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Rock control and tectonism -- their importance in shaping the Appalachian Highlands

by John T. Hack

Abstract

The current interest in contemporary tectonic processes in the eastern United States is turning up abundant evidence of crustal movements in late geologic time. Topographic analysis of the highland areas from the southern Blue Ridge to the Adirondack Mountains indicates that most of the landforms owe their origin to erosion of rocks of different resistance rather than to tectonic processes. Most areas of high relief and high altitude have been formed on resistant rocks. The Cambrian-Ordovician belt, containing mostly shale and carbonate rock, on the other hand, forms an extensive lowland from Alabama to the Canadian border, and girdles the Adirondack Mountains. Differences in altitude can be explained by the presence of resistant rocks outside the belt; these resistant rocks form local base levels on the streams that drain the belt. A few areas may have undergone local uplift at a higher rate than areas nearby -- for example, the Piedmont region northwest of Chesapeake Bay.

Most estimates of erosion rates based on the load transported by streams and of uplift rates based on removal of the known or inferred amount of overburden during a known period of time are of the same order of magnitude, averaging about 4×10^{-2} mm per year. Rates of uplift based on study of tilted Pleistocene beaches and repeated geodetic traverses are at least an order of magnitude higher for comparable areas.

Tectonic uplift of the highlands has been slow and involves mostly warping or tilting on a large scale. Erosion rates keep up with or exceed the rate of uplift and have been sufficient to mask evidence of faulting or other differential movements. The high rates of uplift that are inferred on tilted water planes in the glaciated regions or that are measured by differences in repeated geodetic traverses cannot have been sustained for long periods of time.

Differential erosion of rocks of varying resistance has long been considered a major factor in shaping the landscape of the Eastern United States. Theories to explain the erosional history have been concerned mainly with cyclical landform development, relict surfaces such as peneplains, and the evolution of the drainage pattern (Thornbury, 1965, p. 72-87). Although most geologists agree that uplift did take place in the Eastern United States during late geologic time, that is, since the middle-Mesozoic, few have thought that local topographic forms are of tectonic origin. Faults active in late geologic time have been discovered (York and Oliver, 1976; Mixon and Newell, 1977) but it has not been proved that any modern topographic features owe their origin to contemporary faulting.

Recent studies of precise levelling data along geodetic traverses that have been rerun one or more times indicate that many of the monuments have changed in elevation during the interval between traverses. Some of these changes may be due to errors in levelling or to disturbances of the monuments, but it is generally agreed that many of the changes are related to actual motion of the ground or of the Earth's crust. The differences between traverses expressed as velocities of vertical motion are commonly quite large (see Brown and Oliver, 1976, for discussion of the method of analysis). The interpretation of the repeated traverses poses many problems. The rates of uplift or depression are commonly so high that they could not be long sustained without resulting in much larger topographic features than are observed.

In spite of the difficulties involved in interpreting evidence for recent crustal movements, there is little doubt that rates obtained from precise levelling data as well as from study of ancient strand lines indicate rapid contemporary crustal movements. Analysis of topography of the Appalachian Highlands shows, however, that differential erosion of rocks of different resistance is the major factor that controlled the development of today's topographic forms, including even large features. A few major topographic discontinuities apparently unrelated to rock type do exist, however, and can be identified. The picture that emerges from topographic analysis is that the major drainage system of the area has through long periods of time become closely adjusted to rock type. In places, it still deviates from lithologic controls. Some of the deviations occur in response to competing major drainage systems that have different base levels, lengths, and gradients, which are basically caused by rock control within the different systems. Some may be due to tectonic influence, and others may be inherited from the past.

The rate of topographic adjustment because of erosion is so rapid that detection of the details of tectonic control is difficult or impossible even in areas where it is reasonable to expect that such control exists. On the other hand, the sedimentary record preserved in the Costal Plain and Continental Shelf as well as the grossest topographic relationships are clear evidence of tectonism. The intent of this paper is to describe the major drainage systems and topographic forms of the Appalachian Highlands between Alabama and northern New York and to discuss their adjustment to bedrock. Some topographic anomalies that may be caused by differential movement of separate tectonic blocks are also considered.

I wish to thank C.S. Denny, Richard Goldsmith, Sheldon Judson, and Wayne Newell for their thoughtful criticism.

Major topographic features and drainage systems

The general outlines of the topography and drainage of the Appalachian Highlands are shown in figures 1 and 2. In figure 1, an envelope map originally defined by Stearns (1967), contours are drawn across the ridge crests and drainage divides. Essentially, figure 1 is a map of the high parts of the topography. In this paper, the envelope map is used as a reference map showing the major streams and the continental divide separating the drainage to the Atlantic Ocean from that to the Gulf of Mexico and St. Lawrence River. Figure 2, a subenvelope map (Stearns, 1967; Hack, 1973) is another way to generalize the topography. In this map, the contours are drawn on altitudes along the major streams; it shows the position of the major drainage divides as well as the general configuration of the topography. The degree of generalization is controlled by the spacing of streams used in the generalization. The streams chosen can be selected precisely simply by eliminating the upper reaches of a certain length of all the streams. The length of reach chosen for elimination determines the degree of generalization. The subenvelope is useful for the present purpose because the larger streams are flowing on bedrock, and their channel gradients are adjusted to the bedrock.

