Geomorphology of the Shenandoah Valley Virginia and West Virginia and Origin of the Residual Ore Deposits

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Geomorphology of the Shenandoah Valley Virginia and West Virginia and Origin of the Residual Ore Deposits

By JOHN T. HACK

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A geomorphological study in which it is assumed that the erosion and downwasting of the central Appalachians were continuous and uninterrupted by periods of baseleveling

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GEOMORPHOLOGY OF THE SHENANDOAH VALLEY, VIRGINIA AND WEST VIRGINIA, AND ORIGIN OF THE RESIDUAL ORE DEPOSITS

By John T. Hack

ABSTRACT

The Shenandoah Valley is an example of a mountain landscape that has been formed by erosion during a long interval of geologic time and that has reached a condition of dynamic equilibrium in which the adjustment between the landforms and the rocks beneath is nearly complete. The topography of the area and many aspects of its geology have previously been explained according to the theory of the geographic cycle, in which the mountains are presumed to have been eroded or partly eroded to base level several times in the geologic past. The concept of equilibrium, however, and the adjustment of slope to bedrock conditions through continuous weathering and erosion provides an alternate explanation that closely fits the distribution of landforms and surficial deposits.

On the southeast side of the valley, the Blue Ridge Mountains are underlain by igneous rocks, the most resistant of which are metabasalts of the Catoctin Formation of Precambrian(?) age. The metabasalt thins to the northeast to only about 50 feet; in the northeast it is not a ridge-making rock. The foothills of the Blue Ridge are underlain by clastic rocks of Early Cambrian(?) age, including thick quartzites that form ranges of high hills. To the northeast the proportion of shale and phyllite increases. The main lowland areas of the Shenandoah Valley are underlain by a thick sequence of limestones, dolomites, and shales of Early Cambrian to Late Ordovician age. In this sequence some formations, such as the Athens and Chambersburg Limestones and the Martinsburg Shale, consistently form the lowest areas having the least relief. The more hilly areas are characteristically underlain by other rocks, especially the Beekmantown Dolomite, which generally contains massive chert beds. Highlands on the west side of the valley and in Massanutten Mountain are underlain by sandstones and quartzites of Silurian to Mississippian age.

About 15 percent of the Shenandoah Valley is occupied by alluvial terraces and flood plains. The rest is hilly and mountainous, almost entirely in slope, and is graded by erosional processes. Such topography consists of a skeletal framework of streams separated by interfluves, crudely prismatic in shape, whose crests parallel the streams. In areas underlain by resistant rocks, the stream profiles are steep, whereas in shale and limestone they are as much as seven times gentler. Because of the steeper slopes required to transport the resistant rocks, streams in regions of resistant rock have a greater fall than do streams in soft-rock areas, a fact that accounts for the mountainous topography. Interfluves are also steeper and higher in the resistant rock areas and have fewer concavities. Less contrast appears, however, in the interfluves than in the stream valleys. Measurement of the local relief at various places within the Shenandoah Valley shows that the relation between the elements of topographic form and the kind of bedrock is consistent, and the local relief in any given kind of rock is the same. Major topographic features like the Potomac River also show an adjustment to the bedrock.

The surficial deposits consist of residual deposits, colluvium, and alluvium. The residual deposits and colluvium are closely related in origin to the rocks on which they rest; the alluvial deposits are distributed close to or downstream from the rocks that are their source. In the mountain areas thick residuum is found only on rocks that contain components subject to differential chemical and mechanical weathering. Granitic rocks of the Blue Ridge are commonly covered by thick residuum protected from erosion by a thin mantle of fresh core stones. On the other side of the Shenandoah Valley, shales interbedded with thin sandstones have a cover of residuum protected by a blanket of sandstone flags. Quartzite and greenstone areas, on the other hand, are characterized by many cliffy slopes and thin rocky soil.

Studies were made of the residual accumulations of boulders known as scree that are especially common on the quartzite slopes of the Blue Ridge. One kind of scree rests on or immediately below the rock from which it is derived; it moves, if at all, through the action of gravity. Another kind occurs on slopes far below the source rock, generally in hollows, and shows evidence of having been moved by intense floods caused by concentrated runoff. Essential conditions for the formation of scree are rugged relief, very resistant rocks, and a climate cold enough for frost riving. An intensely cold periglacial climate is not essential, but the formation of scree was probably more rapid in the cold periods of the Pleistocene than at present.

In the lowland areas residuum is thin or absent on the carbonate rocks that do not yield coarse residues on weathering. Such rocks include nearly pure limestone like the Mosheim, shaly limestone like the Athens, and calcareous shale like the Martinsburg. Residuum occurs in thicknesses of as much as 100 feet and more on carbonate rocks that yield coarse residues, such as the cherty dolomite in the Beekmantown Dolomite and the limestone interbedded with sandstone in the Conococheague Limestone. Residuum also occurs beneath alluvial deposits wherever they overlie shale or carbonate rocks. The residuum probably forms during the continuous erosion of the region and collects on the hill slopes wherever it is protected from erosion by a lag deposit of the coarse residues of the bedrock.

The alluvial deposits are mainly typical stream deposits consisting of coarse gravel at the base and sands, silts, and clays near the surface. The essential requirement for the formation of an area of alluvium is a pronounced difference in the resistance of the bedrock in different parts of a drainage basin; the distribution of both modern alluvium and the terraces is closely related to the more resistant rocks. Alluvial-terrace complexes occur on the carbonate-rock lowland wherever a large stream issues from the mountains, and the size of the terrace complex is proportional to the source area in the resistant rocks.

The deposition of alluvium is probably a continuous process. Alluvium accumulates in flood plains and fans in soft-rock areas at about the same rate as it is removed by erosion from the older terraces. The rates of accumulation and erosion of the alluvium may have varied in the past.

Solution is an important erosional process in the carbonate-rock lowlands of the Shenandoah Valley. Some solution goes on in the rocks at depths of thousands of feet below the land surface, but processes at or near the surface are responsible for most of the erosion. Solution is apparently favored both by synclinal structure and the proximity of large streams, which tend to promote concentrated lateral flow of water underground. In some places, especially at the foot of the Blue Ridge, solution goes on beneath a thick cover of alluvial and residual deposits.

Analysis of the sediment records of the South Fork, Shenandoah River, at Front Royal, Va., indicates that considerably more material is transported in suspension than in solution. As almost half of the drainage basin is in carbonate rocks, a very large proportion of the material eroded in the Shenandoah Valley must come from the mountain areas underlain by clastic rocks. The average rate of lowering of the surface of the area is estimated at 0.00023 feet per year, or 230 feet per million years.

Deposits of supergene oxides of manganese and iron are abundant in the Shenandoah Valley and have been mined for more than 100 years. The principal area of mines and prospects lies at the northwest foot of the Blue Ridge, where the oxides occur in residual clay resting on the Tomstown and Waynesboro Formations. The deposits are unrelated to any particular altitude or erosion surface; however, they occur at any altitude at which the weathered surface of the Tomstown Dolomite is found, and manganese has been mined from residuum 100 feet below the level of a large stream. The origin and characteristics of the deposits may be explained by processes similar to those acting today. The oxide ores, derived originally from minerals once thinly disseminated in certain beds in the Waynesboro and Tomstown Formations, are protected from erosion by the talus and alluvium that is washed over or around them from the Blue

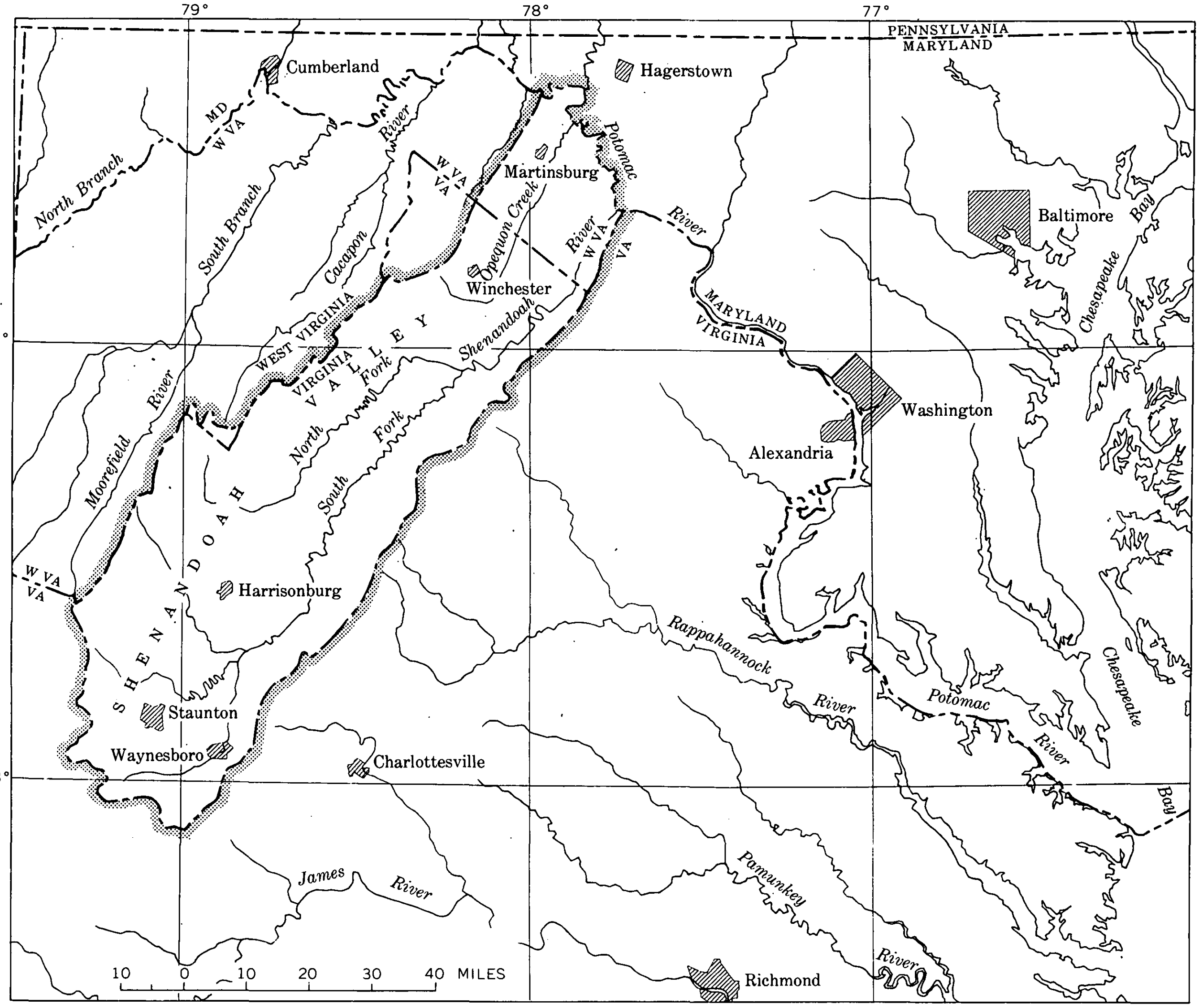

FIGURE 1.—Location of the Shenandoah Valley in Virginia and West Virginia.

Ridge. Chemical-equilibrium conditions in the waters that circulate in the residuum are such that the iron and manganese cannot escape in solution, though some migration within the residuum is possible.

INTRODUCTION

The Shenandoah Valley is an elongate area about 140 miles long that drains into the Potomac River (fig. 1). It lies between the Blue Ridge Mountains on the southeast and the North and Shenandoah Mountains on the northwest. In local usage as well as in this report, the area includes the drainage basin of Opequon Creek, a tributary of the Potomac River that parallels the Shenandoah for about 40 miles. The Shenandoah Valley, or as it is often called in Virginia, "The Valley," is a segment of a long and fertile lowland or trough, underlain by Cambrian and Ordovician limestone and shale, that extends for hundreds of miles along the axis of the Appalachian Highlands. It separates the Blue Ridge province from the main part of the Valley and Ridge province and has long been a main route of migration and travel to the west and southwest, as well as one of the country's rich agricultural areas.

This report is an interpretation of the surficial deposits and landforms of the Shenandoah Valley. The area was selected for studies of geomorphic processes and landform development in an orogenically stable region that has considerable variety of geologic and topographic features. Fieldwork began in the summer of 1952 and continued intermittently until the spring of 1958. About 3 months were spent in fieldwork each year. In 1952–54, collateral investigations were made by Dorothy Carroll and John Hathaway of the U.S. Geological Survey of the source, weathering, and transport of heavy minerals and of the clays in the rock, soil, and alluvium (Carroll, 1959; Carroll and Hathaway, 1954).

The writer's work was divided into phases, partial results of which have already been published. The first phase (1952–53) was a study of the factors that determine the form of the stream profile and its relation to geology (Hack, 1957). The second phase (1954–56) was a study of the processes that influence the forms of interstream slopes in mountain areas. In this study the writer was fortunate in being able to collaborate with Dr. John C. Goodlett, a forest biologist, at that time Research Associate of the Harvard Forest, Petersham, Mass. A study of the forest composition of a small area in the Shenandoah Mountains and its relation to geomorphic forms and processes was undertaken (Hack and Goodlett, 1960). During 1956–58 the writer made a reconnaissance study of both the residual and transported surface deposits of the entire Shenandoah Valley. This study provided much of the data presented herein. In 1957 and 1958 the writer was joined by Paul Ruane of the Geological Survey in a study of the talus and scree deposits in the Blue Ridge. A study of the meanders of the North Fork, Shenandoah River, was also made at this time in collaboration with R. S. Young, then of the Virginia Division of Geology (Hack and Young, 1959). L. M. Durloo, Jr., of the National Speleological Society, very generously spent several days showing the writer features of limestone caverns and invited the writer to work with him in mapping some caverns in Page County, Va. (Hack and Durloo, 1962). The help of two field assistants, Charles A. Ferriter, Jr., in 1952 and 1953 and Karl Funkhauser in 1954, is acknowledged with gratitude. Many other colleagues visited the writer in the field and made valuable contributions by their suggestions and critical discussion. The writer is particularly grateful for the generous amount of time given by Dorothy Carroll, C. S. Denny, John Hathaway, M. M. Knechtel, R. B. Neuman, and M. G. Wolman.

PREVIOUS INTERPRETATIONS OF THE GEOMORPHOLOGY

The geomorphology of the Shenandoah Valley has never been intensively studied. The classic geomorphic work was done farther north in Pennsylvania and New Jersey, where W. M. Davis conceived his cyclic theory of the development of Appalachian landforms (Davis, 1889, 1890a and b). Soon after this theory was proposed, it was applied to an interpretation of the topographic forms of the southern Appalachians, including the Shenandoah Valley. Hayes and Campbell (1894) and Keith (1896) recognized remnants of two ancient peneplains in the region of the Shenandoah Valley: a Cretaceous peneplain corresponding to the surface of high ridges that flank the valley and a Tertiary peneplain corresponding to the general level of the summits of the limestone hills in the Shenandoah Valley proper. Low gaps in the Blue Ridge and other high ridges of the region were interpreted by Watson and Cline (1913) as wind gaps, the remains of water gaps formed during erosion of the Cretaceous peneplain.

Stose believed that four peneplain levels existed and presented a diagrammatic cross section of the valley showing the accordance of summits at the various levels (Stose and Miser, 1922, p. 20). F. J. Wright (1934) believed that valid evidence exists for only two peneplains in the Shenandoah Valley region. The higher one corresponds to the general level of the summit of Massanutten Mountain and the ridges that flank the valley and is correlated by him with the Schooley peneplain of New Jersey and Pennsylvania. The lower erosion level, corresponding to the limestone hills, is correlated with the Harrisburg peneplain of Pennsylvania. Variations in the altitude of the peneplains were believed by Wright to be due to warping.

Hewett (1917) applied the ideas of this school of geomorphology to an interpretation of the origin of manganese deposits that occur in residual clays along the west foot of the Blue Ridge. He inferred that the manganese was concentrated during the rock weathering while the valley floor (Harrisburg) peneplain was being formed. King (1949, 1950), in a detailed study of the Elkton area in southeastern Shenandoah Valley, found extensive gravel deposits on the floor of the valley. The gravel overlies clayey manganese-bearing residuum. King believed that much of the residuum formed in Tertiary time on the Harrisburg peneplain and was preserved by gravel deposited during Pleistocene time.

Some geologists working in the Shenandoah Valley have presented evidence that does not support the peneplain concept. Thompson (1941) analyzed the topography of a large area in the southern half of the Shenandoah Valley and in the region to the west and concluded that the altitudes of the mountain ridges were too varied to be remnants of a peneplain and that their heights showed a close correlation with the character of the bedrock. He believed that no evidence in the region supports the concept of a summit or Schooley peneplain and that little evidence of a valley-floor or Harrisburg peneplain exists. Present topography was explained as the result of progressive stream piracy and local stream superposition, independent of peneplanation. He inferred that the original drainage divide was near the Blue Ridge and that the divide gradually moved westward as erosion progressed downward.

Edmundson (1940) demonstrated a close correlation between the topographic form and geologic structure of Little North Mountain, the ridge that borders the Shenandoah Valley on the northeast. He concluded that the altitude of this ridge and the position of the wind gaps are controlled by the width of the belt of outcrop of resistant sandstone. He further showed that hills on the valley floor in the limestone area are also controlled at least in part by the character of the bedrock (Edmundson, 1939).

THESIS OF THIS REPORT

The validity of the geographic cycle and peneplain concept has always been a subject of controversy. One of the principal weaknesses of the theory is that it requires separate episodes of uplift, separated by long periods of stability. Furthermore, the uplifts must be so uniform that individual straths supposedly produced during a cycle of uplift and stability can be traced for hundreds of miles and vary in elevation within only a few hundred feet. Alternate theories of Appalachian geomorphic development, such as the progressive stream piracy theory of Thompson (1939) or the theory of Permian consequent drainage (Meyerhoff and Olmstead, 1936), are preferred by some geologists, including the writer.

Within the Shenandoah Valley area itself, evidence of a summit or Schooley peneplain is wholly lacking. No ancient deposits on the ridge tops suggest an unconformity, and flat areas of any appreciable extent can be accounted for by local bedrock structures. As Thompson (1941) showed, even a general accordance of summit levels is lacking. When examined in detail, the ridges vary so much in height that if a summit or near-summit peneplain ever existed, no remnants of it remain. Careful investigators like Stose (Stose and Miser, 1922) and Wright (1934) must have postulated a summit peneplain only because they were convinced by evidence and arguments borrowed from other areas.

Although no vestige of the Schooley peneplain occurs in the Shenandoah Valley, the problem of the Harrisburg or valley-floor peneplain is a different one and must be considered separately. In the Elkton area, King (1949, 1950) interpreted the evidence to indicate the existence of an extensive partial peneplain or strath during Tertiary time at an altitude several hundred feet above the present streams. He described a deeply weathered mantle, referred to as residuum, that covers the hill slopes near the foot of the Blue Ridge. The residuum is overlain in places by a complex sequence of gravels, some of Pleistocene age. These relations suggest formation of a deeply weathered mantle during a long period of base-level stability followed by a period of gravel deposition that provided a protective cover for the residual deposits.

The principal subject of this report is an alternate explanation of the formation of landforms and surficial deposits. This alternative has been called the equilibrium concept of landscape (Hack, 1960a). No attempt is made to disprove Davis's concept of the geographic cycle and the peneplain. To do so is unnecessary because, although the idea of the geographic cycle has been useful and its validity has been accepted by many geologists, it is only a concept and cannot be regarded as absolute. The purpose of the report is to demonstrate that the equilibrium concept accounts as well or better for the features that have been mapped and described.

