for a river of its large discharge. The gorge bottom is cluttered with huge boulders, some more than 6 m across, giving the impression that the gorge is downcutting and backwasting rapidly. The quartzite forms a resistant capping that insures steep slopes in the reach below, analogous to the Lockport Dolomite that forms a caprock above shale beds at Niagara Falls, N.Y.

The North Fork of the Catawba River is also in rocks of Cambrian age, the Shady Dolomite, the youngest beds within the window. The dolomite itself does not crop out because it is covered by alluvium and residuum, but its presence in the valley bottom is revealed in an underground cavern as well as by outcrops of residuum. This dolomite explains the comparatively concave profile of the river. Probably the gradient would be gentler were it not for the many quartzite boulders that have been brought down by short tributaries from the quartzite slopes on the east. Details of the relation of these rivers to the geology and topography are clearly shown in the geologic maps and cross sections of Bryant (1965) and Reed (1964).

The hypothesis that the Linville River has captured a former tributary of the North Toe River has been favored by those who believe that the drainage divide and scarp are migrating northwestward by a series of captures. However, no really satisfactory evidence of the place of capture has been found. Low divides between the two drainages occur only above Linville Falls. The lowest of these is Montgomery Gap (M, fig. 28), where the head of a tributary of the North Toe is 75 m higher than the Linville River less than 1 km to the east. Wright (1931) rejected this divide as a point of capture because it would divert only the upper part of this North Toe tributary and would not explain the origin of the reach of the Linville River from Montgomery Gap to Linville Falls.

Other low divides (A and B, fig. 28) are not satisfactory for other reasons. Suppose that a capture of the Linville River took place at the place shown at locality A on figure 28, where the drainage divide between the North Toe and the Linville River is at an altitude of only 1,006 m. This low part is only 175 m above the Linville River but about 230 m above the North Toe River. The divide itself is on the Cranberry Gneiss, and the Linville River flows on the Linville Falls fault along the contact between the Cranberry Gneiss and the quartzite of the Chilhowee Group. On the other hand, the North Toe is in the relatively nonresistant rocks of the Spruce Pine area. It seems impossible that the Linville River could have captured a major tributary of the North Toe, because the Linville River would have to work headward through resistant quartzite to do so.

Wright (1931) believed that the capture took place at the Linville gorge below the meander belt (D, fig. 28) at a time before the cutting of the gorge, when the ridge crest on the west side of the present river was, in his view, part of a peneplain surface north of the escarpment. The stream that is now the upper Linville River was then a tributary of the North Toe but had a curving