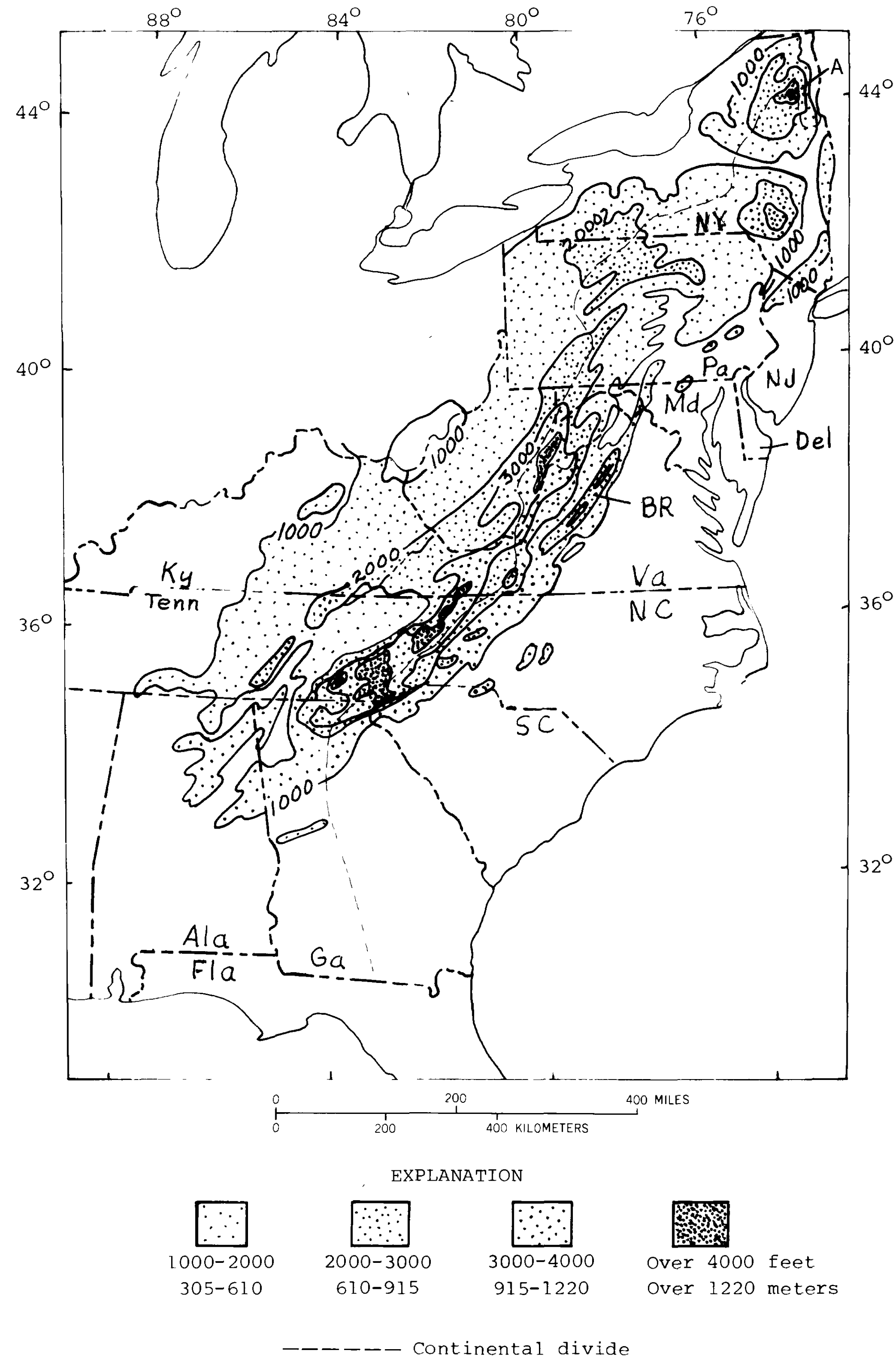
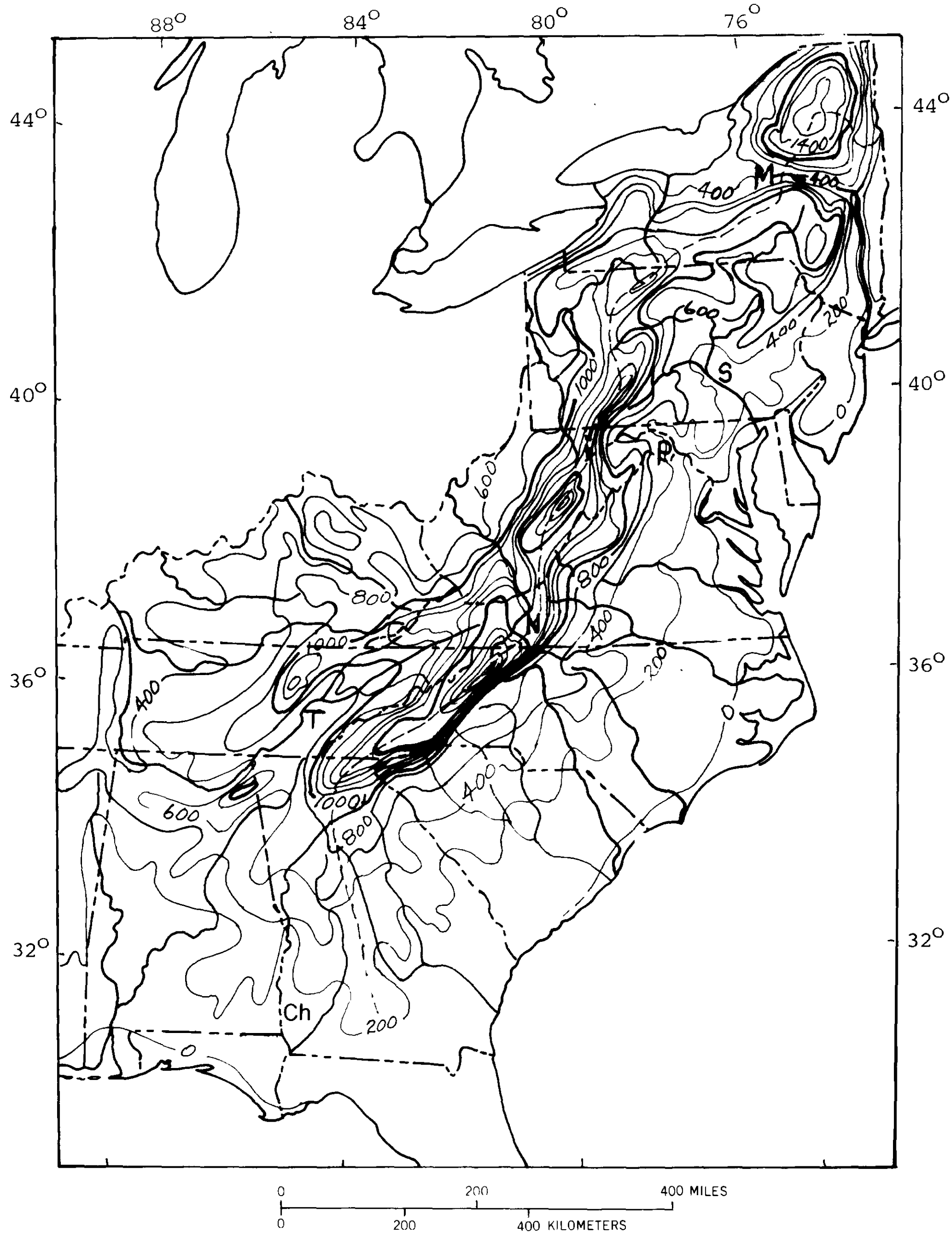


Figure 1.--Generalized topographic map (envelope map) of the Appalachian Highlands, showing the divide between the drainage to the Atlantic and to the Gulfs of Mexico and St. Lawrence. Localities referred to in text include: A - Adirondack Mountains, BR - Blue Ridge of northern Virginia.



Contour interval 200 feet (61 meters)

---- Drainage divide

Figure 2.--Subenvelope map showing contours drawn on altitudes along the major stream channels. Altitudes on streams less than 32 km long as measured from the source are not included. Ch - Chattahoochee River, T - Tennessee River, N - New River, P - Potomac River, S - Susquehanna River, M - Mohawk River.

In preparing figure 2, the contours were drawn on the streams shown on the U.S. Geological Survey's base map of the United States at 1:2,500,000 scale. Contours were plotted only on streams more than 32 km long. The resulting map shows the altitudes to which the major streams have eroded. The true altitudes of the divide areas, of course, are higher than the contours shown. Narrow ranges of mountains that do not correspond to the major divide areas do not have topographic expression on the map. For example, the Blue Ridge of northern Virginia north of Roanoke (BR, fig. 1) contains peaks higher than 1,220 m but as the entire range averages less than 32 km in width, it does not show in figure 2.

The highland area (fig. 2) from Georgia to northern New York is about 1,600 km long and averages about 240 km in width. It is broken by several large reentrants. The Tennessee River valley at T in the South divides it into two prongs. The New River valley forms a prominent reentrant at N and is the only river system that crosses the highland from southeast to northwest. Large reentrants occur at the Potomac River basin P and Susquehanna River basin S. The Mohawk Valley at M separates the Adirondack Mountains from the main highland area.