The entire landscape could have evolved through long-continued erosion of a thick sequence of rocks, during which an approximately balanced condition of the slopes was maintained throughout. The landforms and surficial deposits have a complex but close relation to the geologic structure, and as both forms and deposits can be accounted for by this concept, the necessity to postulate a peneplain is eliminated. If periods of uplift and dissection alternated with periods of stability and planation, they left no remains in the deposits or topographic forms. A generalized map of the bedrock geology of the Shenandoah Valley compiled from published sources (pl. 1) and a reconnaissance map of the surficial deposits (pl. 2) are included.

THE EQUILIBRIUM CONCEPT OF LANDSCAPE

The application of the principle of dynamic equilibrium to the interpretation of landscape is inherited from Gilbert, from his classic description of the Henry Mountains (1877, 1880), and it is analogous to the application of the equilibrium principle in the analysis of soils advocated by Nikiforoff (1942, 1959). As applied to landscape, the principle of dynamic equilibrium states that when in equilibrium a landscape may be considered a part of an open system in a steady state of balance in which every slope and every form is adjusted to every other. Changes in topographic form take place as equilibrium conditions change, but no particular cycle or succession of changes occurs through which the forms inevitably evolve, as was assumed by Davis and most later workers in geomorphology (Hack, 1960a). Differences in form from place to place are explained by differences in the bedrock or in the processes acting on the bedrock. Changes that take place through time are a consequence of climatic or diastrophic changes in the environment or of changes in the pattern and structure of the bedrock exposed as the erosion surface is lowered.

In his Henry Mountain report, Gilbert (1880, p. 93–150) listed several fundamental laws or principles regarding the action of erosive processes: (1) other things being equal, erosion is most rapid where the rock offers the least resistance, (2) other things being equal, steep slopes erode faster than gentle ones, (3) transportation of eroded material is affected by slope or declivity to a degree greater than a simple ratio, (4) transportation is also affected by quantity of transporting water to a degree greater than a simple ratio. The idea of equilibrium is introduced with the statement that a stream tends to equalize its work in all parts of its course. This follows from the laws of erosion just stated. If the quantity of water remains the same in a downstream direction and the amount of material is the same, then the stream smooths its channel and declivity becomes the same throughout the course. However, because the quantity of water does not remain the same downstream but increases as tributaries join, the slope must be modified; and when an equilibrium of action is reached, the declivity of the main stream will be less than that of its parts. This idea is expressed in the form of a law: "Ceteris paribus, declivity bears an inverse relation to quantity of water" (Gilbert, 1880, p. 108).

Gilbert (1880, p. 109–110) believed that erosional forms depend on the interaction of two principles: the law of structure and the law of divides. These laws are the basis for a system of landscape analysis. Since erosion is most rapid where resistance is least, hard rocks are left prominent. The difference is increased as erosion proceeds until an equilibrium of action is achieved through the law of declivities: "When the ratio of erosive action as dependent on declivities becomes equal to the ratio of resistance as dependent on rock character, there is an equality of action" (Gilbert, 1880, p. 110). The corollary states that when there is equality of action then there must be differences in slope because of differences in rock character. The principle accounts, as Gilbert stated, for the infinite variety and character of topographic forms and reliefs. The law of divides follows from the principle that declivity bears an inverse relation to quantity of water flowing down valley and states simply that slopes of stream channels are in general steeper the nearer they are to divides.

Gilbert noted complexities in the application of these ideas and pointed out, for example, that climate has an important influence on forms. He noted the regular arrangement of stream channels and divides in a typical herringbone pattern and the occurrence of hopper-shaped concavities at the heads of the channels. He recognized that it was necessary to have a certain minimum area of overland flow in order to produce a channel, and thus conceived an idea elaborated later by Horton (1945). Gilbert was apparently puzzled by convex-upward forms at the crests of the interfluves and did not explain them in his Henry Mountain report. His analysis in that report dealt almost exclusively with the stream channels and the forms related to them.

Many years later he explained in a very short paper (Gilbert, 1909) the convex form of the interfluves and noted that the processes in the area of overland flow were different from those in channels. This short paper also summarized his earlier concepts of stream action and presented a precise description of a landscape in which equality of action has been achieved. It stated that the landscape may be divided into two domains: The first is the domain of stream sculpture, represented by channels in which the slopes are concave upward because the transporting power of a stream per unit of volume increases not only with the volume but also with the slope. The stream therefore adjusts slope to volume in such a way as to equalize its work of transportation in different parts. The other domain (the interfluves) is that of creep, in which the slopes are mostly convex because the force impelling motion is gravity, which depends for its effectiveness on slope. The slope must be greatest away from the crest in order to provide the force to transport a greater amount of material in the same unit of time.

In the Henry Mountain report, Gilbert also discussed the landforms produced at the margin of areas of resistant rock where streams issued from mountains onto lowlands underlain by softer rocks. Because the banks of the streams in soft rocks are less resistant than the coarse cobbles carried from the hard rock areas and because the slopes on soft rocks are less, the streams tend to migrate laterally and to deposit some of their load, thus producing broad piedmont alluvial plains. The details of this principle have been modified by later work in the Henry Mountains, and the process has been carefully analyzed (Hunt and others, 1953, p. 189).

Gilbert's analysis of topography in the Henry Mountain report, if the 1909 explanation for the convexity of hilltops is added to it, provides the basis for a complete system of landscape interpretation in an erosionally graded terrain. This interpretation was never fully developed because the publication of the report was followed by the development of the concept of the geographic cycle by W. M. Davis (1889), a system of analysis that was enthusiastically received by geologists throughout the world. Gilbert did not dignify his concept with a name, and because geologists at the time were strongly oriented toward the theory of the geographic cycle, his ideas could not be examined independently. Gilbert's concept was obviously based on the principle of equality of action, which he equated with dynamic equilibrium (Gilbert, 1880, p. 117). The idea of base level, so important in the concept of the geographic cycle, is conspicuously absent as a factor influencing the development of landforms, but Gilbert noted (1880, p. 117) that declivity or slope originates in diastrophic action. The total relief determines the rate of erosion and thus the steepness of slopes. However, since every stream and every slope in a drainage basin is dependent on every other, the distribution and form of slopes when equilibrium is achieved are dependent on the laws of erosion.

Davis emphasized the concept of evolutionary change in his analysis of landforms. Chorley (1962) suggested that the concept in reality treats an eroding landscape as a closed system in which the energy in the system becomes less as erosion proceeds toward base level and finally in the last stages approaches zero. Whereas Gilbert emphasized structure as a cause of diversity in landscapes, Davis believed that the effect of structure was lessened and almost nullified as the evolution of topography proceeded beyond the stage of maturity. In the cycle concept, equilibrium is achieved at a certain stage when the streams are said to be graded. At this stage the streams are just able to transport the materials supplied to them by erosion of the slopes above, and no more. From this stage on there is only slow downward cutting. Downwasting is largely manifest by the weathering of the interfluves and lateral planation by the streams. Equilibrium might be achieved in the lower part of a drainage basin as it neared base level long before it worked headward into the upper part of a basin. Eventually, when the entire landscape is in equilibrium, it is in the stage of old age, and very little potential energy remains for further change.

Thus, Davis' idea of equilibrium is very different from Gilbert's "equality of action" and "interdependence." In Gilbert's analysis, equilibrium is achieved when all the slopes in the drainage basin are mutually adjusted to a common erosion rate. Time or stage has nothing to do with the ability of the stream to transport the eroded materials or with the amount of relief or ruggedness of the terrain. The idea of interdependence does not suppost that any stage is necessarily ever reached when the streams must have gradients so gentle that they are loaded to capacity. Equilibrium may exist while the streams are actively eroding their beds. In Gilbert's system the balance between load and capacity cited by Davis may be reached as a consequence of spatial factors. Deposition, for example, occurs where slope is reduced as a stream enters the ocean or passes from a mountain area to a lowland underlain by softer rocks.

Analysis of landscape in terms of the equilibrium concept can be said to be independent of time. This independence does not mean that evolution or change does not occur. That profound and important changes occur is quite evident, but when a landscape is close to equilibrium most traces of its past history are erased so that the forms and thin surficial deposits are explained in terms of the rocks and the environmental conditions that exist now or existed in the recent past. Relict features that formed under different past conditions are preserved only if complete equilibrium has not been reached.

Nikiforoff (1942) constructed a simple mathematical formula to illustrate and explain the equilibrium principle as it is applied to processes of soil formation. The same formula can be applied to many geologic processes. Let us assume that talus accumulates at the foot of a cliff. A certain amount is added every year by rockfalls. This amount we assume to be constant. A certain proportionate amount is removed each year by weathering, solution, creep, and other processes that reduce the volume of talus. The absolute amount lost by these means increases each year as the size of the accumulation increases. The talus accumulation will grow at the bottom of the cliff until the rate at which it is removed equals the rate of accumulation. Then the talus will be in a condition of dynamic equilibrium or steady state and the volume will remain the same. This idea can be expressed matematically:

Let S_n be the total amount of talus present at the end of n years;
A is the amount of talus that accumulates during 1 year;
r is the rate of removal of the talus, expressed as a decimal portion of the amount present;
n is the number of years.

At the end of 1 year,

$$S_1=(A-Ar)+A=A(1-r)+A.$$

At the end of 2 years,

$$S_2=(S_1-S_1r)+A, \text{ or } S_2=[A(1-r)-A(1-r)r]+A,$$

and so forth. The series can be transformed to the following general equation:

$$S_n=A\left[\frac{(1-r)\cdot[1-(1-r)^n]}{r}\right]+A.$$

Nikiforoff (1942, p. 852) said,

It is assumed that the amount, A, is the same each year throughout the period characterized by the stable condition of the environment and that r is constant as regards its relative value although its absolute value increases as a function of S_n until it becomes equal to the amount A. The limiting value of [the] equation . . . is as follows:

$$S_n=A\left(\frac{1-r}{r}\right)+A=\frac{A}{r}.$$

When the absolute amount represented by r equals A the value of S_n remains the same, and no further change occurs. The total amount of material present is then equal to the annual increment A divided by the proportion that is annually removed. The equation may be illustrated graphically as shown in figure 2. In this illustration the dashed line represents the growth of the talus accumulation as a certain amount is added and a

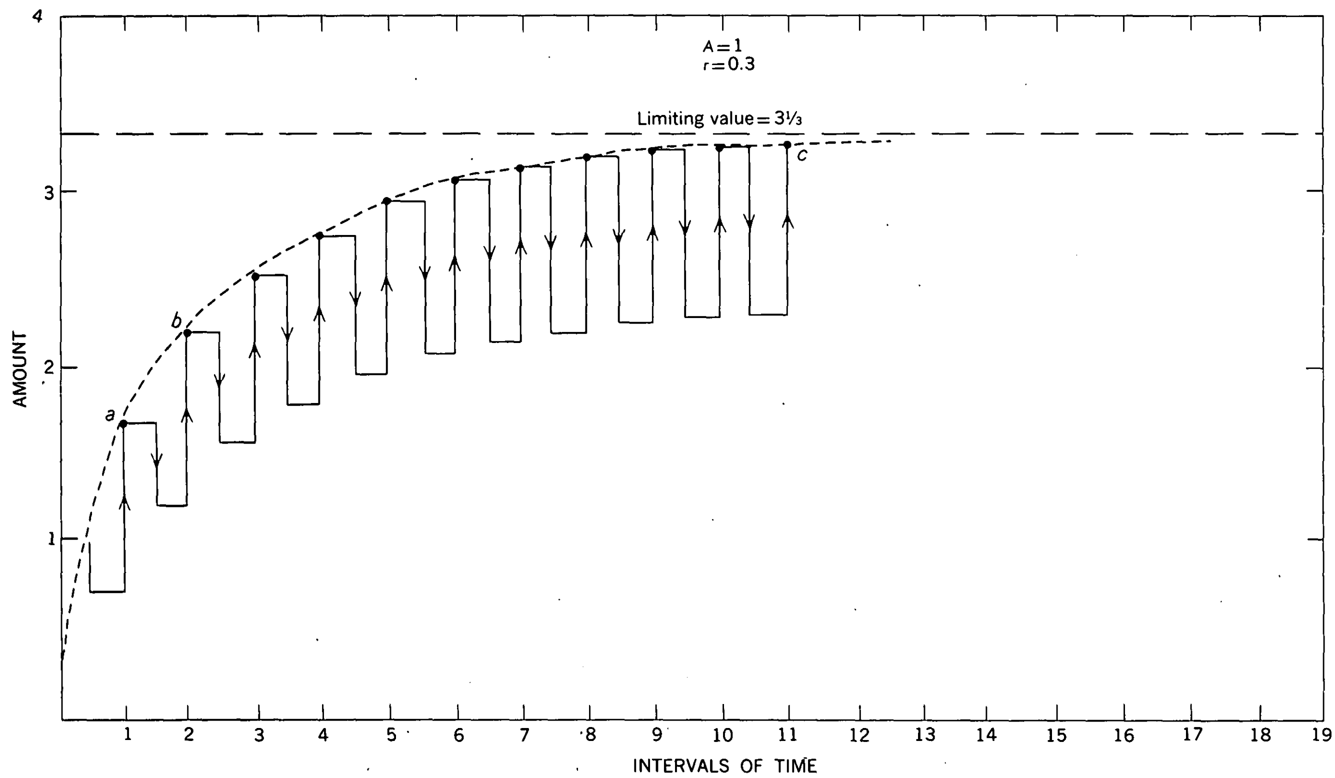

FIGURE 2.—Graph illustrating Nikiforoff's equation for the growth of a deposit toward a steady state in which two opposing processes are involved in the deposit's formation. See text for explanation.

certain amount removed during each interval of time. Starting with an amount A, by the end of the first unit of time, an amount equal to 0.3, or 30 percent, of the original A is removed, but an additional amount equal to A accumulates, bringing the total value of the accumulation to a. During the second time unit an amount equal to 30 percent of a is removed, and an amount equal to A is added, bringing the total to b. During the third time unit 30 percent of b is removed, and an amount A is again added and so on. At c the total amount present is very close to the limiting value, and the system is virtually in equilibrium. The limiting value is A/r, or $1/0.3$.

Many geological processes can be interpreted by use of the equilibrium principle. The volume or area of a piedmont alluvial fan, for example, is limited by the rate of erosion of the gravels that have accumulated in it, as well as by the amount deposited on it each year. How small the annual loss by erosion is makes no difference. Eventually a limit to the size of the fan must be reached, and as long as the environment does not change, no further change in size will occur. The application of this concept to the equilibrium of form assumes that differences in form exist because of differences in the relative effectiveness and kinds of processes acting on materials of different kinds and the function of a slope is to provide for the transportation of rock waste by one or more different processes.

Clearly, the concept of equilibrium is not in itself an explanation for the origin of landforms, deposits, or other phenomena. It is a way of viewing them, or a basis for analysis. In some situations, analyses by the erosion-cycle concept and by the equilibrium concept may seem almost identical and lead to the same conclusion. In other applications, the conclusions may be very different. For example, the erosion cycle concept is frequently elucidated using the example of a flat or nearly flat plain underlain by homogeneous material that is suddenly uplifted above a base level such as the sea and then remains static as it erodes. The manner in which the landscape is supposed to evolve is, of course, very familiar to most geologists. It passes through the stages of youth, maturity, and old age. In maturity the hilltops are rounded or convex upward, the stream profiles are concave upward, and the streams form flood plains. In old age the slopes are less steep and the valley floors are wider, relative to the interfluves, than they were before. They are both covered with a mantle of waste.

If a landscape such as the one in the example were analyzed in terms of the equilibrium concept, one can conclude that a somewhat similar evolution would occur, though the end product might not be quite the same. Gullies will evolve into canyonlike ravines having flat interfluves between them, as in the stage of youth. At this stage in its evolution, the entire landscape is not yet in equilibrium because remnants of the old flat surface are still preserved, and water flowing across the remnants does so on slopes that are not adjusted to the slopes in the ravines. Because the erosion rate is greater in the ravines than on the old land surface, the ravine system will extend itself until all the area is in slope and until the channels through which the water moves and the rounded interfluves are mutually adjusted. The landscape then resembles the hypothetical mature landscape of Davis and may be described as erosionally graded.

The general form of the area, though not the relief, will from this time on remain little changed. The potential energy in the system will be lowered as the average relief is lowered by erosion, and slope will reduce throughout the area. The interfluves and channel slopes will flatten until the area approaches the form of a plain. However, there is no reason to believe that the relative size of the interfluves and valleys will change or that the process of lateral planation will ever become more important. Nor is there any reason to suppose that the streams will at some time be unable to transport the waste products of erosion and, therefore, that the ability of the streams to transport material will be equalled by the material supplied. As the area is in equilibrium and the slopes are balanced throughout, they must always have sufficient grade to transport what is weathered. As the grades of streams are reduced, the rate of weathering and denudation must also be reduced because of the lesser potential energy available in any part of the system.

The difference between the two concepts seems small if applied to such a situation as has just been described; but if the forms of a complex area like the Appalachians are considered, the differences in the two concepts become very important and more evident. In the concept of the erosion cycle, a topography like the lowlands of the Shenandoah Valley, where the streams are actively downcutting and the hills rise a more or less even distance above them, is explained as a partial peneplain, or strath, that has been dissected and has reached the stage of early maturity. The more or less even ridge tops of the Blue Ridge and North Mountain are regarded as remnants of a still older peneplain preserved because of the superior resistance of the rocks in those areas. In the equilibrium concept, however, the entire topography is regarded as belonging to a single system assumed to be in equilibrium, in which the slopes and altitudes of the mountains are adjusted to the slopes of the lowlands and the entire area is down-wasting more or less at the same rate. The differences in form and relief are to be explained by the law of structure and are accounted for by differences in the bedrock.

THE EFFECT OF BASE LEVEL

Under ideal conditions in which the forms of the landscape, including the relief, remained in a perfect steady state, erosion would have to be exactly balanced by isostatic uplift. Under such conditions the generalization could be made that all parts of the landscape are downwasting at the same rate. If, however, the landscape should have a permanent or long-persisting base level, then the total relief must diminish through time, and the rates of downwasting would not be quite the same throughout the system. An area downstream near base level would erode at a somewhat slower rate than an area far upstream. Under such conditions the rate of downwasting on similar rocks would be proportional to the slope both in space and, at any one place, through time. Mountains and divides would downwaste faster than large trunk valleys.

For the Shenandoah Valley, this postulate may not apply. The gradient and size of the bed material transported indicate that the Potomac River near Harpers Ferry is as competent or more competent than the Shenandoah River. Furthermore, the relative erosive power of the Potomac increases before it reaches the Coastal Plain. Whether this condition has persisted for a long time in the past we have no way of knowing.

THE EFFECTS OF ENVIRONMENTAL CHANGE

The argument presented in the foregoing section makes it evident that if the environment remains stable long enough, all elements of the landscape must be in equilibrium, and the forms must be nearly static. This condition, of course, is never reached. Downwasting of an area would never be balanced by uplift so as to maintain exactly the same relief, and climatic factors would never be completely constant. Many changes take place, including major ones such as diastrophic movements of the earth's crust and minor ones such as climatic fluctuations of varying magnitudes. Therefore every area includes features that are in equilibrium and some that are not. Some equilibria are rapidly restored when disturbed, whereas others are restored only slowly. The concept of equilibrium can be applied only in relation to a specific system embracing a certain area and a certain span of time. For example, the bars and riffles in every stream achieve a condition of dynamic equilibrium quickly, but this condition is altered by every flood and rise of water. The fall of a tree into a small stream will change the equilibrium conditions in a small reach. Neither the short-term changes of the bars and riffles nor the fall of the tree have any significant relation to the concept of equilibrium of the stream valley as a whole, which has been achieved through a long span of time and involves processes acting over a large area.