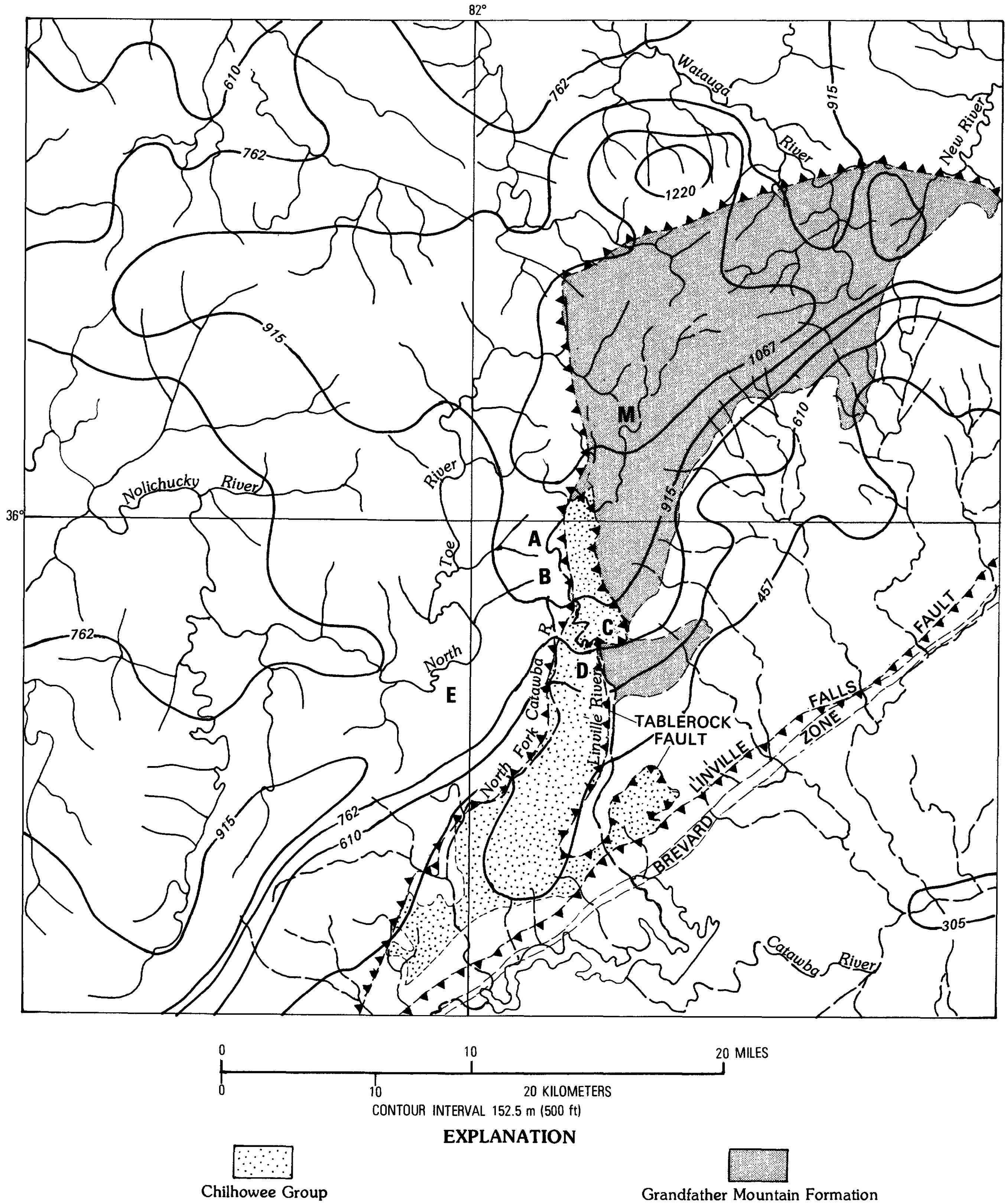


FIGURE 28.—Map of the Linville River and surrounding areas. Drainage toward the Atlantic Ocean is shown by dashed lines. Barbed lines show Table Rock fault and the Linville Falls fault, which bounds the Grandfather Mountain window. Contours show the altitudes of the stream channels. Geology from Reed (1964) and Bryant (1965). Letters M, A, B, C, D, and E refer to localities discussed in the text.

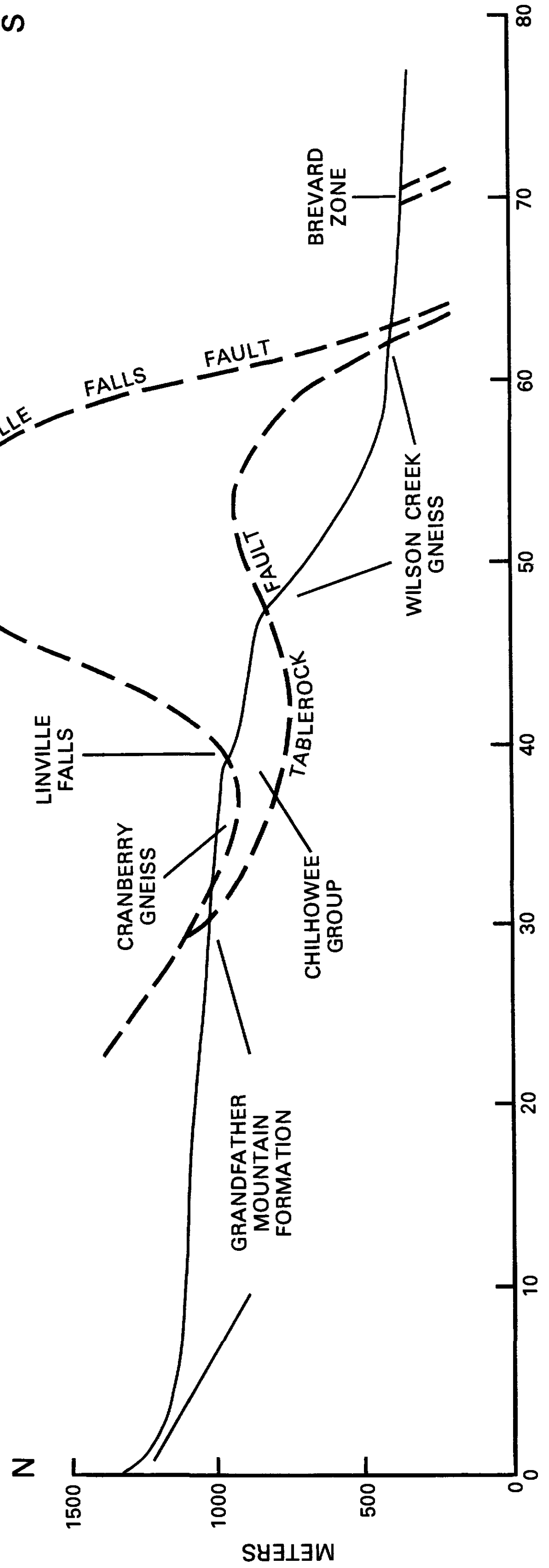


FIGURE 29. – Profile of the Linville River in the Blue Ridge, showing the relation of the channel slope to geologic units.