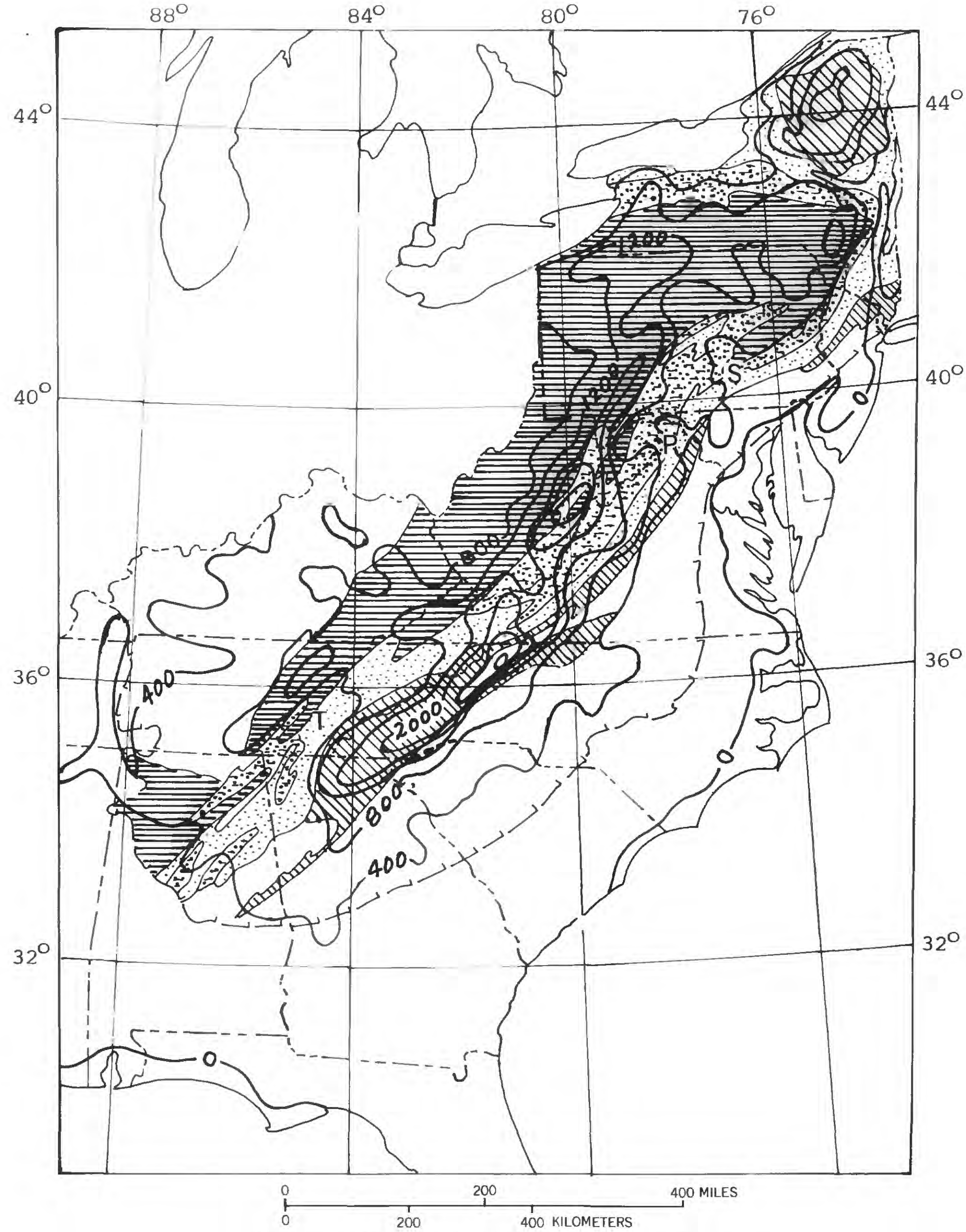
Many of the forms are closely related to the kinds of bedrock directly beneath them. This relationship probably accounts in large part for the features just mentioned, thus obscuring the evidence for the interpretation of the Appalachian and Adirondack Mountains as uplifted features. The problem, then, is the extent to which the present topography has been formed by differential erosion as opposed to tectonism.

In a general way, the highlands owe their existence to uplift, for the rocks of the highlands pass under the Coastal Plain at the southern end. On the other hand, local topographic features such as the individual ridges in the Valley and Ridge province owe their height to differential erosion. The local importance of differential erosion is evident even in the southern Blue Ridge, an area of complex geology in which it is more difficult to identify resistant or nonresistant rock (Hack, 1973, 1976). Were the highlands, including the Adirondack Mountains and Blue Ridge, uplifted as a single tectonic unit or are they composed of a group of separate tectonic blocks? If the highland is a single unit, has the uplift been differential, and at what scale or wavelength?

Relation of the highland form to bedrock geology

Figure 3 is a generalized geologic map of the Appalachian Highlands that has selected contours from the subenvelope map superimposed on it. The rock units are grouped according to broad lithologic types having different resistance to erosion. Units 1 and 4 contain extensive outcrop areas of resistant rock that tend to form highlands or high relief. Unit 2 contains alternating sequences of resistant and nonresistant rocks, which, when folded, produce the elongate ridges and valleys of the Valley and Ridge province. Unit 3 contains mostly nonresistant shale and carbonate sequences. The four units are defined as follows (largely on the basis of the summary by Colton, 1970):

Unit 1: Includes the entire sequence of rocks of Pennsylvania age as well as the Upper Devonian of New York and Pennsylvania and the Permian of West Virginia. The rocks of Pennsylvanian age are clastic sequences, both marine and continental, and contain coal beds. They include thick sandstone,



Contour interval 400 feet (122 meters)

EXPLANATION

Unit 1--Mostly rocks of Late Devonian and Pennsylvanian age.

Unit 2--Mostly rocks of Silurian to middle Devonian age.

Unit 3--Sedimentary rocks of Cambrian and Ordovician age.

Unit 4--Rocks of Precambrian age (from King and Beikman, 1974).

Inner edge of Coastal Plain

Figure 3.--Generalized map of major rock units based on their resistance to erosion. Selected contours from the subenvelope map are superimposed.

especially in the lower part of the section. These more resistant rocks crop out near the eastern edge of the Appalachian Plateau and in the area east of the Susquehanna River basin. The higher areas of the plateau of northern Pennsylvania as well as the Catskill Mountains are underlain by continental clastic rocks of Devonian age. They include coarse conglomerate and sandstone in the units designated as Upper Devonian by King and Beikman (1974).

Unit 2: Includes virtually all kinds of rock typical of platform deposits ranging from limestone and shale to sandstone and conglomerate. This unit includes the following chronostratigraphic map units of King and Beikman (1974): Silurian, Lower and Middle Devonian, Devonian, undivided and Silurian and Devonian, undivided. The Upper Devonian of King and Beikman (1974) is not included. South of Pennsylvania, the unit also includes Mississippian sequences.

Unit 3: Includes the sedimentary rocks of Cambrian and Ordovician age, mostly carbonate rocks and shale. The carbonate rocks increase in thickness from 180 m in New York to more than 3,000 m in Tennessee. Shale, siltstone, and mudstone make up most of the remainder; they increase in thickness to the north. The unit omits the basal quartzite south of Pennsylvania but includes it to the north.

Unit 4: Includes the rocks designated by King and Beikman (1974). These rocks vary widely in their degree of resistance, but they do contain resistant sequences of great thickness, especially in the south. The Ocoee Supergroup, for example, which underlies the Great Smoky Mountains and the Black Mountains, including Mount Mitchell, contains 5,200 m of metamorphic quartz-rich snadstone and siltstone. The narrow Blue Ridge of Virginia is underlain by quartzite, metavolcanic, and coarse-grained granitic rocks.

The subenvelope map superimposed on the geologic units (fig. 3) indicates that except near the southern edge of the Appalachians, the high areas are on the most resistant rocks. The Lower Pennsylvanian sequence (unit 1), the Devonian clastic sequence (unit 2), and the Precambrian rocks (unit 4) form the highest areas. The bifurcation of the southern highland occurs where the main highland shifts southeastward to correspond with the large outcrop area of unit 4. The highland on the Pennsylvanian sequence continues to the southwest, forming the western prong.

In southern Pennsylvania, sandstone of Pennsylvanian age holds up large plateaulike areas. In the northern part, the Devonian and Mississippian clastic rocks underlie the present erosion surface and form the northern glaciated section of the Allegheny Plateau. The Catskill Mountains of New York locally include peaks as much as 1220 m high. The Adirondack Mountains form a separate dome underlain by rocks of Precambrian age.

In contrast, the large areas of nonresistant rocks found mostly in Units 2 and 3 are responsible for the large reentrants in the Appalachian highlands at T, P, and S in Figure 2. The large valley at T is drained by the Tennessee River and its tributaries and corresponds to a broad outcrop area of carbonate rocks (Unit 4). At P, the principal tributaries of the Potomac are subsequent streams flowing in valleys in carbonate and shale (Unit 2). The same is true of the Susquehanna basin at S.

The belt of Cambrian and Ordovician rocks

At the crossover of the subenvelope crest from the Precambrian area to the plateau in southwest Virginia, the altitude of the subenvelope (fig. 3) appears to be unrelated to the bedrock. It crosses the non-resistant Cambrian and Ordovician sequence at an altitude of more than 730 m comparable with, or higher than, altitudes at many other places along the crest. At the Hudson River, for comparison, the Cambrian and Ordovician belt is almost at sea level. Study of this belt of rocks shows that the altitudes within it can be explained at least in part by adjustments of the major streams that cross the belt to bedrock down-stream from the belt. These adjustments, in effect, form local base levels at various altitudes that control the profiles upstream.

Figure 4A, which illustrates the argument for this hypothesis, is a profile along a series of stream reaches that flow along the strike in the Cambrian and Ordovician rocks. The profile is drawn along the streams as the crow flies, and does not include bends or meanders. Altitudes are on the valley floors or flood plains adjacent to the streams. Except for small parts of the Coosa River, the streams are entirely within the Cambrian and Ordovician rocks. However, all have tributaries or upper reaches that head in more resistant rocks, and they depart from the belt to cross terrains having a great variety of rock types. The divides between the streams within the belt are narrow everywhere, so that a continuous longitudinal profile is formed.