The equilibrium concept can be used as a basis for analysis of a landscape only if the system being analyzed is in fact in or close to equilibrium. Thus the topography of a fault-block mountain that is actively rising could not be included in the same system as the topography of a range of hills on the other side of the active fault. A determination of when and where equilibrium can be assumed may be a problem. Whether a bar or segment of a stream channel is in equilibrium can be easily determined because day-to-day changes can be observed, and we can study and measure the evolution that takes place when the environment changes. If we deal with problems on this scale, the steady state and the interaction of environmental factors and materials are taken for granted. Larger systems, however, are a different problem, and we can by no means always determine whether evolution is taking place or whether a near steady state has been achieved. Major diastrophic movements probably occur at least occasionally almost everywhere, and one might fairly question whether the major elements of any landscape ever achieve perfect equilibrium of form.

The topographic elements in the central Appalachians, at least, are generally agreed to be erosional forms. Certainly no one would seriously suggest that any single mountain or range owes its form and altitude entirely to an uplift of the mountain itself. The problem then becomes one of deciding the scale of erosion in terms of both area and time of the graded erosional system with which we are dealing. This determination can probably be made only by trial. If a certain area is assumed to have forms that are mutually adjusted and if our analysis proves to be useful and explains all the facts that we can apply to it, then our basic assumption is probably justified. As a corollary, a lack of adjustment and the limits of the system may be discovered by this means.

Climatic changes present an especially difficult problem, for such changes have probably occurred frequently during the periods of time with which we must be concerned in any landscape analysis, and they must have had a pronounced effect on topographic forms (Gilbert, 1880, p. 111). The climatic effects of the cold stages of the Pleistocene are presumed to have been profound and even though the Shenandoah Valley is several hundred miles south of the margin of the Pleistocene ice sheet, the region must have been greatly affected. The landforms and deposits produced during the colder past are now out of equilibrium, and present processes are obliterating them or modifying them, if they have not already done so. Furthermore no topographic form or deposit can possibly be entirely a result of processes that acted in the remote past. Every part of every landscape is being affected by the environment today and is therefore either in equilibrium in the present environment or is approaching such an equilibrium.

When the new steady state is achieved, the forms or deposits of an earlier equilibrium may be completely erased. This fact is illustrated by the example of the talus accumulation cited on page 7. After the talus has accumulated to a certain volume and equilibrium is achieved, let us assume that the value of r is increased. In other words, the proportional rate at which the talus is eroded is increased. According to the formula of Nikiforoff, the volume of talus will then be reduced until a new equilibrium is reached, when A will again equal the absolute amount represented by r. The volume of the talus in the new steady state is entirely a function of A and r under the new conditions and has nothing to do with the fact that a greater amount of talus was once present.

Inherited forms, especially those produced in cold climates, have been the object of intensive study by many geologists. However, little study has been made of the processes active in temperate climates that tend to modify or obliterate the forms produced in the cold climates. Until the geomorphic processes of the temperate regions are better understood, a somewhat skeptical attitude is in order concerning the interpretation of surficial deposits and forms supposedly inherited from the Pleistocene. Furthermore, many forms and deposits probably have a complex origin, and their development required a span of time that may embrace several periods of climatic change. The fact that the environment undergoes constant change should be emphasized. Since we cannot see what actually happened in the past, we can only infer the causes and evolution of topographic forms through study of the processes, deposits, and forms that exist in the present. A knowledge of the present is essential to the interpretation of any stratigraphic sequence.

SUMMARY OF BEDROCK GEOLOGY

This discussion of bedrock geology is based on many published reports. The principal sources used are Butts (1940), Edmundson (1945), and King (1950). Specialized works of Spencer (1897), Reed (1955), Nickelsen (1956), Cooper and Cooper (1946), Neuman (1951), and Knechtel (1943) have also been consulted. Emphasis is placed here on the physical character and other attributes of the rock that might affect topographic form and on the distribution of surficial deposits. The general geology of the area, compiled in 1960, is shown on plate 1, in which the rocks are grouped in units that probably have like properties. Many minor structural features as well as the intrusive rocks of Triassic age, which occupy very small areas in the Shenandoah Valley, are not shown on this map.

ROCKS OF THE BLUE RIDGE

GRANODIORITE

The Precambrian basement complex in the Shenandoah Valley area consists of crystalline rocks that crop out as inliers in and along the crest of the Blue Ridge. These rocks are extensive in the areas east and southeast of Luray and Front Royal. The rock is mostly a medium- to coarse-grained hypersthene granodiorite composed of plagioclase, quartz, hypersthene, biotite, chlorite, magnetite, garnet, and minor amounts of epidote and albite. The rock is light greenish gray where fresh and brownish gray or white on weathered surfaces. Generally, the rock is only slightly foliated (Reed, 1955, p. 876). Rectilinear joints are common, but the density of joints varies considerably from place to place.

CATOCTIN GREENSTONE AND SWIFT RUN FORMATION

The granodiorite basement complex is overlain by a thick series of altered lava flows commonly called greenstone, or metabasalt. In places, the flows are separated by thin beds of sedimentary rock of pyroclastic origin. At the base of the lava flows and separating them from the granodiorite are thin arkosic quartzite and interbedded pyroclastic rocks that in places reach a total thickness of 100 feet. These rocks are known as the Swift Run Formation (Stose and Stose, 1946). This formation is too thin to have any geomorphic significance.

The Catoctin Greenstone is largely an altered basalt, commonly containing patches of epidote. In places, it is completely altered to this mineral. The other principal minerals of the Catoctin are albite, chlorite, actinolite, and sphene. Columnar jointing and other typical volcanic structures are preserved in places. The Catoctin is an important ridge maker in the Shenandoah Valley and is obviously a resistant rock. Its thickness is reported to be about 1,000 feet in the Elkton area (King, 1950) and more than 1,800 feet in the Luray area (Reed, 1955). It thins to the north, and at the Potomac River, where it lies east of the crest of the Blue Ridge and therefore outside of the Shenandoah Valley, it is 50 feet or less in thickness (Nickelsen, 1956, p. 245).

No fossils have been found in the Catoctin Greenstone. Some disagreement as to its age exists. If the Catoctin underlies the sedimentary Chilhowee Group unconformably, it is probably Precambrian; if it underlies the Chilhowee conformably, it may be Cambrian in age. The age of the Chilhowee Group is itself problematical, but it is generally regarded as Cambrian and Cambrian(?) (King, 1950, p. 14; 1964, p. 78).

CHILHOWEE GROUP

Above the Catoctin Greenstone, the rocks of the Shenandoah Valley are primarily sedimentary in origin. The basal units of the sedimentary sequence are quartzites, graywackes, and argillites of the Chilhowee Group that underlie either the western foothills of the Blue Ridge or the crest of the Blue Ridge for the entire length of the Shenandoah Valley. The group is commonly subdivided into four formations, all Early Cambrian(?) in age that are persistent throughout the area. The lowermost, or Loudoun Formation, is a red or purple tuffaceous slate generally less than 100 feet thick and is similar in character to the Swift Run Formation and to the thin sedimentary layers common in the Catotin Greenstone. The Loudoun is overlain by the Weverton Formation, which consists mostly of graywackes, conglomerates, ferruginous quartzites, and some argillaceous rocks. Above this is the nonresistant argillaceous Harpers Formation. The topmost formation of the group is the Antietam Quartzite, a quartz sandstone containing many beds of massive quartzite.

The total thickness of the Chilhowee Group appears to be rather uniform throughout the area, ranging from 3,000 to 3,500 feet. Detailed studies of areas in the Blue Ridge, however, indicate that the proportion of argillaceous rocks in the Chilhowee Group increases to the north (table 1). In the southern end of the Shenandoah Valley, sandy and quartzitic rocks predominate (Knechtel, 1943). In West Virginia near the Potomac River, argillaceous or phyllitic rocks predominate. The Harpers Formation is predominantly a siltstone at the south but is a phyllite at the north. The phyllite, according to Nickelsen (1956, p. 250), is composed of sericite and chlorite in addition to quartz grains that make up 20–50 percent of the rock. To the south, shaly beds are rarely observed in the outcrop, whereas to the north they are common, and the soil contains small shale fragments. The increase in thickness of the Harpers from south to north is accompanied by a decrease in thickness of both the Weverton and the Antietam. The massive quartzites that are so evident in the Antietam Quartzite to the south appear to be thinner near Harpers Ferry.

CARBONATE ROCKS OF EARLY CAMBRIAN AGE AT THE WEST FOOT OF THE BLUE RIDGE

Northwest of the foothill belt, the Blue Ridge is bordered by a wide belt of dolomite, limestone, calcareous sandstone, and shale of Early Cambrian age, known as the Tomstown Dolomite and the Waynesboro Formation. These rocks contrast markedly with the massive quartzites on which they rest. Outcrops are few, and the rocks are generally buried beneath a blanket 50–100 or more feet thick, composed of residuum and Tertiary and Quaternary gravel.

TABLE 1.—*Composition and thickness of formations in the Chilhowee Group*

[From southwest to northeast along the strike, in the Shenandoah Valley]

Formation	Big Levels area (Knechtel, 1943)	Elkton area (King, 1950)	Luray area (Reed, 1955)	Harpers Ferry area (Nickelsen, 1956)
Antietam Quartzite.	Quartzite and sandstone; contains scolithus tubes. Thickness: 725 ft.	White vitreous quartzite and buff or brown sandstone, in part calcareous. Thickness: 800 ft.	Vitreous orthoquartzite; contains scolithus tubes. Thickness unknown.	Siliceous sandstone, some quartzite, and calcareous sandstone. Thickness: 450 ft.
Harpers Formation.	Shale, fine-grained arkose, and quartzite. Thickness: 1,000 ft.	Fine-grained siltstone and sandstone; contains some thin beds of quartzite. Thickness: 900 ft.	Undivided: Ferruginous quartzite and graywacke grading upward into phyllite and argillite containing thin quartzite beds. Thickness: 1,900–2,500 ft.	Light greenish-gray phyllite; some silty and sandy beds. Thickness: 1,500–2,000 ft.
Weverton Formation.	Quartzite, sandstone conglomerate, and considerable shale. Thickness: 900–2,000 ft.	Conglomerate, felspathic quartzite, argillaceous shale, and ferruginous quartzite. Thickness: 1,000–1,600 ft.		Dark-gray quartzite and some phyllite; bluish-gray ferruginous quartzite. Thickness: 500 ft.
Loudoun Formation.	Not reported	Purple and red slate. Thickness: 100 ft.	Not reported	Dusky red-purple phyllite. Thickness: 50–100 ft.

TOMSTOWN DOLOMITE

The Tomstown Dolomite, which rests directly on the Antietam Quartzite, is of particular interest because it is probably the source of the thick residual clay, silt, and sand that contain commercial manganese deposits (King, 1950, p. 65). Outcrops of the Tomstown in the Shenandoah Valley are scarce, and our knowledge of the formation comes mainly from areas to the north and south. A section of the Tomstown Dolomite 800 feet thick is well exposed in a quarry along the Norfolk and Western Railway near Buchanan, about 38 miles southwest of the southern edge of the Shenandoah Valley (Edmundson, 1958, p. 82). Here the rock is a dark-gray dolomite containing very few impurities and only thin silty and shaly beds. The section, however, is not complete. In the area south of Waynesboro, mapped by Knechtel (1943), the Tomstown is not exposed in outcrop, but samples from shafts and wells consist of finely crystalline dolomite.

In the Elkton area, King (1950, p. 25) described an exposure in which 742 feet of beds probably belongs to the Tomstown Dolomite. This section includes 380 feet of gray dolomite and 360 feet of beds that are predominantly shale. Drill holes near manganese mines and prospects in the Elkton area have penetrated the Tomstown in many places; rocks penetrated have included dolomitic shale and argillaceous dolomite as well as nearly pure dolomite. King believed that the lower beds of Tomstown are sandy and are transitional with the Antietam Quartzite.

The most nearly complete exposures of the Tomstown are in the northern part of the Shenandoah Valley, where the outcrop belt is west of the Shenandoah River. In Clarke County, Va., the exposed part of the formation is fine- to medium-grained light- and dark-gray dolomite that contains thin silty interbeds in the lower part. Though no complete sections have been found, Edmundson (1958) estimated the probable thickness at 1,000–1,500 feet.

The writer examined exposures of the Tomstown on both banks of the Potomac River in West Virginia and Maryland, upstream from Harpers Ferry. In this area the outcrop belt of Tomstown is about 4,800 feet wide, and part of the formation may be repeated by folding. Rock is continuously exposed on the south bank of the Potomac River 1.3 miles upstream from Harpers Ferry in a cut on the Baltimore and Ohio Railroad. The dip is 63° SE. (probably overturned) at the eastern end of the exposure but increases to 90° at the western end. The exposure is 620 feet west of the nearest outcrop of Antietam Quartzite and therefore must be in the lower part of the Tomstown Dolomite. The exposure consists of 200 feet of thin-bedded gray fissile limestone overlain by 180 feet of thick-bedded dark-gray blocky dolomite containing small chert nodules. Laboratory analyses of samples of these rocks by Carroll (written commun., 1961) indicates that the limestone contains 6 percent by weight of silt and clay and that the cherty dolomite contains 4 percent. Several large but discontinuous outcrops of Tomstown occur along the strike about 5 miles to the northeast on the north bank of the Potomac River. These are in Washington County, Md., between Dargan and Antietam (Cloos, 1941). The Tomstown is not sufficiently well exposed here for the structure to be determined; so the proportion of rock types cannot be estimated. Impure fissile limestone beds, like the beds exposed at Harpers Ferry, crop out at two places. At one of these, 400–500 feet of such limestone is continuously exposed in addition to 75–100 feet of sandy dark-gray dolomite.

The scant evidence available within the Shenandoah Valley indicates that the Tomstown Dolomite probably varies along the strike and contains, as believed by King, sandy and silty beds that could produce thick bodies of silty and sandy residuum.

WAYNESBORO FORMATION

The Tomstown Dolomite is overlain by the Waynesboro Formation of Early Cambrian age, which is almost equally poorly exposed. It is a heterogeneous formation composed of maroon and greenish shaly mudstone, fine-grained sandstone, and impure limestone and dolomite that probably total about 2,000 feet in thickness (Edmundson, 1945, p. 13).

LIMESTONE AND DOLOMITE OF MIDDLE AND LATE CAMBRIAN AGE

The Waynesboro Formation is overlain by carbonate rocks that crop out over extensive areas on the floor of the Shenandoah Valley. The rocks are predominantly limestone and dolomite and contain some sandy limestones, calcareous sandstones, and siltstones. Chert is common. These rocks have been subdivided into two formations that, although mappable over large areas, are lithologically not very distinctive. They are not differentiated on the geologic map (pl. 1).

ELBROOK FORMATION

The Elbrook Dolomite of Middle and Late Cambrian age rests conformably on the Waynesboro Formation. It is composed predominantly of interbedded clayey limestone and impure dolomite and has an estimated thickness of about 2,000 feet. One of the distinctive features of the formation is that it is commonly thin bedded and slabby or platy. The rock contains considerable siliceous material, mostly in the form of silt. Thin interbeds of shale are common, and some beds contain lenses and nodules of gray chert. In the Elkton area, the formation is composed predominantly of thin-bedded dolomite but contains blue-gray limestone, especially in the upper part. Here King (1950, p. 33) estimated it to be 3,000 feet thick.

CONOCOCHEAGUE LIMESTONE

The Conococheague Limestone of Late Cambrian age differs from the overlying and underlying formations in that it contains much more limestone, which is generally light bluish gray in color. Dark-gray beds of dolomite are common, however, and in some places compose at least half the formation. The proportion of dolomite is very small in the northern part of the valley but is high in the southern part. The most distinctive features of the formation are interbedded layers of dolomitic and calcareous sandstone in which the sand grains are generally so well rounded that the layers look almost like oolite. These beds are dark gray and gray and where fresh are difficult to distinguish from limestone. They weather, however, to a dark brown and form a soft brown friable porous sandstone. Another characteristic feature is the presence of crinkly ribs formed by thin laminae of siliceous material that stand in relief on the weathered surface. Lenses and nodules of light-gray chert are fairly abundant at several horizons in the formation. Edmundson (1945, p. 14) estimated that the average thickness is 2,500 feet. King (1950, p. 34) believed that the thickness is 2,000 feet in the Elkton area. The proportion of insoluble material in the Conococheague Limestone is high and, except for the Waynesboro Formation, is probably higher than in any other of the predominantly carbonate rocks of the Shenandoah Valley.

BEEKMANTOWN DOLOMITE, INCLUDING THE CHEPULTEPEC DOLOMITE

The Beekmantown Dolomite of Early Ordovician age is one of the most extensive formations in the Shenandoah Valley and is of considerable geomorphological interest because it forms many high ridges and conical hills and because it contains many caverns and sinks. The lower part of the formation, or group as it is called in places, is generally differentiated from the rest of the rocks included under the name Beekmantown and is called the Chepultepec Dolomite, or in West Virginia and Maryland, the Stonehenge Limestone (Sando, 1957). This unit commonly contains 80–90 percent calcium carbonate (Edmundson, 1945, p. 15). It is 300–500 feet thick but apparently thickens to the north. In Maryland, it is mapped as a part of the Beekmantown Group (Sando, 1957, p. 18 and pl. 3) and is 800–900 feet thick. The remainder of the Beekmantown, above the Chepultepec, is described by Edmundson (1945, p. 15) as follows:

> It shows wide variations in composition and thickness. In Augusta and Rockingham Counties, the Beekmantown is composed chiefly of fine-grained, gray dolomite with minor amounts of bluish-gray limestone. Farther northeast along the strike limestone becomes more abundant, and in Frederick County probably comprises slightly more than half of the formation. Limestone also becomes more abundant southeastward across the strike of the belts; thus in Clarke County the Beekmantown is largely bluish-gray limestone with minor amounts of dolomite. In places the formation contains relatively thin beds of high calcium limestone intercalated in dolomitic limestone and dolomite. The layers of dolomite are commonly characterized by distinct intersecting furrows on weathered surfaces. The sampled units of Beekmantown dolomite contain about 30 to 41 percent magnesium carbonate. Chert, although generally not abundant, occurs locally at several horizons in the formation. Field observations indicate that the thickness of the Beekmantown increases northeastward, probably ranging from 2,000 feet in August County to about 3,000 feet in northern Frederick and Clarke Counties.

The Beekmantown, although not predominantly cherty, probably contains more chert than any of the other formations in the Valley. The chert has an important function in the development of the topography, as is discussed on page 48. The chert occurs in several varieties and at various horizons and is apparently far more abundant in the southwestern part of the Valley than in the northeastern part. In Augusta County the formation contains thick units that are covered with coarse boulders of gray spongy chert and that form ridges many miles long. The chert has been described by Sando (1957) in a study of the Beekmantown Group of Maryland just north of the Shenandoah Valley. The chert is common in the lower dolomitic part of the formation but is rare in the limestone-rich upper part. It is of three kinds: "Irregular chert" consists of irregular masses of ellipsoidal shape generally less than 6 inches long; its relations to structural features of the rock suggest that it formed or was introduced at some time after regional fracturing. "Nodular chert" consists of small pods or nodules arranged parallel to the bedding and in some outcrops appears to be in the form of thin perforated sheets; it is a replacement chert, but as it is cut by veinlets of calcite and is bounded by stylolite seams in many places, it cannot be a recent replacement. A third variety of chert, which Sando called cryptozoon chert, has replaced large cryptozoon masses and accumulates in great bouldery deposits in the soil.