course across what is now the North Fork of the Catawaba River and joined the North Toe near locality E (fig. 28). Wright understood the main difficulty with this idea: the valley of the North Fork of the Catawba River has gentler slopes, or, as he put it, is more mature than the valley of the Linville River. He also recognized the speculative nature of the idea. His preferred hypothesis was an attempt to fit the facts into a framework of ideas in which the Blue Ridge escarpment is retreating by a series of piracies and in which the main streams, at altitudes between 760 and 1,060 m, are flowing in valleys that, as he believed, still contain remnants of a valley floor or Harrisburg peneplain. One can truly admire Wright's careful observation. His conclusion was almost inescapable within the limitations of the theoretical concepts of the time. It seems more likely, however, that the Linville River might at some future time be captured at locality A (fig. 28) by a tributary of the North Toe or at locality B by a tributary of either the North Toe or the North Fork of the Catawba River.

The course of the Linville River is peculiar because it is on the most resistant rocks of the area. It is best explained by superposition that took place when the topographic surface was hundreds of meters higher than it is now. The original course may have been established in the Blue Ridge thrust sheet above the Linville Falls fault plane. The fault plain dips northwest, and the river probably migrated downdip in a southwesterly direction within the least resistant rocks. Penetration of the quartzite probably first took place at some point downstream near the Piedmont. As the base of the quartzite became exposed by continued downwasting, the lower river established a course in the Wilson Creek Gneiss. The Linville Falls and steep reach below are presumably working headward and will continue to do so unless the river is captured by another stream such as the North Fork of the Catawba or the North Toe. However, because of the complexity of the geology, it is impossible to reconstruct with any certainty the factors that influenced the changes that could have taken place at higher erosion levels.

The important conclusion, however, is that the Linville River did not enlarge its drainage basin by capture of a stream on the northwest side of the divide unless it did so at a much higher erosion level. Thus, the drainage divide in this area has probably been near its present position for some time during a period involving hundreds of meters of downwasting. Migration of the escarpment is restrained by the extremely resistant rocks on the face of the scarp.

EVOLUTION OF THE ESCARPMENT

The present escarpment is clearly the steeper side of a drainage divide between stream systems draining to the Piedmont and those draining to the Applachian Valley and Plateaus. Although a simple explanation may not be possible, the asymmetry is caused in part by massive and resistant rocks far downstream in the northwesterly drainage system. These belts of hard rock are responsible for the long and steep reaches in the Blue Ridge and Appalachian Valley, as well as the Appalachian Plateaus. They help to maintain the highland northwest of the divide. Differences in the rocks along the escarpment itself cannot be the cause of the steep slopes on the southeast, because in one segment, the escarpment foot and crest are within nearly homogeneous rocks for more than 150 km. Local variations in the rock types, for example, in the Grandfather Mountain window, may cause the escarpment to be steeper and higher in some areas than in others.

The data show that parts of the escarpment crest are retreating toward the Mississippi River drainage, as proved by the Dan River capture. This capture took place in the Alligator Back Formation, which the escarpment follows for 150 km. The evidence in the Linville River area, on the other hand, shows that a capture probably could not have taken place there while the Linville River was lowering its valley through the Grandfather Mountain window, a vertical distance of 500 to 1,000 m.

By using Ahnert's (1970) formula for the relation of the mean relief to the mean denudation rate in areas 20×20 km, a crude estimate can be made of the time that it would take to lower the Linville River to its present position from a position above the window. If a mean relief of 500 m is assumed, it would take 7.5 m.y. to lower the landscape 500 m and 15 m.y. to lower it 1,000 m. If the relief were higher when the river began to erode the rocks within the window, the time might be somewhat shortened. Presumably, then, the drainage divide in the Linville River area has been near its present position or at least within the Grandfather Mountain window since early Pliocene or late Miocene time.

It is generally well known that the major drainage divide within the southern Appalachian highlands corresponds closely to a topographic high and a negative gravity anomaly (p. 6). This coincidence suggests a long period of stability for the divide and suggests that a hypothesis of escarpment evolution should be considered that does not involve substantial migration of the divide.