The profile along the strike of the Cambrian and Ordovician rocks (fig. 4) is divided into concave, upward segments, each of which is part of the drainage basin of a large stream that exits the Cambrian and

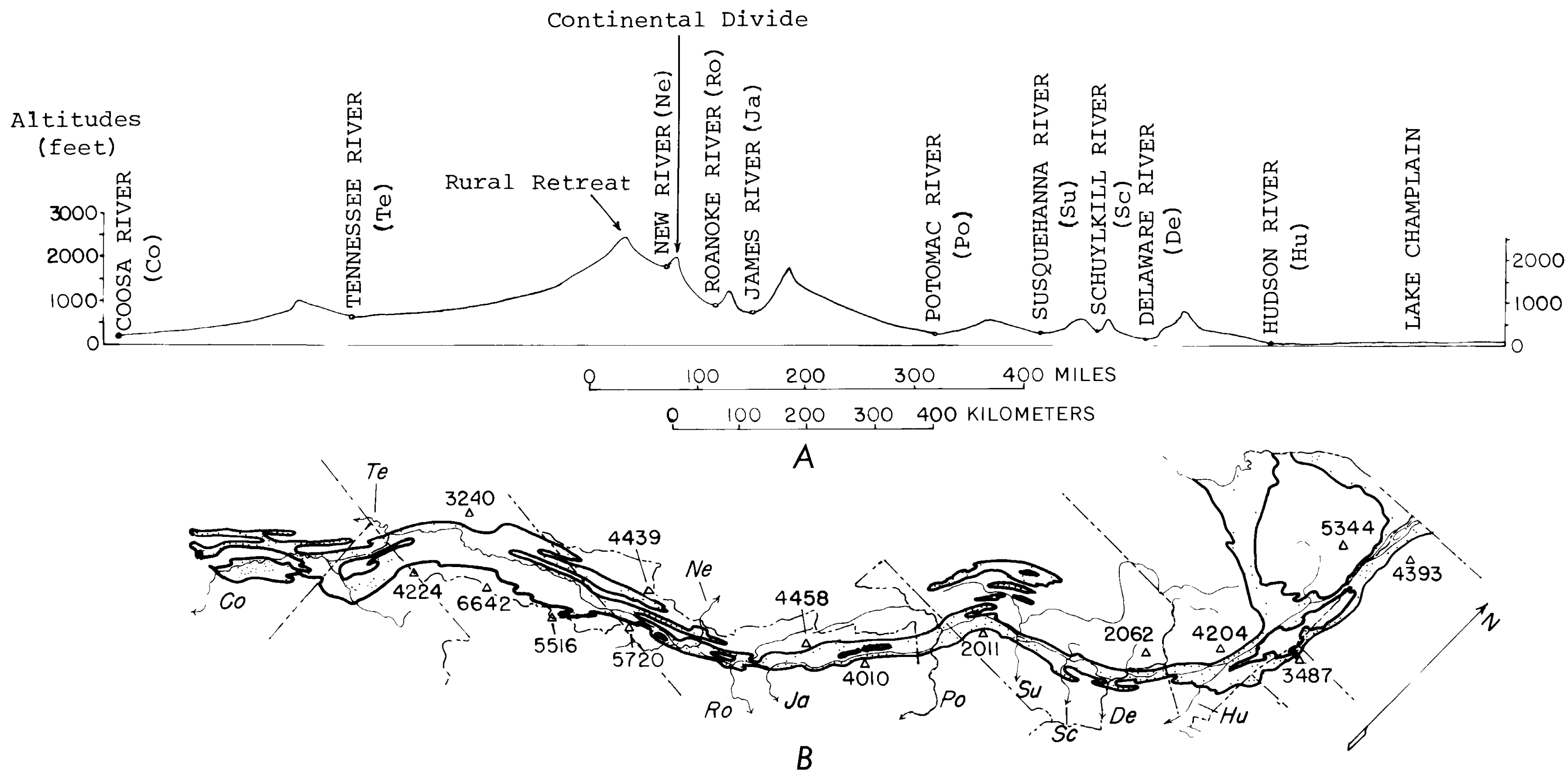


Figure 4.--Profile (A) and outline map (B) of the Cambrian and Ordovician belt from Coosa River to Lake Champlain. Numbers represent altitudes of selected high peaks.

Ordovician belt. The high point is near Rural Retreat south of the continental divide, between headwaters of the Tennessee and New Rivers. This place is close to where the Cambrian and Ordovician belt narrows and is broken up by ridges of more resistant rocks younger than Ordovician. Profiles of streams of small and intermediate size such as those in the belt are generally shaped by the amount of discharge and various factors related to flow resistance of the channel, the most important of which are probably load and size of bed material. The channel slopes of streams are normally inversely proportional to the discharge, and as discharge and length are directly related quantities, channel slopes are generally inversely proportional to a function of length. This explains the concave-form characteristics of most longitudinal valley profiles. As channel slope tends to be directly proportional to a function of size of bed load, the concave profile is modified in steepness and in the details of its form by the resistance of the rocks in its drainage basin. The high point at Rural Retreat on the divide between the Tennessee drainage and the New River drainage is a striking feature of the composite profile (fig. 4A). It can be explained by differences in the geology of the Tennessee River drainage basin and the New River basin (Hack, 1973). The Tennessee system descends from Rural Retreat via the Holston River as far as Chattanooga through a broad valley in shale and carbonate rocks. The altitude at Chattanooga where the river system exits the Cambrian and Ordovician belt is only about 210 m. Below this point it crosses resistant sandstone beds in a short reach below Chattanooga. It crosses northern Alabama through an area underlain mostly by Ordovician and Mississippian limestone. Thus, almost its entire course is in nonresistant rocks.

The New River, on the other hand, heads in the Blue Ridge and crosses the Cambrian and Ordovician belt at a narrow point. It then flows northward toward the Ohio, crossing thick sandstone sequences of Pennsylvanian age. Its gradient is actually steepened in these reaches in spite of the increased discharge. Its steep descent ceases above Charleston where it joins the Gauley River at an altitude of about 200 m.

Northeast of the New River, the regional drainage is to the southeast into the Atlantic by streams that head along the resistant rocks of the plateau and flow more directly across the Cambrian and Ordovician belt. Only in the Potomac and Susquehanna basins are there major streams parallel to the regional strike. The Roanoke and James cross resistant Precambrian rock before reaching the Piedmont lowlands, but the outcrop widths are narrow compared with those of the resistant rocks crossed by the New River. The Potomac exits the belt at a place where the resistant rocks are usually thin (Hack, 1965, p. 28). The Susquehanna and Schuylkill Rivers flow directly across the nonresistant beds of the Newark Group in the Triassic and Jurassic Basin. The Delaware crosses only a narrow band of rocks of Precambrian age in the Reading Prong. The absence of a high resistant barrier between the Cambrian Ordovician rocks and the Piedmont lowlands in Pennsylvania permits even small streams like the Schuylkill to survive without capture by subsequent drainage from the Susquehanna or Delaware. In New York State, the Cambrian and Ordovician belt forms a narrow lowland separating the Appalachian Plateaus province on the west from the New England province on the east. The altitudes are low; the Hudson River is at tide level as far north as Albany. The belt intersects tidewater again at the

St. Lawrence River. Adjustments to resistant and nonresistant rocks do not explain why the Hudson River cuts through the Precambrian rocks of the Hudson Highlands in a preglacial gorge southeast of the Cambrian and Ordovician belt. Many explanations have been offered (Thornbury, 1965, p. 165).

The relationships cited above show that the general topography of the belt can be explained without involving the hypothesis of differential uplift, except, of course, the obvious condition that the entire Appalachian system formed as a result of one or more periods of uplift at some time in the past. The height of the divide at the New River crossing is explained by the necessity for that river to cross many miles of resistant rocks downstream exposed in the Appalachian Plateau. North of the New River, the lower areas are explained by the fact that the drainage out of the Cambrian and Ordovician belt crosses lesser barriers to reach the lowlands.

The altitudes marginal to the Cambrian and Ordovician belt can be judged by the spot altitudes of peaks shown on figure 4B. These altitudes are generally high all along the belt except locally in Pennsylvania, where the belt is bordered on the south by weak rocks and on the north by narrow ridges underlain by thin resistant beds.