King (1950, p. 36) believed that much of the silicification of the Beekmantown Dolomite is a weathering feature, for exposures of secondary chert that preserve structural features of the rock, including even slickensides, are not traceable into the dolomite below.

In the Shenandoah Valley, exposures of chert in altered limestone are common, but chert rarely constitutes an appreciable percentage of the fresh rock. At a deep exposure of Beekmantown Dolomite east of the Middle River on U.S. Route 250 west of Staunton, a lens of chert about 5 feet thick pinches out abruptly down dip. In Luray Caverns, where a section of about 100 feet in the lower part of the Beekmantown is exposed, large cauliflowerlike masses of sponge-like porous chert are abundant. This chert is dark gray, whereas chert on the surface is generally brown or reddish brown. The dolomite at the cavern level is overlain by 90–120 feet of rock. In the writer's opinion some of the chert in the Beekmantown has moved into the dolomite and replaced parts of it at various times since its formation, but no interpretation of its origin is simple. Some of the chert is relatively recent, as King suggested. Much of the chert that forms thick masses in the overlying residum and on the ground surface may be a concentrate formed during the weathering cycle and derived from great thicknesses of rock. The concentration may take place through chemical as well as mechanical processes.

LIMESTONES OF MIDDLE ORDOVICIAN AGE

The rocks of Middle Ordovician age are markedly different from the rocks beneath them and constitute a distinctive unit. They are readily separated from the Beekmantown Dolomite and the Cambrian formations because they are darker gray and contain no dolomite and some, though very little, chert. Because the Middle Ordovician rocks contain high-grade commercial limestone deposits and also because of their distinctive nature, their stratigraphy has been much studied, and they have been subdivided into formations (Cooper and Cooper, 1946; King, 1950, p. 37; Neuman, 1951).

The lowermost part of the Middle Ordovician sequence is commonly a pure light-gray dense limestone known as the Mosheim (Butts, 1940) but renamed the New Market Limestone by Cooper and Cooper (1946). This rock is commonly a high-grade high-calcium limestone, and many quarries occur along its belt of outcrop. The Mosheim is overlain by a granular fossiliferous limestone that contains black chert nodules. This cherty formation is a ridge maker in places in the southern part of the Shenandoah Valley. It was named the Lenoir Limestone by Butts (1940) and the Lincolnshire Limestone by Cooper and Cooper (1946).

The upper beds of the Middle Ordovician sequence are impure limestones that contain calcareous shale and siltstone in the southern part of the area. They contain no chert. In the northern part of the area, the upper beds are mostly thin-bedded limestone and some siltstone beds. These upper beds, called Athens Limestone in the southern part and Chambersburg Limestone in the northern part by Butts (1940), make up the greater part of the total thickness of the Middle Ordovician in the Valley. The thickness of the whole sequence ranges from 600 feet to more than 1,000 feet.

MARTINSBURG SHALE

The Martinsburg Shale of Middle and Late Ordovician age conformably overlies the limestone sequence in the Shenandoah Valley. The formation consist of thick argillaceous and calcareous shale, siltstone, fine sandstone, and, in places, thin limestone beds, especially near the top. As far as is known, the formation is lacking in chert. It appears to lack competent beds, and isoclinal folds are common. Because of the folding, no reliable estimates of thickness have been made. Butts (1940) gave 4,000 feet as the probable maximum thickness.

SANDSTONES OF ORDOVICIAN AND SILURIAN AGE

Included in the sandstones of Ordovician and Silurian age is a group of ridge-making sandstones and quartzites containing some shale. The principal ridge maker is the Tuscarora Quartzite, one of the most resistant rocks in the area. It breaks up on weathering into large boulders that spread over less resistant rocks in the slopes below the outcrop. The total thickness of these rocks varies greatly but probably ranges from 300 to 1,000 feet (Butts, 1940).

Oswego Sandstone.—The Oswego Sandstone of Late Ordovician age crops out in a discontinuous belt along the east flank of Little North Mountain. It does not occur on Massanutten Mountain. It is composed mostly of bluish-gray sandstone and thin beds of red shale. As it occurs generally close to the more massive and thicker Tuscarora Quartzite, it is commonly covered by boulders or other float and rarely crops out.

Juniata Formation.—The Juniata Formation of Late Ordovician age is composed of red shale or mudstone and contains beds of brown to red sandstone. The color is the most distinctive feature. The thickness is 200 feet or less. The formation is exposed in a few places on North Mountain along the west side of the Shenandoah Valley, but it does not crop out on Massanutten Mountain. Because of its proximity to the Tuscarora, it is generally covered by float or talus.

Tuscarora Quartzite.—The Tuscarora Quartzite of Early Silurian age is extremely resistant and is second only to the Antietam Quartzite as a source of boulders and cobbles in the streams of the Valley. It is almost entirely composed of a pure quartz sandstone that is very light gray. In places, it contains layers of quartz pebbles. It breaks up on weathering into large blocks and boulders, so the outcrop is generally covered by boulder fields. The thickness of the formation is variable but probably ranges from 50 to 200 feet. On Massanutten Mountain, the thickness of sandstone at this horizon is about 500 feet, but a large part of it is somewhat more friable than most of the Tuscarora and may belong to the overlying Clinton Formation.

Clinton Formation.—The Clinton Formation of Middle Silurian age directly overlies the Tuscarora. It consists almost entirely of shale and sandstone, but the relative proportion of these components varies along the strike. The shaly facies are greenish-gray shale accompanied by some red shale and thin layers of quartz sandstone. In places the formation includes iron-rich sandstone beds that consist of quartz grains in an iron-oxide matrix. The thickness of the Clinton Formation in the Shenandoah Valley is 300–500 feet.

Cayuga Group.—The designation Cayuga Group of Late Silurian age is used by Butts (1940, p. 251) to include several formations that vary considerably in thickness along the strike but that have a total thickness of 430 feet or less. In this area the group consists of interbedded red and greenish-gray sandstone and shale. On Massanutten Mountain the Cayuga contains at least 140 feet of sandstone containing shale partings and 60 feet of argillaceous limestone and shale. On the west and northwest sides of the Shenandoah Valley, the base of the group is marked by a thinly laminated gray limestone. The Cayuga Group is relatively nonresistant to erosion and therefore generally occurs on the lower ridge slopes. It is covered in most places by float derived from the Tuscarora Quartzite and Clinton Formation.

HELDERBERG LIMESTONE AND ORISKANY SANDSTONE

The rocks belonging to the Helderberg Limestone and Oriskany Sandstone of Early Devonian age are less than 200 feet in total thickness and are nowhere ridge makers. Both formations are of considerable economic importance and therefore merit separate discussion. The Helderberg Limestone, or Helderberg Group, is represented in the Shenandoah Valley by thin but distinctive calcareous rocks, including gray to pink limestone at the base, fine-grained cherty limestone in the central part, and calcareous sandstone at the top (Young and Harnsberger, 1955). In many areas the limestones are deeply weathered and represented at the surface only by a cherty residuum. The Helderberg Group, as discussed on page 76, is presumably the source rock for many manganese deposits.

The Oriskany Sandstone is of economic importance in the Appalachians as a source of oil and gas. In the Shenandoah Valley it is a thin coarse-grained quartz sandstone or a fine conglomerate having a calcareous cement. It crops out along the western edge of the Shenandoah Valley and on Massanutten Mountain.

SHALES OF DEVONIAN AGE

The Oriskany Sandstone in the Shenandoah Valley is succeeded by a thick series of fissile shales, thin siltstones, and some sandstones that commonly underlie lowlands and valley floors. The total thickness of these shaly rocks is about 2,000 feet.

Romney Shale.—As mapped by Butts (1940), the Romney Shale of Middle Devonian age is a group that includes several formations, all of which are characterised by a dark-gray to black color and a shaly character. The group includes the Onondaga Formation of Middle Devonian age at the base. The Formation is composed of an olive to brown shade about 100 feet thick.

It is overlain by thin-bedded dark-gray to black fissile shale that is commonly light gray and has a silvery sheen on weathered surfaces. The Romney Shale varies considerably along the strike, however, and at Massanutten Mountain the rock contains siltstones and fine sandstones. The thickness of the rocks mapped as Romney ranges from about 500 to 1,000 feet.

Brallier Shale.—The Brallier Shale of Late Devonian age consists of sandy and micaceous greenish-gray shale and siltstone containing thin interbedded fine-grained sandstone. The rock weathers to light brown and has well-formed spheroidal structures. The thickness in this area is 1,000–1,500 feet. The Brallier becomes more sandy in the upper part and grades into the overlying Chemung Formation.

SANDSTONES OF DEVONIAN AGE

The Brallier Shale is succeeded in the Shenandoah Valley by about 4,000 feet of thin-bedded sandstones interbedded with siltstone and shale. The sandstones tend to be flaggy and brittle, but because of their great total thickness within the formations of Devonian age and the amount of cobbles they shed, they are important ridge formers on the western side of the Shenandoah Valley and elsewhere in the Appalachians.

Chemung Formation.—The Chemung Formation of Late Devonian age is generally differentiated from the underlying Brallier Shale by fossils because no marked break occurs at the contact. The change from a sequence of beds that is predominantly shale to a sequence that is at least half sandstone is gradual. The sandstone beds in the Chemung are not only proportionately more abundant but are thicker, more massive, and more resistant. The sandstone is mainly gray or greenish gray but at some horizons is red. It is generally arkosic and in places conglomeratic. Shale probably constitutes at least half of the formation and is greener, softer, and less fissile than the shale of the Brallier. The rocks are cut by closely spaced joints, and because of its somewhat thin-bedded character, the formation breaks on weathering into slightly flat rectilinear cobbles. The Chemung is about 2,000 feet thick.

Hampshire Formation.—The Chemung is succeeded by the Hampshire Formation of Late Devonian age, a body of rock similar in character to the Catskill Formation of Pennsylvania. However, it is probably younger and not traceable into the Catskill (Butts, 1940, p. 333). The Hampshire is mostly red and thereby easily distinguished from the Chemung. It is a thick-bedded arkosic and micaceous sandstone interbedded with mudstone and fissile shale, which also are red in color. The shale makes up nearly half of the formation. In places the sandstone is extremely flaggy.

Like the Chemung, the Hampshire Formation breaks down into resistant rectilinear cobbles of fairly uniform size. The thickness of the Hampshire is about 2,000 feet.

POCONO FORMATION

The Pocono Formation of Mississippian age is a ridge-making unit that crops out in a relatively small area on the southwest side of the Shenandoah Valley. Like the Antietam and the Tuscarora it contains massive sandstones that produce extensive boulder fields and talus. The Pocono consists of thick, massive sandstones interbedded with shale. Sandstone is especially abundant in the lower part, where some beds are quartzitic sandstone at least 20 feet thick. The shale is generally greenish gray and makes up about half of the total thickness. Some coal beds containing plant fossils are exposed in the upper part of the formation. The Pocono is easily distinguished from the underlying formation because it weathers to yellow colors that contrast markedly with the red and brown colors of the Hampshire. The Pocono does not seem to provide nearly as much cobbly rock waste as do the Hampshire and Chemung, but the sandstones within it are more massive and break up into very large boulders that collect as residual talus on the slopes below the outcrops; the cobbles transported in the streams are apparently very resistant.

The top of the Pocono is not exposed in the Shenandoah Valley; its thickness is not known but is estimated to be at least 1,000 feet in places.

DIKES

Intrusive igneous rocks occur in widely scattered small areas. They are mostly diabasic dikes that crosscut rocks ranging in age from Cambrian to Mississippian, but several small plugs and a sill are also present. Since these rocks are no older than Mississippian and are similar in composition to intrusive rocks of known Triassic age in the Piedmont Province, they are generally presumed to be Triassic.

STRUCTURE

The rocks of the Shenandoah Valley are intensely deformed, and large overthrusts, isoclinal folds, and similar structural features occur. The valley is bounded on the southeast side by the northwestern limb of the Blue Ridge anticlinorium. The outcrop of the younger beds of this anticlinorium, including the Catoctin Greenstone, form the crest of the Blue Ridge. In many places the northwestern limb is overturned against the carbonate rocks of the valley and in places is thrust over them (King, 1950, p. 47). The central part of the valley coincides with the Massanutten syncline, a long structural feature that extends the entire length of the Shenandoah Valley and in which Silurian and Devonian rocks are exposed at the present erosion surface. In this syncline the Martinsburg Shale is exposed for many miles and is deformed into isoclinal or shear folds, with the result that the major cleavage planes and bedding planes are inclined at nearly the same angles.

West of the Massanutten syncline are two great thrust faults, known as the Staunton and North Mountain faults. The North Mountain fault in places thrusts Cambrian rocks over Devonian rocks, and in much of the area it separates carbonate rocks on the southeast from younger clastic rocks on the northwest. Northwest of the North Mountain fault the Silurian and younger rocks are less intensively deformed than the carbonate rocks to the southeast.

Certain structural features such as cleavage, lineation, and jointing are of interest because of the control they exert on drainage and topography. These features have been described in detail only in the Blue Ridge area. In the Elkton area the most conspicous feature is a slaty cleavage that strikes northeast and dips southeast, generally parallel to the axial planes of the folds (King, 1950, p. 50). Similar cleavage has been noted by Reed (1955) in the Blue Ridge near Luray and by Nickelsen (1956) in West Virginia. Nickelsen refers to the cleavage as a flow cleavage. Similar cleavage to that described in the Blue Ridge is extensive in the Massanutten syncline in the Martinsburg Shale and in other shaly beds in other parts of the valley.

A lineation in the plane of slaty cleavage that strikes generally down the dip of the cleavage has been noted by King, Reed, and Nickelsen. It is manifest principally as an elongation of mineral grains. The siltstone in the Martinsburg Shale in the Massanutten syncline is broken into match-size prisms, which are alined down the dip of the cleavage.

Joints are especially important geomorphically, for they determine the courses of many rivers (Hack and Young, 1959). The most extensive joint system in the area trends in a vertical plane that strikes northwest at right angles to the fold axes. It can be seen in almost every outcrop and is particularly well formed in the more brittle rocks such as sandstone and quartzite. In places the rocks are broken by fracture zones that extend in straight lines for many miles. In these zones the rocks near the surface are so extensively broken that they form reservoirs for soil moisture and have influenced the pattern of vegetation. Hack and Goodlett (1960, p. 5) traced one such fracture zone for nearly 6 miles because it was marked by trees that differ from and characteristically require more moisture than the trees on adjacent ground. Most of the Triassic dikes were intruded along this joint system. Other joints strike northeast and dip northwest parallel to the fold axes and generally normal to the axial planes of the folds. This joint system is described by Nickelsen (1956, p. 257), and has been observed by the present writer in the Massanutten syncline.

TOPOGRAPHIC FORMS

Study of the topographic forms in the Shenandoah Valley reveals a close dependence of the forms, relief, drainage pattern, and other aspects of the topography on the kind of bedrock. Generally, areas of similar bedrock have similar forms. Some dissimilarities or differences in form that do exist within the same rock areas probably indicate that the equilibrium of form is not complete.

The Shenandoah Valley may be subdivided into several subareas or physiographic provinces that are determined by the kinds of rock that underlie them (fig. 3). The North Mountain area is underlain by rocks of Silurian age and younger, which are mostly clastic and contain thick sandstone sequences. Massanutten Mountain is similar and is a synclinal belt that contains the lower part of the same clastic sequence, ranging from the Silurian Tuscarora Quartzite to the Devonian Romney Shale. The Blue Ridge Mountains on the southeast are supported by resistant clastic and igneous rocks of Early Cambrian age and older. The lowlands in the central part of the Valley are divided into two areas: (1) a lowland on the carbonate rocks which surrounds (2) a long narrow central lowland belt on the Martinsburg Shale.

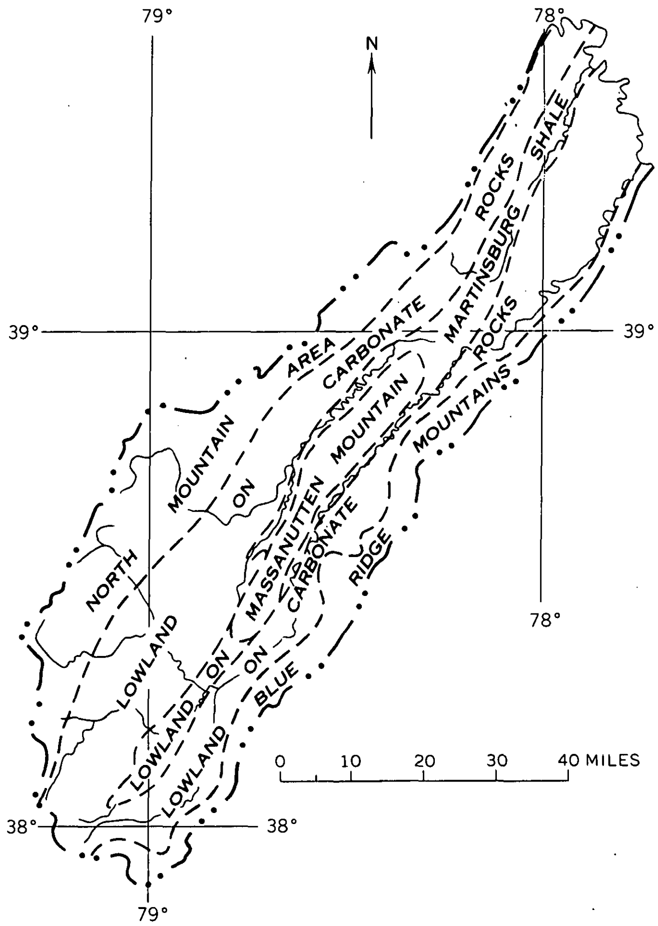

FIGURE 3.—Physiographic subdivisions of the Shenandoah Valley.

In the section that follows, the elements of form that make up the erosionally graded terrain of these areas are described in terms used in more detailed studies (Hack, 1957; Hack and Goodlett, 1960). The description is based primarily on the consideration of the longitudinal stream profiles of the valleys and on the forms of interstream areas.