The configuration of the southeastern margin of the Blue Ridge from the Roanoke south into Georgia (fig. 25*A*) suggests such a hypothesis. In figure 25*A*, the major valleys on the escarpment are shown. Many of them are structurally controlled, especially in the area southwest of Mount Mitchell, where the escarpment crest becomes an interior divide. Many valleys have easterly trends. The Catawba River follows an easterly trending lineament parallel to the Laurel Creek trench valley, but offset from that valley by the Brevard zone (fig. 22). The Yadkin River follows the Brevard zone. Many small valleys follow structural trends in the outlying hills south of the escarpment, such as the Brushy Mountains. These facts suggest a long history of landscape evolution in which the drainage became established along major structural trends, including shear zones and joint sets. The process did not necessarily involve substantial migration of the continental divide, because it took place southeast of that divide. The headwaters of the Green River follow structural trends and show the process at an early stage. As the stream slopes are lowered to match the gradient of the Piedmont, the highlands on either side recede, leaving residual hills or mountains. Migration of the divide may also take place, but it depends on the gradients and relative resistance of the rocks on either side of the divide.

North of the Yadkin River, the divide is migrating northward in the weak rocks of the Alligator Back Formation. Migration is also going on in the headwaters of the Roanoke River, where the divide crosses the Appalachian Valley. In that area, the migration is rapid in the limestones.

The South Mountains and Brushy Mountains and other hills south of the Yadkin and Catawba Rivers are true residuals or inselbergs formed by the backwasting of steep slopes, as discussed by Kesel (1974). Recent mapping by Goldsmith, Milton, and Wilson (1978) at 1:250,000 scale shows that the boundaries of the high-relief areas cut across the geology, although there is a tendency for the high areas to be in rocks such as granite, sillimanite schist, and Henderson Gneiss. The mountain areas south of the Hendersonville bulge are in similar rocks, and the margin of this high area cuts across the geology, an indication that it too has been backwasting.

Divide migration on a large scale need not have taken place. The general form of the topography was determined at a much higher erosional level, possibly as long ago as the middle Mesozoic. The crest of the Appalachian highlands was probably not too far from its present geographic position, that is, trending northward from northeast Georgia to western Pennsylvania. Whatever the initial drainage pattern, it became adjusted to the weak rocks, especially in the limestone and shale of the Appalachian Valley, probably in a manner analogous to the drainage evolution of the Zagros Mountains, as described by Oberlander (1965). In the crystalline rocks, the controls of migration and headward erosion leading to stream captures may have been largely determined by structural features inherited from late Paleozoic or early Mesozoic diastrophism. Locally, however, migration of the continental divide as well as of many lesser divides is still going on.

Another problem is the sharp change in form between the escarpment foot and the Piedmont, where within a short distance, the relief may vary within a factor of 5 to 10. The residual outlying hills have similar abrupt changes in slope at the foot. In some places, the changes are within the same rock unit. This phenomenon probably is explained by the development of saprolite, which can be retained on a low-relief landscape more readily than it can on one having steep slopes. A threshold value for local relief and slope probably exists that determines the retention of a thick blanket of saprolite. When this threshold is reached, as for example, along a retreating escarpment, a sharp change in relief would take place.

TIME FRAME OF LATE TECTONISM

An important problem that has not been resolved is the time frame of the late tectonic activity. Unfortunately, stratigraphic information is available in only a few areas adjacent to the Coastal Plain, and the history of the Blue Ridge can be inferred only by an indirect and tenuous argument. Two areas are considered here. One is the Northeastern Highlands, where we have the most convincing geomorphic evidence of renewed uplift, as well as some stratigraphic evidence. The other area is the Southern Blue Ridge province, where we have only geomorphic evidence to go on.

The southern part of the Northeastern Highlands is marginal to the Coastal Plain at a place where a fault system has been carefully studied, between Fredericksburg, Va., and Washington, D.C. (Mixon and Newell, 1977). A monocline and a zone of en echelon faults lie directly west of the Potomac River estuary along the reach where the river trends parallel to the Coastal Plain boundary. The age of the faulting is Cretaceous and middle Tertiary, but later faulting is possible. The sense of movement is up on the northwest side of the fault system. The Brandywine fault system is southeast of the river and parallel to the Stafford system. The sense of motion on this fault is up to the southeast. The Potomac estuary is bracketed by the two fault systems and is within the downthrown block; thus, the southwesterly course of the river was probably determined by these fault systems. We also know that the original course of the river was toward the southeast when it flowed in sediments of Miocene age (Schlee, 1957). Thus, the diversion must have taken place in Miocene time or later.