Duration of erosion

The general evidence for a close adjustment of the surface forms to the rocks, especially obvious in the Cambrian and Ordovician belt, is an indication that the region must have been exposed to subaerial erosion for a long time; however, more positive evidence does exist. A narrow zone of deposits containing lignite, wood, leaf fragments, and other organic material was called the Brandon Formation by Clark (1891), who regarded these occurrences as of Eocene age. The deposits are now known to range in age from Cretaceous through Tertiary. They extend down the southeastern margin of the Appalachian basin from Vermont as far south as Alabama. These deposits are small and generally occur in residuum overlying the Lower and Middle Cambrian carbonate rocks of the Shady and Tomstown Dolomites and their equivalents. The surface topography in the zone of these deposits is in many places pitted by modern sag ponds that commonly contain peat and other organic matter of Pleistocene to Holocene age. The close association suggests a common origin for the organic deposits. They have evidently been formed by solution in the underlying carbonate rocks as part of a continuous process operative since the Cretaceous. The literature on the occurrences is scattered, but a discussion of the implications of a Cretaceous locality in Pennsylvania has been presented by Pierce (1965). Some modern sag ponds have been described by Craig (1969). Additional references have been cited by Mathews (1975). As the fossil deposits could have slumped downward thousands of feet since they formed because of solution of the underlying carbonate rocks, they do not necessarily inform us about the past nature of the relief, but their occurrence and the lack of marine fossils indicates that the Cambrian and Ordovician belt has been exposed to erosion for a long time.

Evidence for late tectonic activity

Abundant evidence indicates that tectonism has taken place in late geologic time, that is, during the late Mesozoic and Cenozoic. The evidence includes changes in the rates and kinds of deposition through time on the

Continental Shelf, late faults, tilting or warping of features of the coastal zone, differences in height of land, and areas of anomalous relief. The sedimentary record in the Coastal Plain and Continental Shelf has, of course, been used by many to interpret the history of the hinterlands, and this record will surely become increasingly useful as exploration of the shelf continues. Analysis of the Coastal Plain and shelf deposits, however, is outside the scope of this paper, though the use of sedimentary volumes to interpret erosion rates on the land should be mentioned. Mathews (1975) estimated, by analysis of a series of sections off the Atlantic coast, that at least 2,000 m has been eroded during the Cenozoic on the adjacent land, an average rate of 0.027 mm per year. The data also suggest that erosion was greatly accelerated during the Miocene.

Data (table 1) from the COST No. B-2 well on the outer shelf off New Jersey (Scholle, 1977) permit an estimate of sedimentation rates since the beginning of the Cretaceous, though density, compaction, and other factors are not taken into account. As this well seems to penetrate a typical sequence of major units, the sedimentation rates probably are roughly indicative of relative erosion rates in the source area. The data support the idea that rates were high in Early Cretaceous and Miocene time.

Judging by the quick response of the crust by uplift to the removal of the continental glacier of Pleistocene age, changes in erosion rates probably have been balanced closely by changes in uplift rates; the reverse was probably also true. Thus, high uplift rates probably prevailed in the Early Cretaceous and Miocene, and lower rates, from Late Cretaceous

Table 1.--Thickness of sediments penetrated in COST No. B-2 well (Scholle, 1977) and estimated deposition rate

Period or epoch	Duration (millions of years)	Thickness of sediment (meters)	Rate (mm per year)
Pliocene and Pleistocene	12	130	0.011
Miocene	14	860	.061
Oligocene	12	150	.012
Eocene	16	220	.014
Paleocene	11	60	.005
Late Cretaceous	35	980	.028
Early Cretaceous	36	2470	.068
Total	136	4870	
Average			.036

to Eocene time. Nevertheless, uplift was probably continuous throughout post-Jurassic time.

That faulting took place in late geologic time in the Eastern United States is not easy to demonstrate. Nevertheless, actual faults cutting beds of late geologic age have been observed at more than 30 places in eastern North America (York and Oliver, 1976). Probably many more faults will be discovered. Most of them are along the Fall Zone where thin sedimentary strata overlap the basement. Two of these faults have been studied in some detail. The Stafford fault zone, west of the Potomac River (Mixon and Newell, 1977) consists of en echelon faults and a parallel monocline 56 km long. The displacements are small, not exceeding 60 m. They involve sediments as young as middle Tertiary, but some movement in the late Tertiary is possible. The sense of motion is reverse. A somewhat similar group of faults that had small displacements in late geologic time has been found near Augusta, Ga. These faults are also reverse and involve Cretaceous and younger sediments, possibly as young as Miocene (David Prowell, oral commun., 1977).

The displacements observed are too small to account for the rate of uplift of the area inland from the Coastal Plain as determined by the sedimentary record. Other faults farther inland would be difficult to detect because of the sparse sedimentary cover younger than the basement.

Another tectonic process of late geologic age is the general downward tilt of the coastal zone to the northeast. This tilt can be seen on figure 3, which shows the Fall Zone or inner margin of the Coastal Plain. In Georgia, the Fall Zone is crossed by streams at an altitude of 120 m.

At the northern end of Chesapeake Bay the zone is at sea level and in New England it is below sea level. The tilt of the Fall Zone is matched by the tilt of the Continental Shelf. These features imply a broad tilt of the continent down to the northeast. The head of tidewater extends inland different distances in different rivers, as shown by the zero contour in figure 3, but this phenomenon is not necessarily related to the tilt of the coastal zone. Eardley (1964) explained the tilt of the Coastal Plain and Continental Shelf as part of a worldwide process of poleward rise of sea level related to a long-term decrease in period of the Earth's rotation, beginning in the Cretaceous. The fact that some river valleys appear to be drowned more deeply toward the north, however, suggests that a process acting more rapidly may be involved.

Differences in relief or sharp changes in altitude may be evidence for unequal uplift in the past, provided that relief differences are in rocks of similar resistance to erosion. Figure 5 is a simplified relief map of the Piedmont and Blue Ridge provinces south of the glacial border. The two provinces have extreme differences in relief. The Piedmont relief is comparatively low over extensive areas, and the map is intended to emphasize differences in relief within the Piedmont. Most of the areas of moderate relief on the outer Piedmont are directly related to resistant bedrock. For example, the Pine Mountain belt containing quartzite beds forms a high relief area at A. Cataclastic rocks of the Brevard zone occur at B. Resistant beds of the Kings Mountain belt are found at C. At area D, high relief is caused primarily be felsitic rocks of the slate belt. A large area of moderate relief (E) lies between