THE ELEMENTS OF TOPOGRAPHIC FORM

The largest part of the Shenandoah Valley is an erosionally graded landscape. Some areas, amounting to about 15 percent of the whole valley, are underlain by alluvial deposits in flood plains, terraces, and thin pedimentlike aprons at the mountain foot. Exclusive of these alluvial areas, which are relatively flat, the surface is almost entirely in slope, and the land slopes toward a stream, either steeply or gently. The only exceptions are the rounded crests of ridge tops, and if the form were perfect, even the crests would be entirely sloping except for a line along the very top. The overall shape or form of the landscape is determined by the network of stream channels, each channel being concave to the

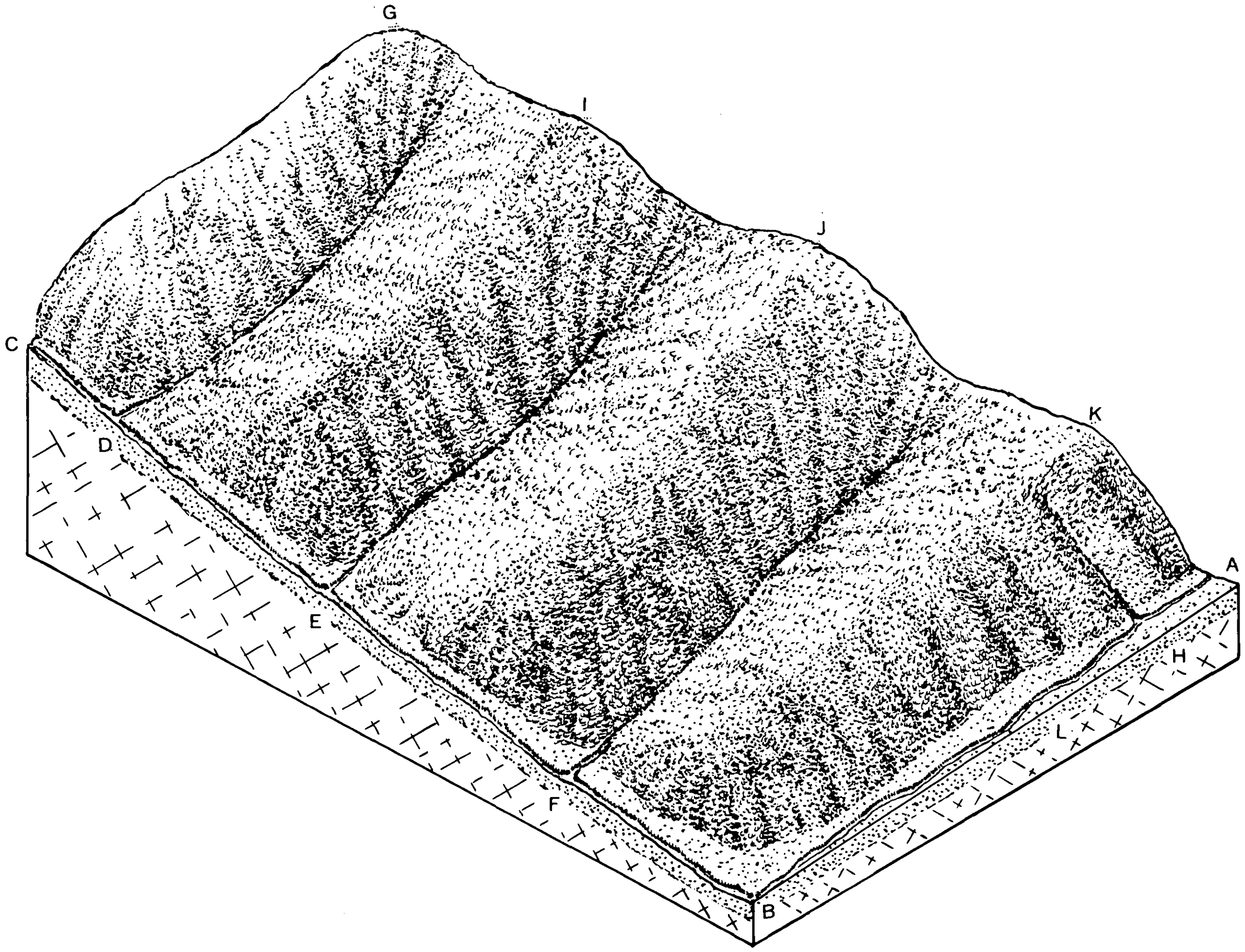

FIGURE 4.—Idealized relation of the interfluves and ridge crests to the framework of the streams. See text for discussion.

sky. The local relief, however, is determined by the interfluves or ridges, which rise to a more or less even height above the streams. Thus the general plan of the landscape is a spoon-shaped multiribbed surface, concave upward. Although the network of streams determines the overall shape of the area because it controls the absolute altitude of the intervening ridges, the details of form in the interstream areas are molded not only by flowing water but by other processes as well. The forms in the interfluves are therefore in part independent of the form of the stream channels.

The geometric relation between the streams and the interfluves is illustrated in figure 4. The streams are like ribs that form a skeleton supporting the entire framework. The interfluves in the zone of overland flow, although rather irregular in shape, tend to rise to a constant height above the streams. Three orders of streams are shown in the diagram. Stream A–B is a segment of a 3d order stream, and since channel slope is a function of discharge or size of stream, the profile of A–B is gentle. Stream C–B is a segment of a 2d order stream whose profile is somewhat steeper. The streams at D, E, and F are 1st order streams, and their profiles are even steeper. The ridges or interfluves rise an even distance above the streams. As a consequence the network of ridge crests parallels the drainage network. Thus the ridge A–G parallels the stream C–B, and the ridges that culminate at points I, J, and K are parallel to the 1st order streams that terminate at D, E, and F. Since the stream profiles in an area like the Shenandoah Valley are concave upward, the ridge crests are also concave upward in a longitudinal direction. The lowest point in the area is at B, on the largest stream. The highest point is the point farthest away at G.

The interfluves in the domain of overland flow are crudely prismatic in shape. Strictly, however, the cross profiles are not triangular as they would be if the interfluves were truly prismatic; rather, they are parabolic. The side slope steepens downward, and its curvature becomes gradually less until it approaches a straight line. The degree of curvature of the interfluves and their height above the streams tend to be constant in any homogeneous area because a channel must be maintained by a certain minimum area of slope (Schumm, 1956, p. 607). Thus the width of the interfluves tends to be a constant, and since the shape of the slope is also a constant, the height must be constant. Where the area of slope adjacent to a channel becomes significantly larger than the average, a small stream channel tends to form (as at H, fig. 4), or simply, a concavity or hollow forms (as at L, fig. 4).

Within a wide area of homogeneous topography of this type, the relief of any two areas of the same size will be approximately the same. The total relief, however, is also a function of the total size of the drainage basin. In general, since the ridges parallel the channels, the general slope of the upland parallels the skeletal framework of the streams. This kind of topography has been called a ridge-and-ravine topography (Hack, 1960a, p. 89) and is widely formed. It makes up most of the Shenandoah Valley except for the 15 percent that is alluvial-terrace land. Though the details of form vary widely, the arrangement of the form elements is generally similar. The differences are largely a function of the kind of bedrock.

LONGITUDINAL STREAM PROFILES

Studies of longitudinal stream profiles made in the Shenandoah Valley have been described elsewhere (Hack, 1957). They are briefly reviewed here as they apply to the description of the landscape. In general, the rate of increase of drainage area in a downstream direction is remarkably regular. In the Shenandoah Valley the drainage area increases with respect to length in accord with the following formula:

$$L=1.4A^{0.6},$$

where L is the distance from a point on the stream to the head of the principal stream above the point, and A is the drainage area at the same point. Differences of rock and geologic structure appear to affect this relation to a small degree, and in some areas the coefficient in the above equation is as high as 2. Since the average discharge increases in direct proportion to the drainage area, the increase in discharge in a downstream direction is proportional to the 1.66 power of the stream length. This is, of course, a very high rate of increase, and a stream 10 miles long has an average discharge over 25 times as great as a stream 1 mile long.

Channel slope is partly a function of discharge and decreases as discharge increases. Slope at any given point on a stream is also a function of the kind of bedrock (Brush, 1961) and of the bed and bank materials deposited by the stream (Hack, 1957). In any area of uniform lithology where the streams originate in and pass through the same kind of rock, the materials carried by the stream, the banks, and the enclosing rock will be the same for any given drainage area. Therefore, the channel slope will be the same, and the stream profiles will be similar.

Figure 5 is a graph showing the relation between channel slope and stream length at various localities in the Shenandoah Valley in areas of three different kinds of rock. Note that in areas of Devonian sandstone and in areas of Martinsburg Shale the relation between the two is very well defined. On the average, streams of a given stream length in the sandstone areas tend to have channel slopes about seven times as steep as streams in Martinsburg Shale. For limestone areas, the scatter of points on the graph is very much greater, but the values for channel slopes in general lie between those for slopes in the sandstone areas and those for the slopes in the shale. The rocks in the limestone areas are very much more diverse in character, and in some areas they shed large quantities of chert into the streams; because of its resistance, the chert affects the steepness of the profile. Much steeper longitudinal profiles are produced on the cherty dolomite of the Beekmantown Dolomite than on the relatively soft fissile limestones like the Athens Limestone.

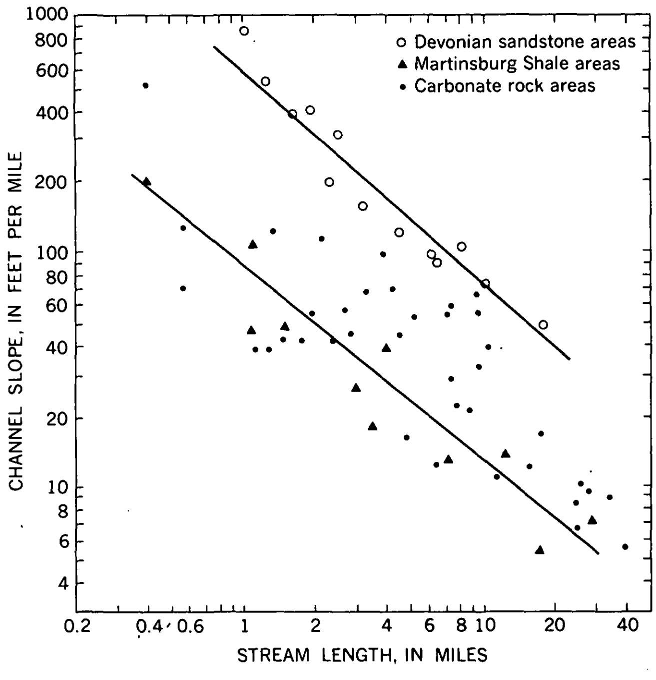

FIGURE 5.—Relation of channel slope to stream length (measured from the source) on different kinds of rock at various localities for streams in the Shenandoah Valley. Logarithmic coordinates.

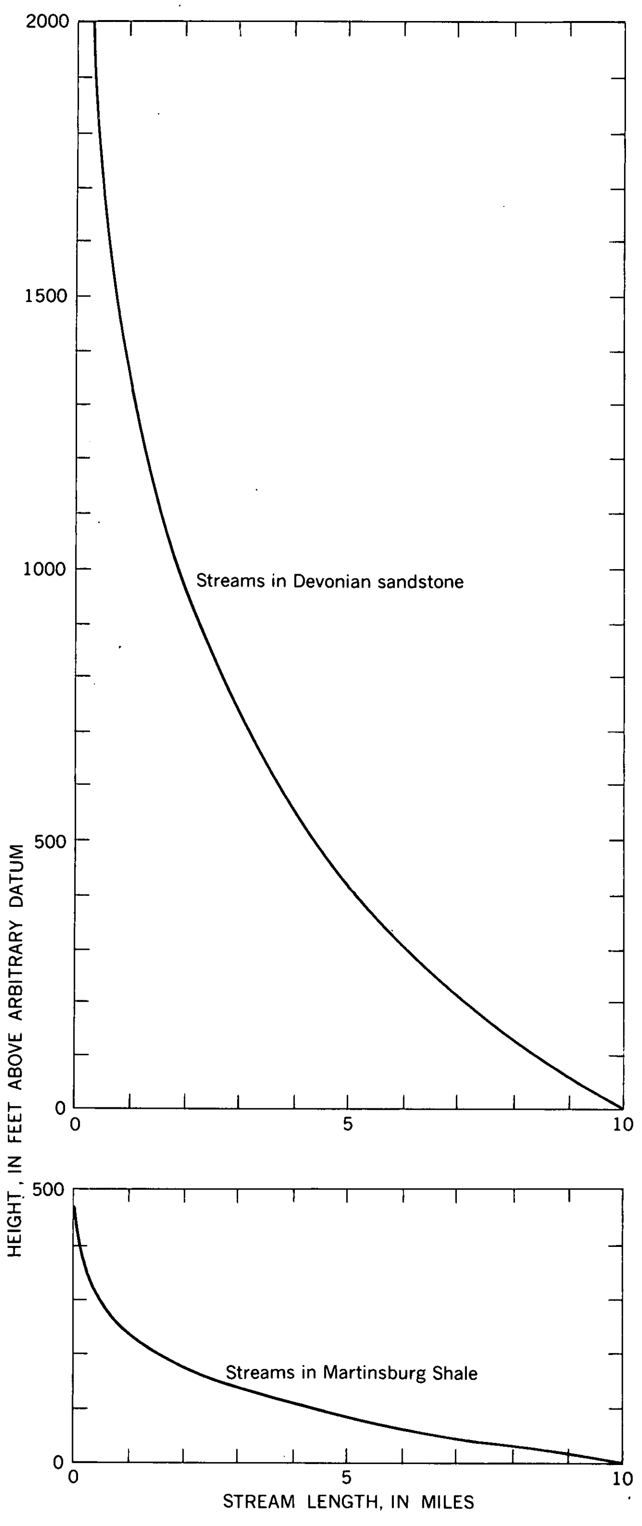

FIGURE 6.—Profiles of typical streams within areas of sandstone and shale in the Shenandoah Valley. The profiles are constructed by integration of the corresponding curves in figure 5.

The relation between channel slope and stream length may be expressed as a differential equation. By integration of such an equation, a curve may be constructed that represents an average profile of all the stream segments within a given homogeneous rock type. Average profiles of streams in the Martinsburg Shale and streams in the Devonian sandstone area of North Mountain are shown in figure 6, which indicates the great difference between the two kinds of profile. An average profile for the limestone region is not as meaningful because of the large differences between individual streams. Such an average curve, however, would be similar to, though slightly steeper than, the curve for the shale. In the Blue Ridge Mountains the stream profiles are similar to those in the sandstone area, though they are somewhat steeper in Catoctin Greenstone areas and somewhat gentler in the rocks of the Chilhowee Group.

FORMS IN INTERSTREAM AREAS

Though the streams form the skeletal framework of the topography, the distinctive character of the topography is mainly evident in the forms of the interstream areas. The elements of form in typical interfluves have previously been defined for a small area of Devonian rocks in the North Mountain area (Hack and Goodlett, 1960 p. 5–10). The valley side or interfluve may be divided into three classes of slope:

The hollow.—The first is the hollow, which consists of any area on the slope in which the contours are concave outward away from the ridge. Hollows generally occur in the valley axis at the stream head as a sort of extension of the channel, and because of the spoon shape, drainage or runoff tends to concentrate downslope. In the lowlands, hollows are also common on the valley sides as tributaries of the stream channels, which break the uniformity of the slope. As the hollow is an area of concentration of moisture, it is wetter than the remainder of the interfluve and contains different vegetation. In the mountains the hollows commonly support a northern hardwood forest, whereas the remainder of the interfluve has stands of oak forest or yellow pine. In the lowlands in areas of open pasture, the hollow is commonly evident because it supports more lush grasses and herbs of different species than does the remainder of the slope.

In mountain hollows the ground consists of fields of subrounded to angular cobbles and boulders that are much coarser than the surficial mantle of other slopes, a fact indicating that mechanical sorting of the mate-

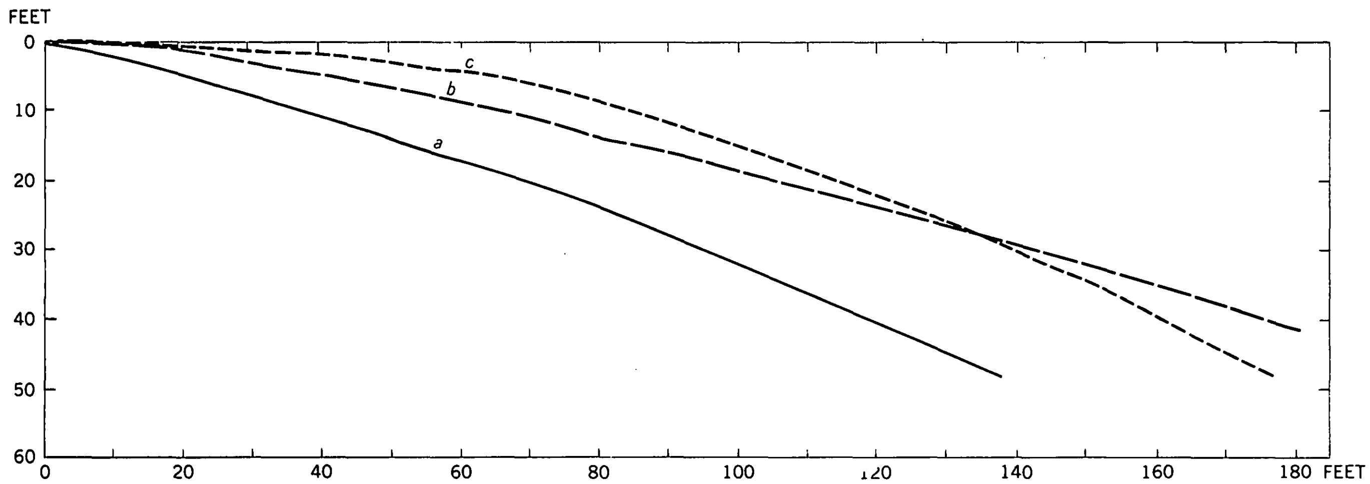

FIGURE 7.—Profiles of slopes at the crests of typical ridges in the Shenandoah Valley. The zero point is at the ridge crest. *a*, West side of Ridge on Chemung Formation West of Staunton, Va.; *b*, east side of ridge on Conococheague Limestone, north of Staunton, Va.; *c*, north side of hill on Brallier Shale, west of Harrisonburg, Va.

rial by large discharges of water is a factor in molding the form. In lowland areas, however, soil samples of slope areas indicate little difference in texture between the hollows and other slopes.

The side slope.—Much of the area of the interfluve has slopes that are straight and neither concave nor convex. This area, the second slope category, is called the side slope. In most small valleys in the mountains, the side slope begins at the hollow in the stream head and extends downvalley, forming the valley side above the stream. It ends at the intersection of the valley side with the side of another valley where the slopes curve outward. Straight side slopes occupy a much smaller part of the interfluve in lowland areas than in the mountains.

The nose.—The third slope category is called the nose, which consists of the area on the interfluve in which the slopes are convex outward toward the valley. In this area, percolating moisture or runoff is dispersed radially downslope and shed from the nose to adjoining areas. The nose is therefore a dry environment, whose character is reflected in the vegetation (Hack and Goodlett, 1960).

A profile drawn on the map of any interfluve in an erosionally graded terrain from a ridge crest down the slope in a direction at right angles to the contour lines tends to follow a simple parabolic curve, and its form may be expressed as a power function of the type

$$H=CL^{f},$$

where H is the vertical distance from the ridge crest to a point on the slope, L is the horizontal distance, and C and f are constants. Many measurements of such profiles have shown that the slope profiles are generally steeper and more peaked at the top in the mountain areas underlain by sandstone than in lowland areas in softer rocks, where the profiles are more curved (Hack and Goodlett, 1960, p. 7–8). In the Blue Ridge Mountains the profiles range from rather peaked in the Chilhowee Group to rounded in granitic areas. Three typical profiles of ridge crests are shown to scale in figure 7.

Other factors than the kind of rock affect the form of the profiles of the interfluves. If in an area of homogeneous rocks one area has a higher relief than another because of some disequilibrium, the form of the interfluves will be different. Where a stream is engaged in lateral planation at the foot of a slope, the horizontal distance from the foot to the crest is smaller on the side adjacent to the stream than on the other side. Figure 8 shows a cross profile of a hill in the Brallier Shale in the Calfpasture Valley just west of the Shenandoah Valley. The north slope is adjacent to an actively cutting stream. The south slope ends in the headwater area of a small drainage basin. In spite of the asymmetry of the hill, both profiles have a fairly well-graded form.