Other faults along the strike of the Stafford fault zone cut low gravel terraces in the Washington, D.C., area and are probably as young as Pleistocene. Therefore, this region was subject to tectonic stress until very recently. The landforms show clear evidence that the

Northeastern Highlands are, or have been recently, increasing in relief. The tectonic uplift or tilting responsible for the rejuvenation must have taken place in middle to late Tertiary and possibly Quaternary time. If rates of erosion of 20 to 30 m per million years are assumed (Hack, 1980), the geomorphic evidence is unlikely to have survived unless some uplift continued into the Pleistocene. One can speculate that, after an earlier period of uplift in the Cretaceous, the topographic relief became subdued. A second uplift of the area started in late Tertiary time, producing the present forms, including the convex lower reaches of the Potomac and Susquehanna Rivers, but leaving gentle slopes in the upland areas, such as those in the Patapsco River basin.

Because of the spectacular difference in relief between the southern Piedmont Lowlands and the Southern Blue Ridge province, it is of interest to speculate on the tectonic history of that general area. The average relief in the Blue Ridge highlands is at least 5 to 10 times greater than the average relief in the adjacent Piedmont. Such an abrupt difference in relief could not be the result of lithologic differences between the two regions, because the physiographic and topographic boundary between the two does not correspond to the boundary between the Piedmont rocks and the Blue Ridge rocks. Furthermore, an assessment of the common rock types in the region does not suggest that great differences in their resistance can be the cause of the different relief. The example of the different topographic expression of granite and other rocks along the Blue Ridge escarpment and the outlying hills shows that the same rocks do have greatly different topographic expression within short distances.

Little empirical evidence of a stratigraphic nature is available related to late tectonism in the Blue Ridge. The Brevard zone is probably the youngest major fault zone in the region, and it is believed to be Paleozoic (Hatcher, 1978), or it may have existed for a long span of time, the latest motion being in the Triassic Period (Bryant and Reed, 1970). A diabase dike crosses the Brevard zone in the Grandfather Mountain window without offset. It was intruded into cool rocks, as chilled contacts show, and is not metamorphosed. It is believed to be Late Triassic but could be as young as Early Cretaceous (Bryant and Reed, 1970, p. 161). Several of the large trench valleys in the Blue Ridge highlands end at the Brevard zone. Presumably, they are deeply eroded brittle-fracture or shear zones related to the Brevard zone. The age of the shear zones is probably also Paleozoic or early to middle Mesozoic. At least one silicified breccia zone in the Piedmont near the foot of the Blue Ridge escarpment offsets a diabase dike. It indicates that some kind of tectonic movement that broke the rocks probably took place as late as middle Mesozoic. Although no direct evidence of younger tectonism has been identified in this region, differential movement associated with regional warping of the crust could have taken place along existing fracture zones. Small offsets would be difficult to detect.

Outlying hills in the North Carolina Foothill zone that are residual from a more extensive upland suggest that the Southern Blue Ridge province owes its present relief to vertical uplift younger than the features discussed above. It probably belongs to the same erosional and tectonic block as the Valley and Ridge province to the northwest. This is suggested by the deep valleys on weak rocks along its northwest margin that are similar to those in the Valley and Ridge province. The Murphy Marble belt in the Mountain highlands suggests the same, because it is eroded hundreds of meters below the peaks on either side.

Ahnert's (1970) work on the relation of denudation to relief provides some basis for speculation. It suggests that if the only uplift is isostatic, then the mean relief of any terrain will probably be reduced to 10 percent of its original value at the end of a 30-m.y. period. Thus, if the Blue Ridge has been absolutely stable and the present mean relief is approximately 300 m, the area must have had a relief of 3,000 m in Oligocene time and 30,000 m in Paleocene time. That such relief is virtually impossible, supports the idea that later tectonic uplift was involved.

In conclusion, the evidence indicates that the major features of the topography of the crystalline rock area are related to differential uplift of very late geologic age that has involved both the Piedmont and Blue Ridge. The area northwest of the Tallapoosa-Rappahannock line has been uplifted more or at a later time than the Piedmont Lowlands southeast of the line. The tectonic histories of both areas, however, have many complexities. In general, the area in which relief is highest and erosion most rapid shows the effects of rock control to the greatest degree. The present drainage divide may not be very far from its position at the end of postorogenic time, that is, early to middle Mesozoic. Nevertheless, some migration has taken place locally in the southern Blue Ridge area and probably somewhat more in the Appalachian Valley.

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