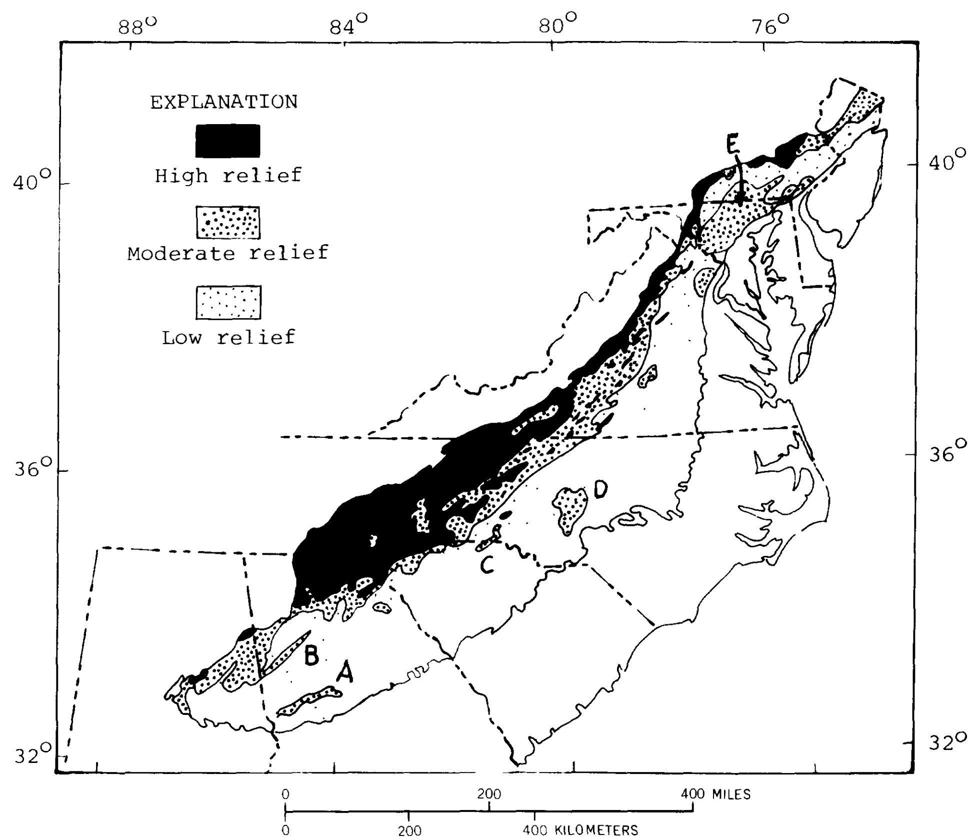


Figure 5.--Generalized relief map of Piedmont and Blue Ridge provinces. The relief classes are based on the number of contours within square areas 10 km on a side, as indicated on the 1:250,000-scale series of topographic maps by the U.S. Geological Survey. Less than 90-120 m in low-relief areas, 120-240 m in moderate-relief areas. High-relief areas are generally more rugged and have relief of more than 240 m and as much as 1,200 m. Letters indicate areas discussed in the text.

the Coastal Plain and the Culpeper and Gettysburg Triassic and Jurassic basins. Its highest point exceeds 305 m above sea level near Westminster, Md. This area was referred to as the Westminster Anticline by Campbell (1929, 1933), who believed that it represented an upwarped peneplain of Miocene age on the basis of warped stream terraces along the Potomac, Susquehanna, and Schuylkill Rivers. Although the existence of a peneplain of lower relief in Miocene time cannot be proved, the higher relief that exists there now is probably the result of upwarping of the area at a higher rate than other parts of the Piedmont farther south.

Several arguments support this idea. The rocks in the area of the Westminster anticline, or uplift, to use a less specific term, are both igneous and sedimentary, similar to rocks of the Virginia Piedmont farther south where relief is lower. They include gabbro, granite, gneiss, diamictite gneiss, quartzose schist, volcanic rock, phyllite, and marble.

The altitude of the highest parts of the Westminster uplift exceeds 305 m. Total relief within 100-sq-km areas was measured on 1:24,000-scale maps of the following quadrangles: Downington, Pa. (162 m), and Mt. Airy (125 m), Winfield (125 m), and Rockville (105 m), Md. In low-relief areas shown on figure 5, total relief is generally less than 90 m in areas 10 km on a side.

On its west side, most of the high-relief area is bounded by a low escarpment that overlooks Triassic and Jurassic basins to the west. As these basins are underlain mostly by shale and mudstone, the relatively low altitude and low relief are perhaps related to rock control. On the east, the Piedmont margin is overlapped by a sequence of sedimentary deposits ranging in age from Early Cretaceous to late Tertiary (Darton, 1951). The surface of basement and base of the Cretaceous has a slope of about 0.02 in Washington, D.C. The slope of the Miocene in Washington is only 0.005. A river gravel deposit resting on the Miocene, believed to be of late Tertiary age has an even lower slope, averaging about 0.0025 at the same cross section. Although the slope of the surface on which these sediments were laid down is not known, the Lower Cretaceous gravel is similar in size composition to the upper Tertiary gravel. Thus, the difference between the two slopes (approximately 0.0175) may reasonably be taken as a measure of the tilt that has taken place. Assuming an age difference of 100 million years between the two gravels, the rate of tilt would average about 1.75×10^{-4} mm per km per year. If the present slope tangent of the basement were projected 50 km inland from the Fall Zone to a point near the center of the highest altitude near Westminster, Md., assuming no erosion and no bending, the altitude of this point would be about 1,000 m above sea level or 700 above the present surface. The average rate of lowering at this point since middle Cretaceous time would have been 7×10^{-3} mm per year. As will be shown in another section, these rates of uplift and erosion are somewhat lower than average rates calculated for other places and other time spans using a variety of methods. (tables 2 and 3).

It should be noted that the present slope of the basement beneath the Coastal Plain in the Washington area, as indicated on the Tectonic Map of the United States (U.S. Geol. Survey and Am. Assoc. Petroleum Geologists, 1962), is greater in the northern Piedmont near Washington east of the Westminster uplift than it is farther south. At the Virginia-North Carolina boundary, the slope tangent of the basement near the inner edge of the Coastal Plain is about 0.005 as compared with 0.02 in the Washington area.

In the northern Piedmont, the increased relief is also related to the general depression of the Coastal Plain in the Salisbury embayment, which brings the edge of the Piedmont to sea level in the larger river valleys like the Potomac. The Pleistocene lowering of the sea level lowered the base level still more and probably affected the relief inland, at least close to the larger streams (Hack, 1975, p. 96).

The most dramatic difference in relief unaccounted for by geology is along the Blue Ridge escarpment. As shown by figure 3, a steep escarpment in Georgia and North Carolina crosses the boundary between the Precambrian terrane and generally younger metamorphic rocks of the Piedmont. The same escarpment also crosses the Cambrian and Ordovician belt, and its crest forms the continental divide between the Gulf of Mexico and Atlantic drainages. The relief and altitudes of the terrain on either side are strikingly different. Within areas 10 km wide, the relief averages less than 90 m in the Piedmont. In areas of comparable size in the Blue Ridge, the relief ranges from about 180 to more than 1,000 m and altitudes rise to more than 2,000.

Davis (1903) explained the escarpment by the differences in lengths of the rivers on either side in their descent to the sea. White (1950) thought the escarpment was essentially a fault scarp. Hack (1973) expalined the escarpment as due to differences in the geology crossed by the streams on either side. In other words, the westward-flowing streams cross a series of local baselevels formed on resistant rocks, thus in effect, maintaining the high altitude of the continental divide. The faulting hypothesis remains a possibility, though clear evidence for a fault of faults has not been found. Nevertheless, over long distances, the rocks forming the present escarpment are not more resistant than are those on the adjacent Piedmont at the toe of the scarp. If no faulting is involved, tilting must have taken place on the southeast side of the Blue Ridge.

Rates of Erosion and Uplift

Many estimates have been made of erosion and uplift rates based on a variety of methods. In addition, contemporary uplift or tilting of the land has been measured in the field by repeated geodetic traverses. Some of these estimates that relate to the Appalachian Highlands are summarized in tables 2 and 3. Table 2 deals with erosion and uplift rates measured at a single locality within an area. These rates are absolute values related to a constant but unspecified data base expressed as erosion in millimeters per year or as velocities of uplift in millimeters per year.

The estimates in table 3, on the other hand, are differences between rates at points along a linear feature or traverse. In other words, they are relative rates or angular tilt rates rather than uplift or subsidence. They are expressed in millimeters per year per kilometer (compare with Brown and Oliver, 1976, p. 17).

The first five items in table 2 are based on amounts of material being carried out of one or more drainage basins over a period of at least a year. Item No. 1 in the table is from a general paper by Judson and Ritter (1964), which contains a review of earlier literature as well as newer data. The first three estimates deal with fairly large drainage basins. Items 4 and 5 are small watersheds that may not be typical, though

Table 2.--Estimates of erosion rates

Item No.	Process	Rate (mm per year)	Sources of data
1	Modern erosion, North Atlantic drainage basins	4.8×10^{-2}	Judson and Ritter, 1964
2	Modern erosion, South Atlantic and Gulf basins	4.1×10^{-2}	Judson and Ritter, 1964
3	Modern erosion, South Fork of Shenandoah River watershed	4.3×10^{-2}	Hack, 1965
4	Modern erosion, small watershed, Coweta, N.C.	9×10^{-3}	Berry, 1977
5	Modern erosion, small watershed, Maryland Piedmont	4.6×10^{-3}	Cleaves and others, 1970[1/]
6	Cenozoic erosion rate based on sediment volumes on Continental Shelf (Atlantic drainage)	2.7×10^{-2}	Mathews, 1975
7	Phanerozoic erosion rate for U.S. based on sedimentary volumes. Figure given is minimum; estimated by authors as up to six times too low)	1×10^{-2}	Gilluly and others, 1970
8	Uplift and erosion (some possibly tectonic) based on depth of intrusion of Stone Mountain Granite, Georgia	4.1×10^{-2}	Whitney and others, 1976[1/]
9	Uplift and erosion (some possibly tectonic) based on thickness of Paleozoic sequence in eastern Pennsylvania	5.3×10^{-2}	This paper, data from Colton, 1970
10	Removal of overburden from Middle Devonian rocks and conodont color in Appalachian basin - rate calculated since Permian	4.8 to 8.8×10^{-2}	This paper, data from Epstein and others, 1977

1/ The rates reported here were calculated by Hack from data in reference cited.