The differences in the interfluves on rocks of different kinds are evident in several ways. The most striking is in the average height of the interfluve, though height does not differ as much as does the steepness of the stream profiles. In Martinsburg Shale areas, interfluves between first-order valleys average between 50 and 100 feet high. In carbonate rock areas they are between 100 and 200 feet high; and in sandstone areas, between 200 and 500 feet. Although the Hampshire Formation of Devonian age composes the peaks of some

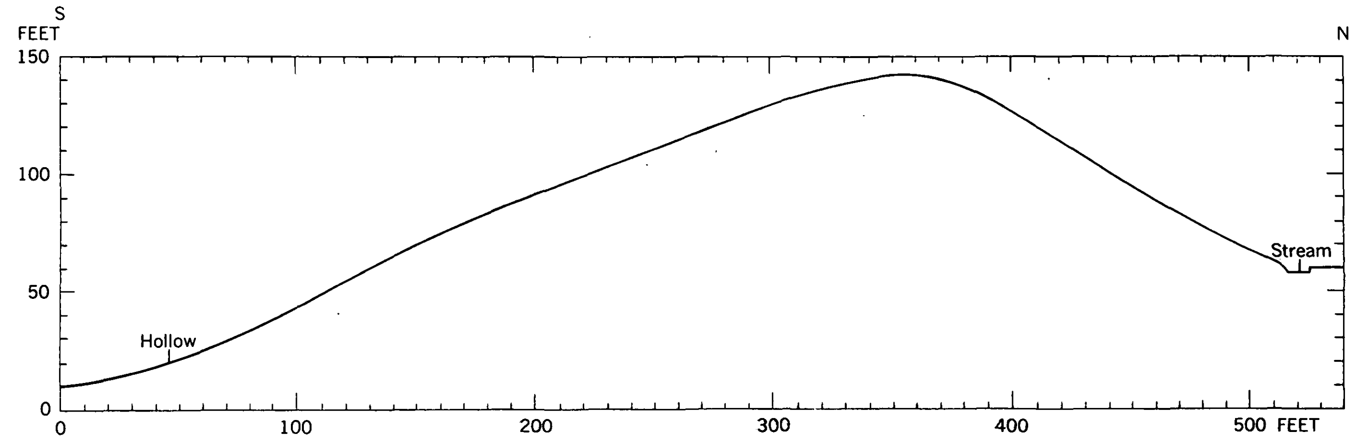

FIGURE 8.—Profile of an asymmetric hill in an area of Brallier Shale west of Staunton, Va.

of the highest mountains in the area, the height of the interfluves in many places is less than 200 feet. Most of the relief is supported by the skeletal framework of the streams, which commonly have an average gradient of 20 percent for a distance of more than 2,000 feet below the stream heads.

In lowland areas, a greater part of the relief is contained in the interfluves, especially in the limestone terrain where the headwater slopes of the streams are very gentle. As explained on page 27, the difference in the way in which the relief is manifest is reflected in the hypsometric curves of drainage basins in the two kinds of areas.

The size and frequency of hollows on the interfluve is also a function of kind of rock as well as the amount of relief. Four typical interstream areas at the heads of small valleys were mapped in the field with plane table and alidade and are shown in figure 9. The difference between the mountain valley and the three lowland valleys is very apparent. The small valley on Crawford Mountain has only one hollow, which occupies only 9 percent of the slope area. The area of side slope is relatively large. In the lowland interfluves the proportion of slope occupied by hollows is much greater and seems to be proportional to the relief. The turkey farm area, having the lowest relief, is thirty percent in hollows. The head of Eidson Creek in a limestone area is only 20 percent in hollows. In all three lowland areas, straight side slopes are relatively insignificant, and the valley walls either curve inward to a hollow or outward to a nose.

In the mountain areas the stream channel has a rather definite beginning and originates sharply below a field of blocks that generally occupies the hollow. In the lowland areas, the channel beginning is less marked, and in places, as in the Eidson Creek valley head, several discontinuous gullies have formed upstream from the continuous channel. In every locality, the channel generally begins a long distance upstream from the place where perennial flow begins. In some places this distance may be more than a mile.

The inflection point of slope is another topographic feature that has a relation to the rock type. A profile made in the longitudinal axis of a valley that begins in the interfluve and extends downslope toward the channel must change in curvature at some point from convex upward near the crest to concave upward downslope either in the channel or above the channel. The point at which the curvature of the longitudinal valley profile changes is called the inflection point. Examination of many headwater areas has shown that the inflection point is not by any means at the same place with relation to the hollow but varies in position considerably. The change from convex to concave may occur at a high point in the hollow, at the lower end of the hollow, or even in the stream channel itself. Presumably the inflection point occurs wherever the drainage area of the slope above it reaches a sufficient value so that the kinetic energy of flood runoff exerts an influence on the form of the slope; thereafter the gradient decreases and is inverse to the increase in drainage area. Examination of several headwater areas on several rock types shows that the inflection point is located much closer to the crest in soft-rock areas than in hard-rock areas and that the area of slope that drains to the inflection point is smaller in the soft-rock areas by a factor of 2–25. Furthermore, the slope of the ground at the inflection point, which is always the steepest point along the axis of the valley, is much steeper in the hard-rock areas than in the soft-rock areas. The data for several valleys are summarized in table 2.

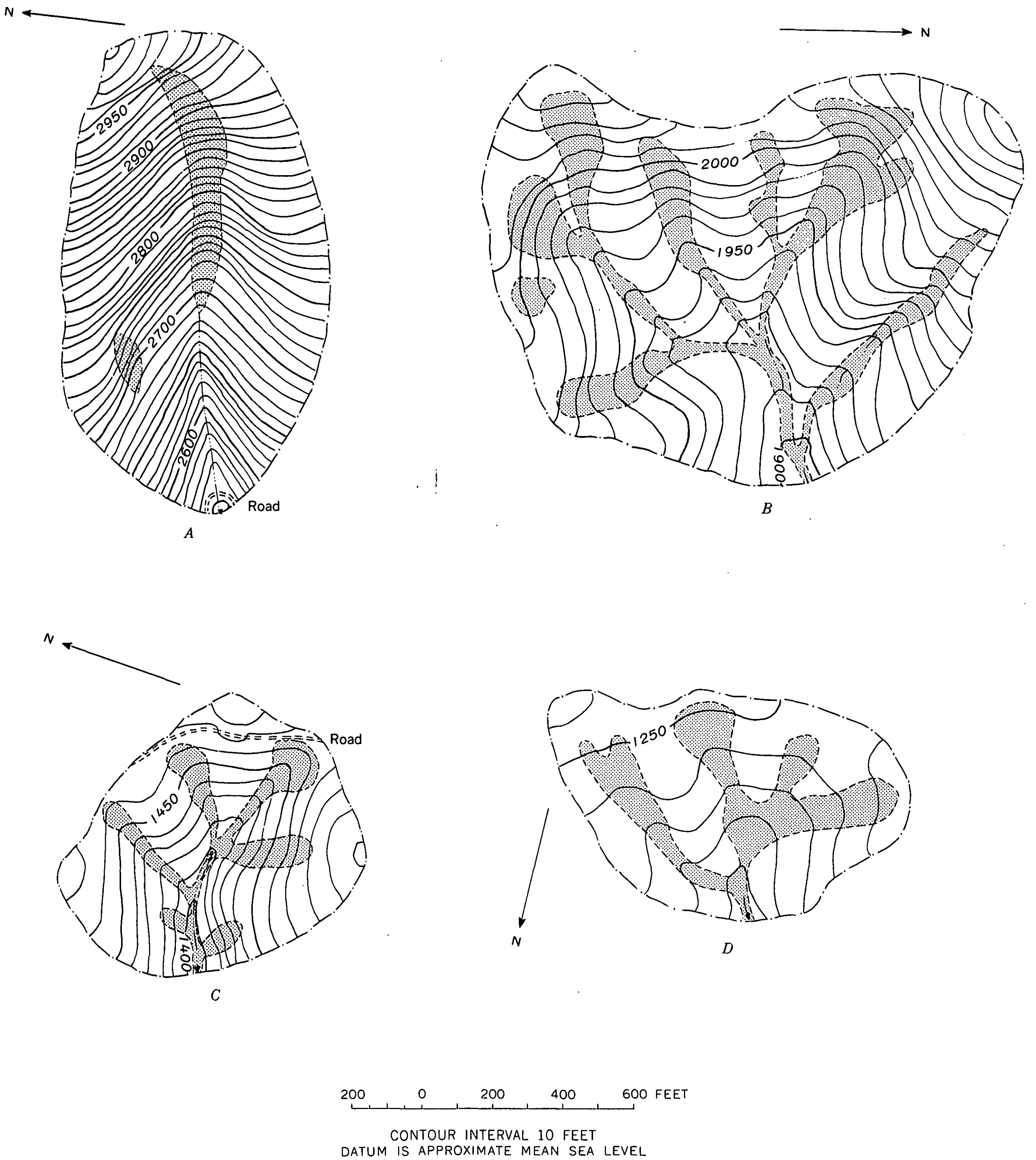

FIGURE 9.—Headwaters of four valleys in areas of different relief, showing the area of the slope that consists of hollows (stippled pattern). Runoff is concentrated in these areas. *A*, Valley head on Crawford Mountain, west of Staunton, in sandstone of the Chemung Formation; *B*, headwaters of Eidson Creek, southwest of Staunton, in carbonate rocks; *C*, valley head on Moffet Farm, south of Staunton, in Martinsburg Shale; *D*, valley head on turkey farm northeast of Staunton in Martinsburg Shale.

TABLE 2.—*Comparison of inflection points of hollows formed in several kinds of rock*

Locality	Kind of rock	Distance from divide (feet)	Fall from divide (feet)	Area of watershed at inflection point (square feet)	Slope at inflection point (percent)
Moffet Farm, 1 mile southeast of Good Shepherd Church, Staunton quadrangle.	Martinsburg Shale.	100	16	28,000	15
Turkey farm, 1 mile west of Damtown, Waynesboro quadrangle. Traverse A.	Martinsburg Shale.	190	17	23,000	10
Turkey farm. Traverse B.	Martinsburg Shale.	110	10	2,000	10
Bonnie Hills Farm, at the head of Eidson Creek, Staunton quadrangle. Traverse A.	Conococheague Limestone.	130	32	5,000	28
Bonnie Hills Farm. Traverse B.	Conococheague Limestone.	180	28	24,000	17
Head of White Oak Run, McDowell quadrangle.	Brallier Shale	140	33	14,000	37
Valley north of Dry Branch Gap, Craigsville quadrangle.	Chemung Formation (sandstone).	300	120	50,000	55
Valley on south slope of Crawford Mountain, Craigsville quadrangle.	Chemung Formation.	320	125	51,000	53

CONCAVE AND CONVEX VALLEY SLOPES

The slopes of valley sides have commonly been thought to be crudely S-shaped in cross profile—that is, convex upward at the top and concave upward at the bottom. In the Shenandoah Valley the apparent concavity of the lower slopes is clearly in part an illusion created by the fact that in looking at the wall of a valley of the second or third order we often are looking obliquely across the interfluves of first-order streams, whose crests are concave upward (fig. 10). True concave valley slopes exist, however—for example, in hollows. In mountain areas like those on the Devonian sandstones of the North Mountain area, concave slopes are rare except in the hollows or at the very foot of side slopes where colluvial material has collected on a flood plain or terrace. On most valley sides in these areas, the side slope is fairly straight all the way to the stream or flood plain.

In the lowland areas, on the other hand, concave slopes are common, probably because hollows are larger and more numerous and the interfluves consist of an almost continuous succession of alternating hollows and noses. In the limestone area the valley floors of small streams are wide relative to the channel. In these broad meadowlike valleys in which the channel migrates from one side to the other, the valley slopes opposite the channel have had a fixed base long enough so that the slope above has been altered by creep and slope wash. In such wide valleys the steepness of the valley sides varies considerably, and the lower slopes are in places marked by concavities, narrow benches, and gently sloping deposits of colluvium.

RELIEF AND HYPSOMETRY

The foregoing analysis of the topography indicates that in a region of homogeneous rocks the local relief should be approximately the same throughout—that is, the vertical distance from a hilltop to a valley or stream adjacent to it should vary little from one place to another. On the other hand, the local relief is very different in areas of different rock types because drainage

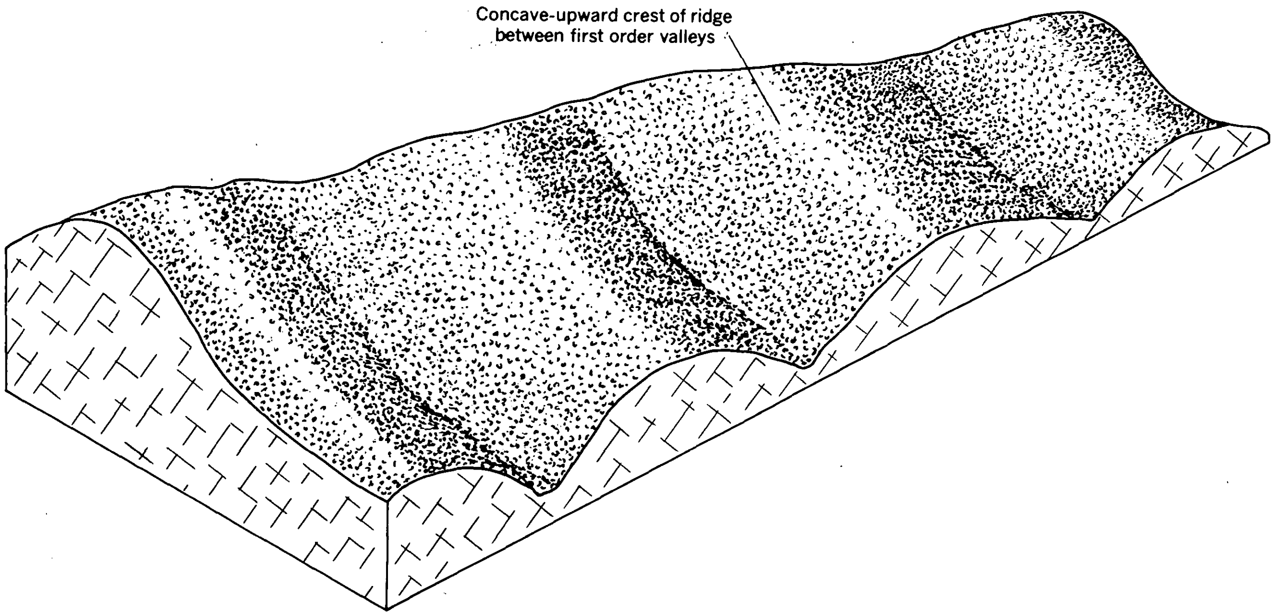

FIGURE 10.—A main ridge and several tributary ridges; the concave longitudinal profiles of the ridges are adjusted to the first-order streams that divide them.

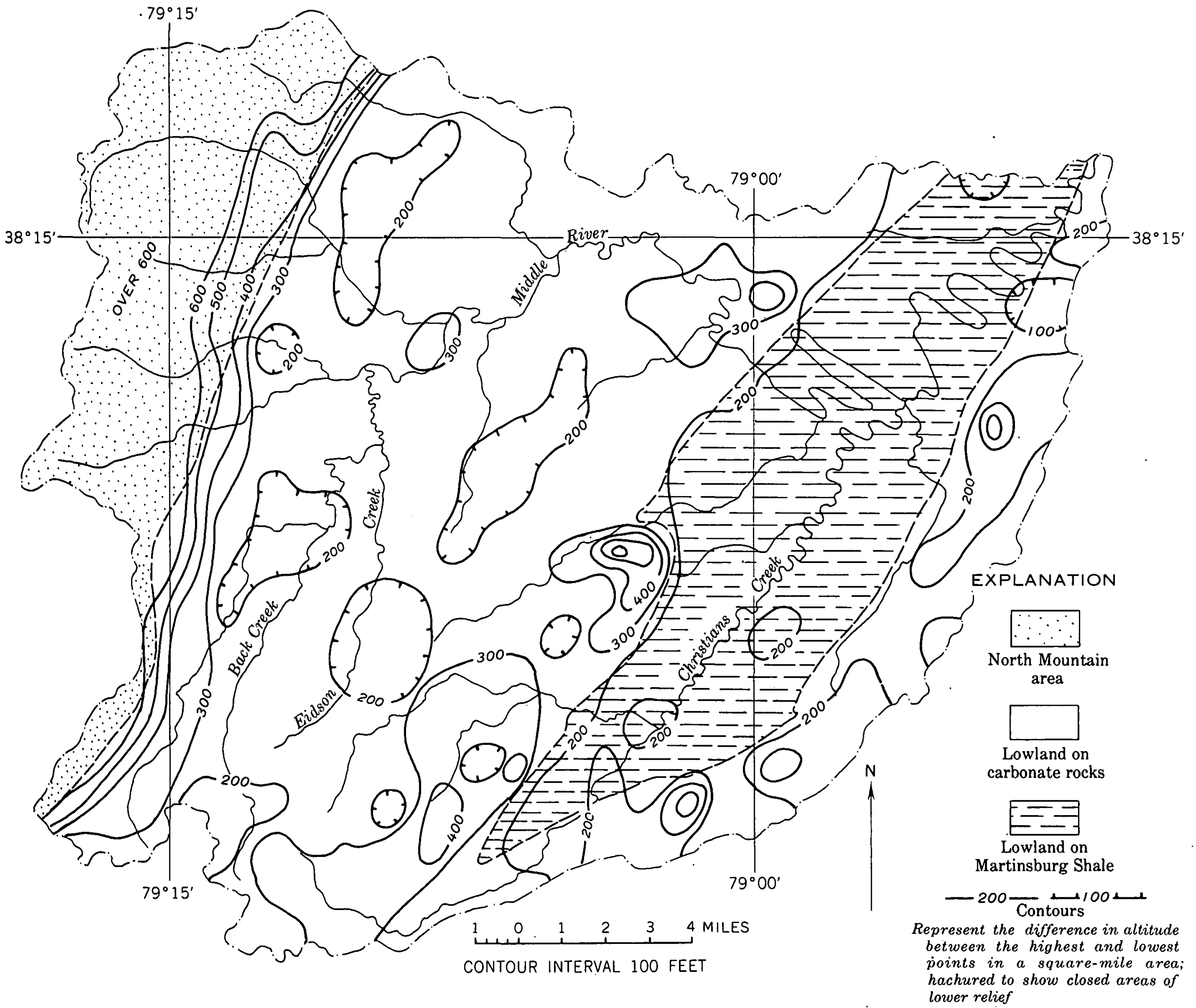

FIGURE 11.—Middle River basin; contour lines show the relief per square mile.

densities, stream profiles, and the shape of the interfluves may all differ. This generalization is confirmed by figure 11, a map of the Middle River basin in the southern Shenandoah Valley in which the local relief is superimposed on a map showing the major drainage system and its principal subdivisions. The contours showing the amount of local relief were obtained by dividing topographic maps of the area into squares a mile on a side. The difference between the highest and lowest point on each square was then plotted and the resulting values contoured. Correlation between relief and general rock type is quite close. On Martinsburg Shale the relief per square mile averages less than 200 feet. On carbonate rocks the relief averages between 200 and 300 feet, though there are high areas along a belt of Beekmantown Dolomite in the center of the basin where the rocks contain considerable massive chert. In the North Mountain area the relief becomes much higher and exceeds 600 feet per square mile.