Table 3.--Estimates of tilt rates

Item No.	Tilted feature	Rate in mm/year/km	Source of estimate
1	Modern tilt rate in Potomac basin if balanced by average erosion rate of 4×10^{-2} mm per year	3.5×10^{-4}	This paper
2	Tilt rate of basement rocks from the Cretaceous to late Miocene, Washington, D.C.	1.75×10^{-4}	This paper
3	Domal uplift of Canadian Shield since 7,000 B.P. (average tilt of southeast part of dome)	2.2×10^{-2}	Walcott, 1972, fig. 16[1/]
4	Downwarp to south of beach of high stand of Champlain Sea since 10,000 B.P.	9.3×10^{-2}	Denny, 1974, p. 42[1/]
5	Upwarp to south of level line west of Lake Champlain, Rouses Point to Ticonderoga after 18 years	3×10^{-2}	Isachsen, 1975[1/]
6	Downwarp of rerun level line 100 km long between Atlanta and Columbus, Ga.	5.5×10^{-2}	Brown and Oliver, 1976, fig. 11

[1/] The tilt rates reported here were calculated by Hack from data in the references cited.

the methods of analysis used in these two cases are rigorous. Items 6 and 7 are estimates based on the volume of sedimentary sequences that have been redeposited after erosion from the land. These estimates are the same order of magnitude as rates in Items 1-3. Item 8 (Whitney and others, 1976) is an entirely different kind of analysis based on depth of emplacement of a granite body in metamorphic rocks. By using isotope ratios, the age and temperature, and hence, by inference, the depth of emplacement can be estimated. The rate of removal of overburden can, of course, be determined from these data, though whether all the removal was due to subaerial erosion is not certain. It is interesting that Item 8 is comparable in magnitude with Items 1, 2, 3, and 6.

Items 9 and 10 are similar in concept to Item 8 in that uplift rates are estimated by inferences about age and depth in the earth before the time of uplift. It is assumed that as the rocks are now at the surface, the average uplift rate and erosion rate have been the same. Item 10 is taken from a study of conodonts (Epstein and others, 1977), which shows that conodont colors in the Appalachian basin consistently indicate depth of burial. These data give results similar in magnitude to the others. The two modern erosion rates from small watersheds (items 4 and 5, table 2) are significantly different, but it should be noted that the rates in these two places are lower rates rather than higher.

The results obtained from direct measurements of tilt cannot be compared with the uplift velocities or erosion rates given in table 2 without making some sort of conversion that involves assumptions about the distribution of the uplift or erosion rates within an area. To do so, the absolute rates must be converted to tilt rates. Item 1 in table 3 is such a conversion. It is based on the assumption that the Potomac Basin is now being tilted upward at its western headwaters near the continental divide relative to the mouth at sea level and that uplift is balanced by an erosion rate of 0.04 mm per year (from table 2). As the erosion rate varies throughout the basin, the average rate is assumed to be centered at a point about halfway (112 km) from the river mouth to the continental divide. On this basis, the tilt-rate estimate in the basin is 3.5×10^{-4}, expressed as millimeters per year per kilometer. This rate can be compared with tilt rates measured along level lines (items 5 and 6, table 3) or along water plans whose age is known (items 3 and 4, table 3). The data show that the measured modern and post-Pleistocene tilt rates are at least an order of magnitude higher than rate number 1, on the basis of the erosion rates estimated in table 2. The same conclusion is reached by comparing erosion estimates with absolute uplift rates of domal features that have been measured. One example is the Canadian Shield, which near its center at Hudson Bay is estimated by Walcott (1972) to be rising at a rate of 20 mm per year, a rate much higher than any of the absolute rates of table 1. In this example, the uplift area has a radius of about 1,000 km, and this rapid rate is, of course, compatible with the calculated tilt rate for this area (table 3, no. 3).

Another example of a high absolute rate is the Adirondack dome, studied by Isachsen (1976). On the basis of differences between two geodetic traverses, assuming zero base at the south edge of the rising area at Utica, 240 km from the center, the Adirondacks are rising at a rate of 3.7

mm per year. This translates to a tilt rate of 2.4 X 10^{-2}, approximately the tilt rate of the southern margin of the Canadian Shield. The northern flank of the Adirondack domal uplift also shows on a geodetic traverse along the shore of Lake Champlain (table 3, item 5), but the direction of tilt is the reverse of the postglacial tilt of the water planes parallel to it (table 3, item 4). Therefore, if the tilt is real, it has only recently reversed its direction.

A brief consideration of the rate of change of landforms in postglacial time is useful. Glacial deposits of Wisconsinan age in the central United States, as much as 55,000 years old (Flint, 1971, p. 559), have a constructional topography that has undergone little change in form since deposition. On the other hand, glaciated terrain of Illinoian age, more than 100,000 years old, has been considerably modified, though the deposits themselves are mostly preserved. Nebraskan glacial deposits, which are more than 400,000 years old, have been eroded extensively, and none of the original topographic form survives. These general conditions are consistent with an average erosion rate of 0.04 mm/year or 4 m per 100,000 years.

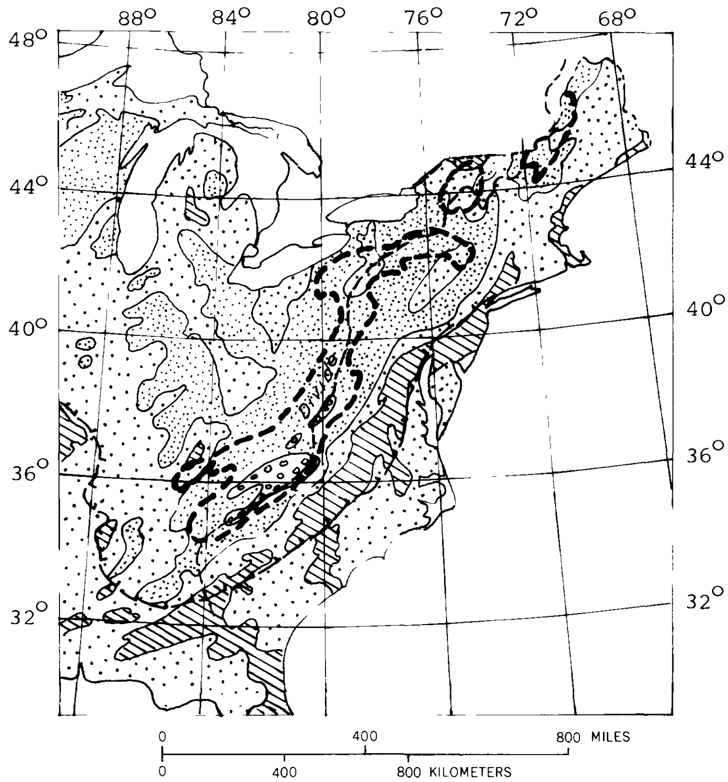
The data indicate that erosion rates in Eastern United States, both modern and Cenozoic in age, are, on the average, low. For large areas, they average as much as about 4 X 10^{-2} mm per year. Judging by two anomalously low rates estimated by more detailed analysis of very small watersheds, the true rates may be even lower. All these rates are much lower than uplift rates at the center of large areas like the Canandian Shield, where the rate at the center is 20 mm per year. However, uplift rates depend on the size of the area being deformed, so it is more reasonable to compare tilt rates than absolute uplift rates. If erosion rates are converted to or are considered the inverse of tilt rates, then a large discrepancy remains, amounting to at least one order of magnitude. It thus appears that the rates of deformation now observed cannot have been sustained for long periods. The rates of rebound after the removal of the ice are consistent, however, with measured tilt rates. Rebound can take place rapidly, with only a short lag time between removal of a load and compensation by uplift. Much of the data contained in repeated geodetic traverses, however, is not readily explained. For example, the tilt downward south of Atlanta, Ga., involving a 100 km traverse (table 3, no. 5 if sustained for 100,000 years would result in a depression 550 m deep. No depression exists at this locality, and the topography is a normal erosional topography resembling its surroundings.

Development of the Appalachian Highlands in late geologic time

The present topography of the Appalachian Highlands is closely adjusted to the rocks. However, the available evidence shows that the highlands have undergone warping and minor faulting in late geologic time. No positive evidence has yet been found that the bedrock at the surface has been broken by differential movements of any great magnitude since at least Cretaceous time. Though uplift took place, it must have involved broad warping of the crust. Data from the Coastal Plain and Continental Shelf indicate that though the tilting of this area has probably been continuous, it was particularly rapid in early Cretaceous and Miocene to Pliocene time. Assuming these conditions, the present topography could have been derived by normal erosional processes.

The outlines of the present landscape began to form some time in the middle Mesozoic after the breakup of the continent and the establishment of a drainage system toward the opening Atlantic. The first coarse clastic sediments deposited on the Coastal Plain and Continental Shelf are Early Cretaceous in age. The outlines of the hinterlands from which this material came are not known except by inference. The material deposited on the Coastal Plain contains elements from both the Piedmont and Coastal Plain (Glaser, 1969). As the coarser gravel contains a high proportion of quartzite and finer gravel contains a significant amount, at least some of the material probably came from the Appalachians west of the Blue Ridge. On the other hand, the heavy minerals and the paucity of chert indicate that, in general, the greatest proportion of material probably came from the Piedmont, perhaps indicating that the relief on the Piedmont was high at the time.

Geomorphic evidence concerning the limits of the initial drainage system and the position of the divide has been controversial, and much has been written on the subject (Thornbury, 1965, p. 82). Judson (1975) noted that the continental drainage divide corresponds to a gravity low in the central and southern Appalachians (fig. 6). The meaning of the gravity low is not understood, but the low may relate either to a thick crust or to the thick sequence of low-density rocks in the eastern part of the Appalachian basin. In any case, the correspondence of the low to the drainage divide is consistent with the idea that the divide has had a stable position for a long time. The correspondence of the divide to the low is lost north of lat 40°. In a general way, low anomalies do tend to correspond to high topographic areas, though, in detail, this correspondence



EXPLANATION

-120 to -80	-80 to -40	-40 to 0	0 to +50

Bouguer anomalies in milligals

Subenvelope contour at 1000 feet (305 m)

Inner edge of Coastal Plain

Figure 6.--Map of Eastern United States showing Bouguer gravity anomalies. Data from Defense Mapping Agency, Aerospace Center, St. Louis, Mo.

breaks down in the Adirondacks and New England (fig. 6).

During the Cretaceous, base level of erosion was established at about its present position or higher; it could have varied little more than 300 m in altitude since then. During the Early Cretaceous, when large amounts of coarse sediments were poured onto the Continental Shelf, derived largely from crystalline terrane, base level was lower than it is at present. By Late Cretaceous time, the rate of sedimentation had slowed, and base level rose so that the sea probably covered some of what is now the Piedmont. In Oligocene time, at least in the north, sea level again was lowered. In Miocene time, base level was generally lowered, though in some areas the sea was higher than now. The positions of base level were never parallel to those at present, as certain areas like the Cape Fear and South Jersey uplift tended to be higher than others (Owens, 1970), and there has been a general down tilt of the shelf to the northeast.

It is assumed that continuing uplift as well as erosion inland were responsible for the high rate of sedimentation on the Continental Shelf. The uplift was in the form of a broad arch, its axis centered somewhere near the present continental divide, as the absolute rate of uplift relative to base level must have increased inland. Thus, topographic relief that formed as a result of uplift would have been higher inland near the crest of the arch, especially in areas of resistant rocks. West of the Blue Ridge, where streams were eroding sedimentary rocks arranged in distinct and contrasting layers, the drainage system became adjusted to differences in the bedrock by multiple piracies, and local differences in relief were marked. This condition was especially true along the Cambrian and Ordovician belt in which solution was a major factor in the erosion mechanism. In the crystalline rock areas of the Blue Ridge and Piedmont, some rocks, such as the Ocoee Supergroup, felsitic volcanic rocks, or the quartz kyanite rocks of the Kings Mountain belt had great resistance to erosion. As Flint (1963) showed in southern Connecticut, rocks having high quartz content tend to form high relief. Massive, coarsely crystalline rocks tend to be more resistant than sheared or foliated rocks. When erosion was rapid, differences in relief related to rock type would have been large. Under conditions of slower erosion in the crystalline-rock areas because of the polymineralic nature of most of the rocks, a residue of saprolite would collect at the surface. When the residual mantle became generally thick, differences in relief were largely eliminated over broad areas.

At present in the Piedmont and Blue Ridge, saprolite underlies most areas that have low relief. It may be more than 300 feet thick, though it averages about 60 feet (Hack, 1976). In areas of high relief it is thin, or lacking, or it is found mostly in valley bottoms.

The shape of the domal uplift that produced the highlands is a significant problem. The steep gravity gradient that parallels the Blue Ridge front is suggestive of different crustal blocks that may separate areas having two rates of uplift. On the other hand, as shown by figure 5, the gradient loses its character in New England north of 40°, yet great contrasts in topographic relief are found in New England. An analysis by Best and others (1973) based on modeling of various assumed crustal conditions indicates that in North Carolina at about lat 35° 45', the steep gradient is related to high-density rocks in the Carolina slate belt at high levels in the crust.

Although the argument for a tectonic break near the east foot of the Blue Ridge is strong, the present topography could have resulted from a warping over the axis of the Appalachian Highlands. If this were true, the Blue Ridge escarpment south of Roanoke would have been tilted eastward (on the southeast side of the upwarp), and the escarpment could have reached its present position by retreat.

The gentle gradient and presumed rate of tilt of the basement just east of the Fall Zone may provide a clue. At the North Carolina-Virginia boundary, the gradient of the basement is only about 0.005. If projected inland as far as the Blue Ridge, the top of the basement would reach an altitude of only about 1,200 m, almost equal to the crest of the present mountain range. We have reason to believe, however, from estimated erosion and uplift rates, that the middle Cretaceous crest of the range would have been at about 4,000 m relative to present sea level. Therefore, if faulting is ruled out, the warping of the Piedmont must have taken place at a higher rate inland toward the Blue Ridge than at the Fall Zone. Farther north in the Washington area where the Piedmont narrows, the warping of the basement at the Fall Zone is steeper and may account for much of the uplift.

Conclusion

Analysis of the landscape, the major concern of this study, indicates that the topography is quite varied in altitude and relief, but this variation is in general related to differential erosion of the bedrock rather than to differential uplift. Even the Adirondack Mountains, which have the form of an uplift, are girdled by a wide belt of nonresistant rock. Much of the Appalachian Highlands region, however, can be interpreted as an upwarp of large proportions, but the entire highland cannot be explained that simply. For example, differential uplift, or warping, must be invoked to explain the sharp break between the Blue Ridge and Piedmont. This is attested to by the gentle slope of the Piedmont at its margin along the Fall Zone and by the lack of large-scale faulting along that zone. The Blue Ridge escarpment, at least in the south, cannot be explained by differential erosion alone. Thus, the up-arching of the Appalachian Highlands must have been differential, and it could have been accompanied by small-scale movements along fault zones parallel to the structural grain.

Data from a variety of sources indicate that the Earth's crust probably responds quickly to loading or unloading and that erosion rates and uplift rates are probably mutually dependent. The effect of a high rate of uplift is to increase erosion rates gradually eventually producing topography of greater relief. The converse would also be true, that decreased erosion rates would be correlated with lower relief. Local variations in the relief, even of great magnitude, are related to rock control and in places to differences in uplift rates as well, thus obscuring the outlines of tectonic blocks that may exist.

Data obtained from repeated geodetic traverses indicate rapid rates of tilt in many areas. These rapid rates are matched by similar rates along ancient water planes in deglaciated regions. Clearly, rapid changes in elevation or rates of tilt can and do occur locally. Nevertheless, many of those observed cannot have been sustained for long periods of time. The erosion rates that have been estimated are an order of magnitude or more lower, and if the high rates of tilt or uplift were long sustained, they would result in much larger topographic features than are observed.

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