This phenomenon, the uniformity of local relief, demonstrates the close dependence of the landforms on the skeletal framework of the streams. The Middle River, for example, heads in the limestone area at an altitude a little higher than 2,000 feet. It leaves the limestone area north of Staunton at an altitude of 1,200 feet, having dropped 800 feet. The country close by the river, however, rises to just about the same height throughout the river's course, except where the stream crosses exceptionally resistant beds.

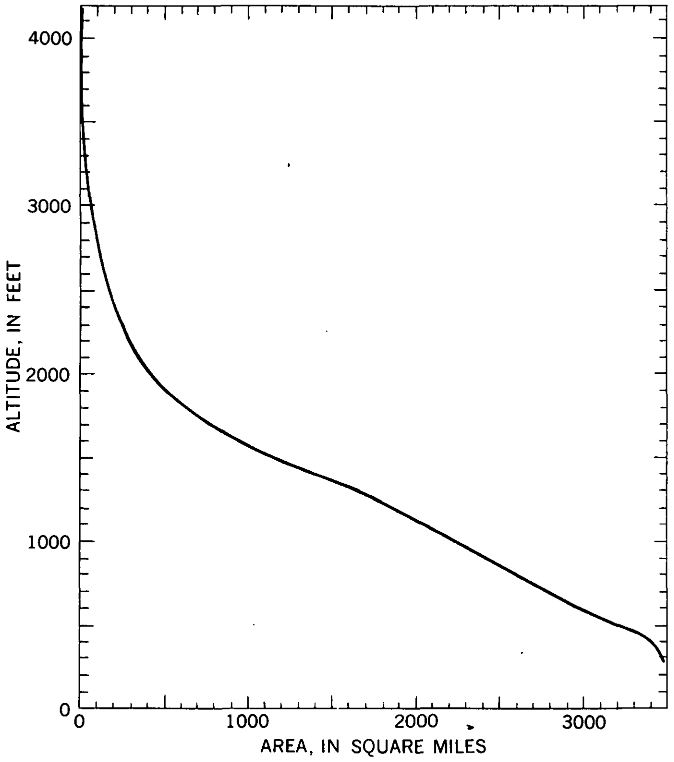

FIGURE 12.—Hypsometric curve for the entire area of the Shenandoah Valley and Opequon Creek valley.

The differences in relief in different parts of the valley can be compared by constructing hypsometric curves. This kind of curve is constructed by measuring with a planimeter the area on a map of all the terrain above successively lower contour lines. The values thus obtained are plotted on a graph. Altitude is plotted on the ordinate; and the area above that altitude, on the abscissa. The relief may be represented by several single numerical figures. The median relief is the difference between the altitude of the lowest point in the basin and that altitude above which half the basin is higher. This value can be read directly from the graph. The average relief can be calculated and is the average height above the lowest point of the basin of the various area-altitude classes measured with the planimeter.

A hypsometric curve showing the altitudes in the entire Shenandoah Valley and Opequon Creek Valley is shown in figure 12. The average altitude is 1,340 feet, and as the lowest point in the area at the junction of the Shenandoah and Potomac Rivers is about 275 feet, the average relief is 1,065 feet.

If smaller and more homogeneous areas are considered, the resulting values for relief are lower and confirm the conclusions drawn from the study of the relief of the Middle River basin. Figure 13 is a map of the entire study area and shows the median relief of small parts of the different subdivisions. The Martinsburg Shale area in the Massanutten syncline has the least relief. The relief in the carbonate rock lowland is somewhat greater, but the two parts of the lowland in the south part of the Shenandoah Valley have identical relief. The carbonate rock area in the northern part of the valley, however, has less relief. The difference in relief is perhaps because the Ordovician carbonate rocks in the northern part contain less chert than in the areas to the southwest. (See p. 13.)

In any hypsometric analysis of a drainage basin, the average relief increases as the drainage area along the principal stream increases because the lowest point descends with the stream and the range of altitudes increases. On a percentage hypsometric curve, however, this increase does not occur. Percentage hypsometric curves (fig. 14) can be constructed from area altitude curves. The ratio of the area between a given altitude and the highest point of the basin to the total basin area or percentage of basin area is plotted on the abscissa, and the ratio of the difference between the given

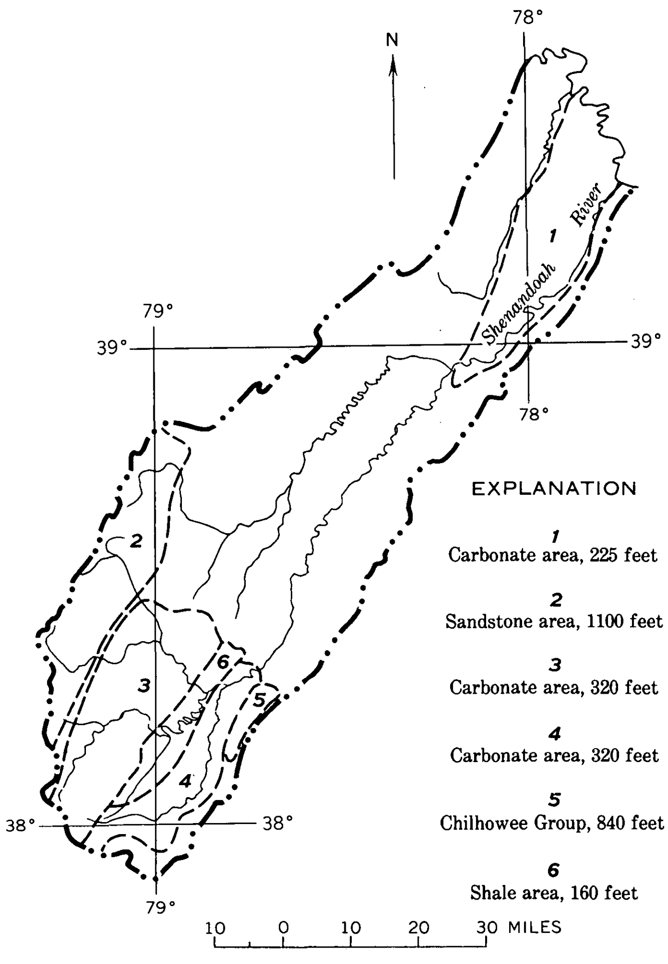

FIGURE 13.—Median relief in selected parts of the study area as obtained from hypsometric curves.

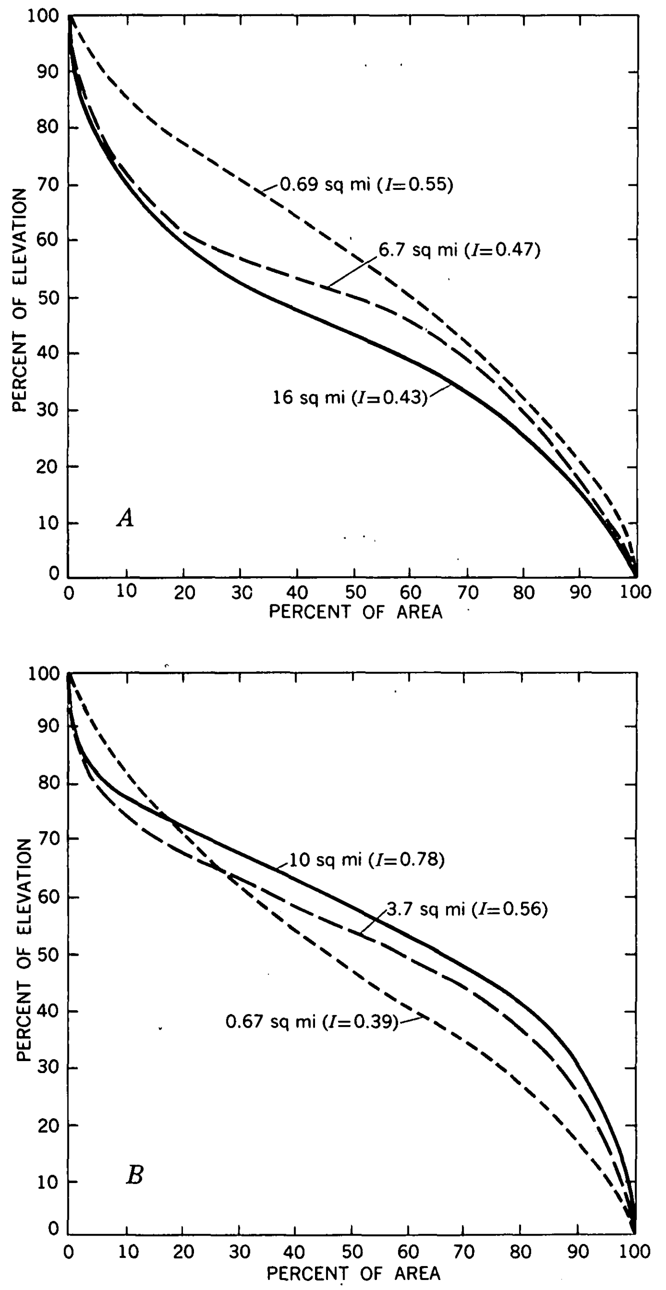

FIGURE 14.—Percentage hypsometric curves for (*A*) the Little River basin (tributary to the North River and underlain by sandstone) and for (*B*) Bell Creek basin (tributary to the Middle River and underlain by carbonate rocks).

altitude and the base to the total height of the basin or percentage of basin height is plotted on the ordinate. (See Strahler, 1952, p. 1119, for method of construction.) The area below the constructed curve on the resulting graph can be measured with a planimeter. The ratio of this area to the total area of the graph is known as the hypsometric integral. In terms of drainage-basin form, a low hypsometric integral means that the largest part of the area is close to the lowest point of the basin, whereas a high hypsometric integral means that the largest part of the area is close to the highest point. A basin that consisted of an upland cut by a deep narrow gorge, for example, would have a high integral. A low plain rimmed by conical hills, on the other hand, would have a low integral.

Construction of percentage hypsometric curves for several small drainage basins disclosed that mountain areas, like the Little River basin in the Hampshire Formation west of Harrisonburg, have rather high hypsometric integrals, but as drainage area increases downstream, the hypsometric integral decreases (fig. 14*A*). In the lowlands this change does not occur. In the Martinsburg Shale lowland, large drainage basins have approximately the same hypsometric integral as small ones. In the carbonate lowland areas, the hypsometric integral increases markedly in larger basins, as is shown by the curves for Bell Creek (fig. 14*B*).

As suggested on page 21, the difference in topography between the sandstone and carbonate rock areas is probably a consequence of differences between the forms of the stream profiles in the two areas and between the forms of the interfluves. In the sandstone areas, exemplified by the Little River basin, the interfluves are rather narrow and fairly low for so rugged a terrain. Hollows are few and are confined to stream heads. Most of the relief, therefore, is accounted for in the skeletal framework of the streams. Because their pattern is concave upward, large drainage basins tend to be bowl shaped or spoon shaped, and the hypsometric integral is low. In small basins, however, the interfluves are proportionally large, and as their form is more prismatic or even convex upward, the higher parts of the terrain compose a larger proportion of the total basin area. In an area like Bell Creek (fig. 14*B*) a much higher portion of the total relief is accounted for by the interfluves themselves. In a small drainage basin consisting only of headwaters, a fairly large part of the area in slope is in either hollows or valley floors. In larger drainage basins downstream, however, the part of the basin that is interfluve is greater. Because the stream profiles are less markedly concave and rise less steeply near the headwaters, their effect on the hypsometric curve is also less marked.

ADJUSTMENT OF MAJOR TOPOGRAPHIC FEATURES

Perhaps the most spectacular phenomenon related to the geomorphology of the Shenandoah Valley is the adjustment of the major elements of the topography to the bedrock.

The correlation between the heights of the principal mountain ridges and the outcrop belts of the resistant rocks is very consistent; the highest altitudes in the area occur in the headwaters of the North and Middle Rivers, where the resistant Silurian, Devonian, and Mississippian sandstones have the broadest outcrop area. The lowest parts of the Valley are, in general, closely related to the outcrops of the least resistant rocks. The Martinsburg Shale thus forms troughlike belts of topography that extend almost the entire length of the valley and are occupied by large streams (pl. 1).

In the Blue Ridge the highest peaks occur east of the town of Luray, where the metabasalt in the Catoctin Greenstone is nearly flat lying and is thicker than elsewhere. The close correlation between the structure and form of Little North Mountain, a ridge formed on the Tuscarora Sandstone, has been studied in detail by Edmundson (1940). Wind gaps are closely correlated with localities at which the quartzite is especially thin or absent.

Although the major wind gaps of the Blue Ridge and Massanutten Mountain have been attributed to erosion by ancient streams (Watson and Cline, 1913), at least some of them are now known to be controlled by local structures. For example, Thornton gap, east of Luray where the Blue Ridge is crossed by the Lee Highway (U.S. Route 211), is the site of a large normal fault disclosed by the detailed mapping of Reed (1955). New Market gap, where Lee Highway crosses Massanutten Mountain, corresponds to a flexure in the Massanutten syncline in which the Tuscarora Sandstone is sharply folded upward above the present erosion surface. Manassas gap, near Front Royal, corresponds to a belt a few miles wide in which the basal Cambrian quartzites are absent because of faulting.

The most remarkable example of adjustment in the region is the course of the Potomac River itself. The Shenandoah Valley is bordered on the northwest by a continuous belt of Tuscarora Quartzite and on the west by a broader belt of Devonian sandstones. The Potomac River crosses this belt at the only place in many miles where the Tuscarora is cut out by a thrust fault. Just north of Martinsburg, West Virginia, the Potomac enters the valley through a gap in the ridge of Silurian sandstone about 5–6 miles wide in which Cambrian limestone is thrust over the relatively nonresistant Devonian Romney Shale. East of this point the Potomac River winds through the shale and carbonate rocks of the Shenandoah Valley, and at Harpers Ferry the river leaves the valley where the ridge-making rocks that border the valley on the east have an unusually narrow outcrop.

In order to leave the valley the Potomac River must cross not only the quartzite of the Chilhowee Group but also the Catoctin Greenstone. As stated on pages 10 and 11, the Catoctin Greenstone, which forms the crest of the Blue Ridge farther south, thins in the Harpers Ferry area to only 50 feet and no longer forms the crest of a ridge. Instead, the Weverton is the most resistant rock and forms Elk Mountain, through which the Potomac passes in a deep gorge. The Catoctin crops out on the southeast flank.

To the south the Weverton is not as resistant as either the Catoctin Greenstone or the Antietam. It forms a ridge near Harpers Ferry only because the other two formations have so changed in character and thickness that they are not high ridge-makers. To the north in Maryland, the outcrop areas of the quartzitic rocks are greatly widened by broad folds (Cloos, 1941).

The possibility that the Potomac River would enter and leave the Shenandoah Valley through weak places in the bounding belts of resistant rock if its course had been determined by superposition is too remote a coincidence to be believed. Furthermore, the adjustment of so great a river to these structural weaknesses could hardly have taken place while the river was eroding vertically only a few thousand feet. Such perfect adjustment must have occurred as the result of the erosion of many thousands of feet of rocks and may have involved major changes in the drainage pattern.

Examples of lack of adjustment in the topography also exist. Some of these can be seen by examination of plate 1, the bedrock geologic map of the Shenandoah Valley. At Columbia Furnace, Stony Creek, a large tributary of the North Fork, Shenandoah River, crosses the Silurian Tuscarora Quartzite. A few miles to the south, however, this resistant bed is cut out by a thrust fault. Perhaps this stream has been superposed on one rock type from another, for it is possible that when the erosional surface was a thousand or more feet higher, the Tuscarora may have been missing at this place because it was cut out by the same fault. Bloomer (1951) suggested a similar mechanism on a more spectacular scale to explain the gorge of the James River in the Blue Ridge to the south of the Shenandoah Valley.

Other examples of apparent lack of adjustment exist. South of Staunton the topography on the Martinsburg Shale to the west of Christians Creek is steeper and has a higher relief than the topography east of the creek. This phenomenon can be demonstrated on plate 1, because west of Christians Creek most of the Martinsburg Shale area is above 1,400 feet, whereas to the east it is mostly below 1,400 feet. The difference is striking on the 15-minute topographic map of the Staunton quadrangle published by the U.S. Geological Survey. The reason for the difference is not known. Perhaps the streams draining the rocks on the northwest limb of the Massanutten syncline have somewhat steeper gradients because they are not yet adjusted to the rest of the drainage network. Perhaps their profiles are in part inherited from a time when the streams west of Staunton drained past Harrisonburg into the North Fork of the Shenandoah. Possibly, however, the apparent disequilibrium is due to the high terrain southwest of Staunton where the Beekmantown Dolomite is covered with residual chert. The chert may exert an important influence on the profiles of the streams flowing across it. If this is the correct explanation, the higher topography west of Christians Creek is actually in equilibrium with the topography east of the creek.

LONGITUDINAL PROFILE OF THE SHENANDOAH VALLEY

The longitudinal profile of the Shenandoah Valley steepens sharply upvalley from Front Royal (fig. 15). In the 36 miles from Harpers Ferry to Front Royal, the average gradient of the lowland as measured along the hilltops bordering the river is about 4 feet per mile. Above Front Royal the gradient of the lowland along the South Fork as far as Shenandoah increases to 12 feet per mile. The gradient of the more open lowland country bordering the North Fork is even steeper, and between Edinburg and Strasburg, it is 17.5 feet per mile. Above Edinburg the gradient declines again to about 11 feet per mile.

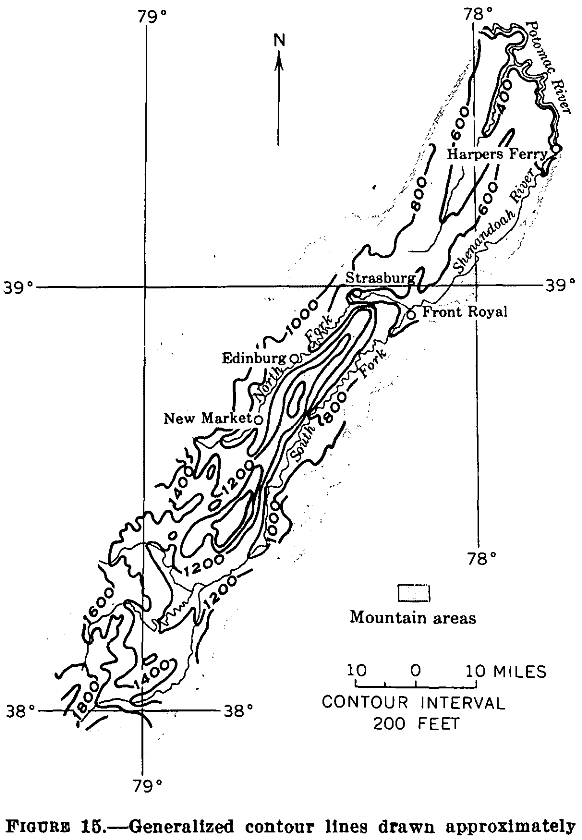

FIGURE 15.—Generalized contour lines drawn approximately at the level of the hilltops in the carbonate rock and shale lowlands of the Shenandoah Valley.

Fenneman (1938, p. 247–248) thought that the gradient of the lower valley could have been produced during peneplanation but that the upper and middle valley were too steep and must have been warped. King (1950, p. 62) noted that the gravel deposits bordering the South Fork were graded to the present steep longitudinal profile of the main river and the tributary streams. He suggested that if a peneplain once existed at the level of the hilltops in the lowland, it must have been warped before the gravel was deposited.

If the topography of the Shenandoah Valley is in erosional equilibrium and the lowland area is adjusted to the streams that drain it, the steep longitudinal profiles of the lowlands along the North and South Forks can be explained without postulating warping. In a study of the North Fork Shenandoah River, Hack and Young (1959) showed that the steep downvalley profile of the lowland is related to the unusual meanders formed along joint planes where the river flows through Martinsburg Shale between Strasburg and Edinburg. The river channel itself has a smooth profile throughout its entire course; the profile follows closely a simple logarithmic curve and is gently concave upward. The average gradient decreases from 9 feet to the mile in the reach between New Market and Edinburg to 5.5 feet per mile in the reach between Edinburg and Strasburg. The river is graded throughout to transport cobbles and boulders of sandstone shed into it from Massanutten Mountain and from the mountainous country in the headwater area. The gradual decrease in the gradient is consistent with the gradual increase in discharge in a manner normal for streams in this region. The sinuosity of the river (defined as the ratio of the distance measured along the channel to the distance measured straight downvalley) is 1.4 above Edinburg but averages 3.2 from Edinburg to Strasburg. Thus the change in sinuosity is much greater than the change in gradient in these two reaches. In the meandering reach the river must travel more than twice as far laterally as it does downvalley, and the downvalley gradient of the lowland that borders the river obviously must become much steeper in order to maintain the river grade. The river descends the valley like a graded highway in a series of switchbacks. The hills bordering the river are graded to the river and rise about the same distance above the river all along the course. Because the gradient of the hilly lowland is determined by the gradient of the river, the ratio of the two gradients must be equal to the sinuosity.

Statistics comparing the river gradients and the downvalley gradients of the lowlands bordering the river are listed in the following table:

River	Average gradient (feet per mile)		Sinuosity	Ratio of lowland gradient to river gradient
	River	Lowland		
North Fork:				
New Market to Edinburg	9	11.4	1.4	1.3
Edinburg to Strasburg	5.5	17.5	3.2	3.2
South Fork: Shenandoah to Front Royal	6.0	11.5	1.8	1.9
Shenandoah River: Front Royal to Harpers Ferry	3.1	3.9	1.4	1.2

The data indicate a close adjustment of the river to the topography along it, and the explanation of the sharp rise in the carbonate-rock lowland is clearly identified with the high sinuosity of both forks of the Shenandoah River. The origin of the meanders has been discussed by Hack and Young (1959) and is related to the pronounced planar structures in the Martinsburg Shale in which the meanders occur.

THE REGOLITH

The mantle of unconsolidated surficial material in the Shenandoah Valley consists partly of residual material—known as saprolite—derived from the rocks directly beneath; partly of colluvium, which has moved short distances by gravity or surface wash; and partly of alluvium, which has been transported and sorted by running water. In some areas the regolith is thin or lacking entirely, but in others it has accumulated to thicknesses greater than 200 feet and roadcuts more than 40 feet high in residual material are commonplace. The close relation of the regolith to the bedrock has long been noted and taken for granted by geologists who have mapped the bedrock in the region, particularly the carbonate rocks. Rapid field identification of many of the rock formations is considerably aided by examination of the regolith above them. To the writer's knowledge, however, no systematic study or map of the regolith covering an appreciable area has been made in this part of the Appalachians.

With the idea that a knowledge of the character and distribution of surficial materials is essential to an understanding of the geomorphic history, the writer undertook a reconnaissance survey of the regolith of the entire valley. Presumably, if the hilltops in the lowland areas are remnants of an erosion surface of low relief, or peneplain, then one should expect to find a fairly widespread alluvial mantle deposited by a river system graded to the level of the hilltops. Residual deposits thought by some geologists to be remnants of a mantle produced by Tertiary weathering should be concentrated in parts of the region where the peneplain is best preserved. If, however, the peneplain never existed and the topography formed by long-continued erosion of a great thickness of rocks so that the landscape is now in dynamic equilibrium, then the materials in the surface mantle should be closely related to the rocks from which they are derived, and their distribution should be unrelated to the remnants of any particular level or near-level surface. To resolve this question it is more important to learn the distribution of different kinds of regolith in a very large area than to study the details within a small area. Plate 2 is a reconnaissance map of the regolith in the entire valley of the Shenandoah River and Opequon Creek. Because of the practical limits of time, the units mapped were confined to those that could be delineated readily without digging or drilling. Every public road was traversed, and the roadcuts were briefly examined. Widely spaced traverses were made between the roads in the mountain areas in order to delineate roughly the largest areas of alluvium. Mapping was done on 15-minute quadrangles, and the data were later transferred to a base at a scale of 1:250,000.

With the mapping method used, only a few units could be differentiated. Alluvial deposits could be easily recognized and mapped. They are subdivided into two map units (pl. 2); unit ab, bottom lands and lower terraces that are little if at all dissected and are often or occasionally flooded, and unit at, higher terraces that are distinctly dissected and rarely if ever flooded. This subdivision serves to distinguish recent alluvial deposits presumably produced by a stream regimen like the present from older deposits that might have been deposited under different conditions or by different streams.

Colluvial and residual deposits could not be separated by the reconnaissance methods used and were lumped together as residual mantle. The mantle on shale was different from the mantle on other rocks and could very easily be distinguished by only casual examination (unit s, pl. 2). It contains many small shale fragments that are scattered on the surface in both woodlands and pastures. Except where covered by alluvium, the residuum on shale is thin, and bedrock is found in many roadcuts.

On carbonate rocks two kinds of mantle can be differentiated: unit c (pl. 2), a thin mantle generally less than a few feet thick consisting of silty saprolite and colluvium in which fragments of limestone are abundant and through which project many outcrops, and Unit t (pl. 2), a thick residual mantle—in places more than 50 feet thick—that is commonly oxidized and contains fragments of siliceous materials like chert and sandstone in addition to clay, silt, and sand.

Saprolite.—In this report the term "saprolite" refers to unconsolidated material (1) that is derived from the rock on which it rests, (2) that preserves almost intact the structure of the original material, and (3) that has undergone little or no loss of volume on weathering. Such material is characteristic of weathered igneous and metamorphic rocks but also occurs on some very impure carbonate rocks.

Residuum.—According to the usage of King (1943, p. 54), residuum is a residual mantle that has undergone loss of volume on weathering and occurs more typically on shale and carbonate rocks. The original structures of the rock are partially preserved in much of it, but in some areas the residuum is chaotic in structure or completely structureless. Structureless material that may have moved slightly from its original position commonly rests on top of less disturbed material. Most of the residual mantle in the Shenandoah Valley belongs in the category of only slightly disturbed residuum.

Colluvium.—The term colluvium refers to material that has moved downslope some distance by creep or slope wash. It may be slightly sorted as a result of its movement. As colluvium is generally in very thin small bodies on the surface and generally cannot be differentiated from residuum and saprolite in reconnaissance mapping, the term residual mantle as used herein includes all three kinds of material.

In the descriptions that follow, color designations are based on the "Rock Color Chart" of the National Research Council (Goddard, 1948).

RESIDUAL MANTLE ON RESISTANT ROCKS IN MOUNTAIN AREAS

The residual mantle in mountain areas is not subdivided in plate 2. Most of it is thin and stony, and in places it is lacking entirely. The character of the mantle depends on the topography as well as on the kind of rock. In the mountains the hollows or hopper-shaped areas at the heads of streams have a considerable size. Runoff in such places during heavy rains is sufficient to do much sorting and to remove most of the fine soil particles. The hollows are generally floored by fields of rounded to subrounded cobbles and boulders that are the residue of material that has weathered on the slopes above and that has concentrated above the stream (Hack and Goodlett, 1960, p. 6, 13). The size of the boulders varies depending on the kind of rock. The coarsest are found in the granitic rocks of the Blue Ridge, where the average size of the boulders in one hollow is 450 mm. The size is smaller in the Catoctin Greenstone as well as in sandstone and quartzite. In rocks of the Chemung Formation, which contains much sandstone, the boulder fields in hollows have an average size of 100–150 mm. They are about the same size in the Hampshire Formation. In the Pocono Formation areas, they are intermediate in size, averaging 200–300 mm.

On side slopes and noses the residual mantle is finer grained, and stones are mixed with finer weathering products such as sand, silt, and clay. Outcrops and cliffy slopes occur in places where especially resistant beds such as massive quartzite occur. Such slopes may be covered with scree or have extensive deposits of talus at the foot. These deposits have been given special attention and are described on page 32. The general character of the mantle on the principal rock types in the mountains is now briefly described.

Mantle on metabasalt.—The Catoctin Greenstone, or metabasalt, is one of the most resistant rocks of the Shenandoah Valley. It forms the crest of the Blue Ridge for many miles. The mantle is generally thin and stony, but varies considerably in thickness and stoniness. The material is generally light-brown stony clay loam (5*YR* 5/6) containing many basalt fragments of pebble and cobble size. Insofar as known, thick masses of either residuum or saprolite are rare. Bedrock outcrops are numerous, stones are scattered on the surface, and scree is common on cliffy slopes.

Mantle on sandstone and quartzite.—Sandstone and quartzites occupy high areas in many parts of the Valley and outcrop extensively in the foothills west of the Blue Ridge, on Massanutten Mountain, and on the high ridges to the west of the lowland belt. The rock formations involved include the Chilhowee Group, the Tuscarora Quartzite, and the Pocono Formation. The mantle is physically much like that on metabasalt. Outcropping ledges are abundant, the soil is thin and stony, and the surface is littered with fragments of quartzite.

Mantle on interbedded sandstone and shale.—Large areas are underlain by rocks less resistant than the massive sandstones and quartzites but nevertheless support considerable relief. These rocks include argillite, like the Harpers Formation in the Chilhowee Group, and interbedded sandstone and shale, like parts of the Chemung Formation, the Hampshire Formation, and the Clinton Formation on Massanutten Mountain. On weathering, such rocks produce abundant resistant fragments that are concentrated on the slopes. The fragments creep downward and provide an armor that protects the underlying weathered shale from erosion, and therefore a fairly thick residuum consisting of stony loam is produced. In general the mantle is considerably thicker on these rocks than on more resistant rocks that contain less shale. Rock outcrops are few. Many exposures in landslide scars in the Hampshire Formation indicate that the mantle averages 3–4 feet thick. On Massanutten Mountain, where the Silurian quartzite occurs upslope from the Martinsburg Shale, sandstone fragments creep down the mountainside and form a thick mantle that covers a clayey residuum derived from the shale. In one road cut 40 feet of such mantle is preserved.

Mantle on granitic rock.—Several areas of considerable size in the Blue Ridge are underlain by coarse-grained igneous rock, mostly granodiorite. Steep slopes are strewn with boulders. The noses and ridge crests are rocky and have some outcropping ledges, whereas the hollows are filled with large subspherical boulders. Valley-side slopes are also bouldery, but the boulders cover less than half of the ground. Roadcuts or sand pits reveal the material beneath the surface in a few places, as at the head of Hawksbill Creek on a road that leads to Skyline Drive. Fresh granodiorite here is overlain by coarse sandy saprolite. The saprolite is light brown (*5YR* 5/6), and 2 feet from the surface it has a pH of about 4.5. The saprolite contains many subrounded, subelliptical boulders some of which exceed one-half meter in diameter. Such boulders have been termed core boulders (Scrivenor, 1931) or core stones (Linton, 1955; Ruxton and Berry, 1957). They are relict boulders in the grus or sandy saprolite that have survived the gradual process of spheroidal weathering (fig. 16).

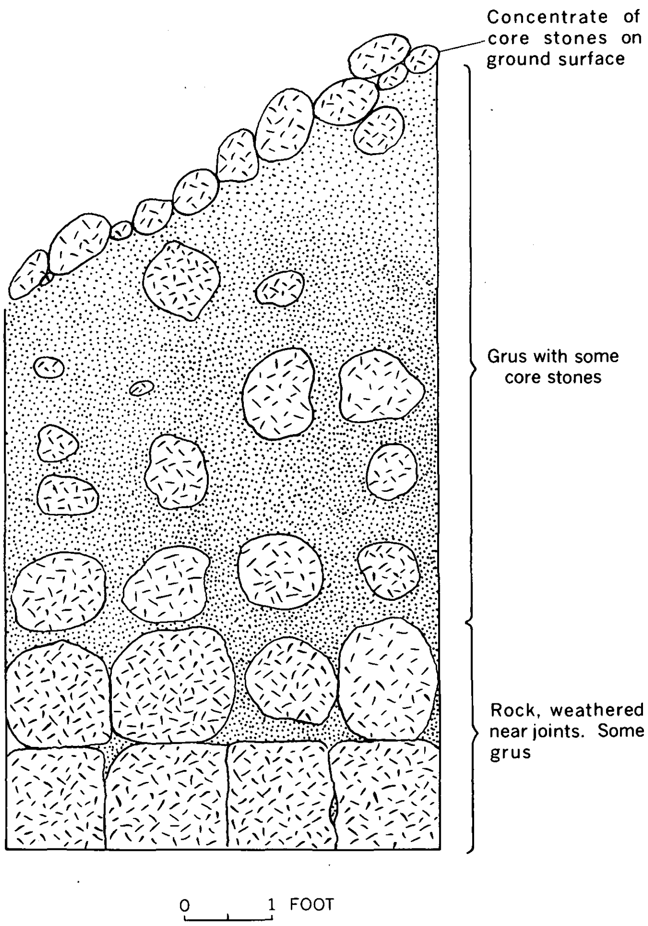

FIGURE 16.—The hypothetical formation of core stones on a mountain slope.

In massive crystalline rocks, weathering begins along joint planes and gradually works inward; so, rounded boulders, which are the remnants of former joint blocks, are left (fig. 16). As the saprolite and grus are eroded from the surface of the ground and the weathering of the rock proceeds to greater depth, the core stones may collect as a lag concentrate on the surface. Being less readily removed by creep and wash than the grus, they form an armor beneath which the saprolite and grus are protected and preserved.

SCREE

Large block fields that are treeless or only sparsely covered with trees and shrubs are common features in the mountain areas. Some of these block fields are composed of a typical talus that has accumulated at the base of cliffs of resistant rock. Others consist of blocks that have accumulated on the rock from which they were derived and that have moved very little. Accumulations of coarse rock fragments of this kind have been called scree (Stokes and Varnes, 1955). In the Shenandoah Valley, scree occurs on the principal ridge-making sandstones and quartzites and on the Catoctin Greenstone. It is most abundant in the outcrop area of the Chilhowee Group, particularly on the Antietam Quartzite and downslope from the contact between this group and the Harpers Formation. Scree is abundantly associated with the Tuscarora Quartzite of Silurian age and occurs especially on Massanutten Mountain, where it accumulates at and below the contact between the Tuscarora and the Martinsburg Shale. Scree also occurs sporadically on the Pocono Formation.

In the Blue Ridge of Virginia, scree is thought by some geologists to be relict from the cold climates of the Pleistocene, and it has been regarded as proof that a cold climate existed there (Smith and Smith, 1945). The production of scree probably is favored by a cold climate, and the writer has observed that scree is much less extensive and is in fact a rarity on the Weisner Quartzite of Early Cambrian age exposed near the southern edge of the Appalachian Mountains in

Alabama.[1] In the Appalachians of Pennsylvania, on the other hand, it is more widely formed on the resistant quartzites than it is in Virginia.

Probably the Pleistocene climate in the Shenandoah Valley was much colder than today's climate, and some evidence for this has accumulated. For example, dune areas apparently of Pleistocene age have been found in what are now forested areas of the Maryland Coastal Plain (Hack, 1955). Pollen of fir and other northern species has been found at depth in bogs on the Coastal Plain as far south as South Carolina (Dillon, 1956). Dillon believed that the pollen and other ecological evidence suggest that Appalachian Highlands had at least a Canadian-type climate during the Pleistocene.

On the other hand, as shown in the following pages, evidence exists that some scree is forming today in the Shenandoah Valley and that it moves downslope. Probably the processes acting in the region at the present time are effective in producing and moving scree and the climate is probably cold enough for frost riving of the quartzite to occur. Periglacial climates and the existence of perennially frozen ground are therefore not necessary conditions for the formation of scree, though an intensely cold climate might greatly favor its production.

Studies of scree derived from the Antietam Quartzite were made in several areas where scree is particularly abundant and where many areas of bare scree are preserved.

ST. MARYS RIVER VALLEY

One of the most spectacular areas of scree is in the St. Marys River valley (fig. 17), where the scree lies on or immediately downslope from its source in the Antietam Quartzite. The St. Marys Valley contained manganese mines for many years, and it was served by a spur railroad that connected with the Norfolk and Western Railway at the mouth of the St. Marys River. Ballast for the road bed was quarried from a scree on the north side of the valley (fig. 17, loc. 917). The exposure shows the internal character of a scree deposit. The Antietam Quartzite dips toward the river only slightly less steeply than do the valley sides (fig. 17). Scree derived from cliffy ledges high on the slope and moved down by gravity collects on the quartzite. It extends from the valley floor to a height of approximately 150 feet on a slope of about 36°. Quarrying operations were carried on at the foot of the scree, and as cobbles and boulders were removed, slides took place. As a result of the operations, vegetation was completely removed from the lower part of the scree, and the protecting cover of coarse blocks was also removed from parts of it.

The weathered scree surface on the undisturbed part is almost white but generally has a slightly pinkish-gray cast (5*R* 8/2), except where modified by lichens or other vegetation. The fresh scree uncovered by quarrying is pale yellowish-orange (10*YR* 8/6). The greatest difference, however, is in the size of the rock fragments. The upper few feet of undisturbed scree consists of boulders that average about 120 mm in diameter. At least half of the material beneath the surface is sand, and the mean diameter of the exposed particles is 3.0 mm.

A cross section of another scree on the south side of the St. Marys Valley (fig. 17, loc. 918) is shown in figure 18. This slope is on the south limb of the syncline, and the situation of the scree almost mirrors the conditions at the scree quarry (fig. 17, loc. 917), except that less of the relatively soft Harpers Formation is exposed at the base of the slope and the slope is undercut by the river with the result that a thick deposit cannot collect. The scree is formed just below outcropping ledges of resistant quartzite. Vegetation covers most of the slope, but bare patches occur; beneath the ledges where the scree appears to be coarsest, these patches form an almost continuous belt or series of belts running horizontally around the mountain.

Where vegetation occurs the scree is finer grained and the surface is more stable and smoother than on bare scree. The difference is not very striking but was demonstrated by measurement of boulders in sample plots along a tape traverse. Sample areas were selected at nine localities about 60 feet apart on the scree surface, four in vegetation-covered areas and five in bare areas. At each locality 20 boulders were selected at 2-foot intervals on a tape and grouped by size classes. The data are shown in table 3 and support the observation that the scree is coarser in the bare areas.

The vegetation in the area is not dense, and rock fragments in the vegetation-covered areas are visible on the rocky ground, although some patches are completely covered with moss or humus. The forest is typical of the yellow-pine forest characteristic of dry sites defined by Hack and Goodlett (1960) in the Little River area. It consists of pitch pine, sassafras, black gum, scrub oak, chestnut oak, red maple, rhododendron, and various ericaceous shrubs. Lichens are abundant in both bare areas and vegetation-covered areas but are

[1] In May, 1961, the writer visited Choccoloco Mountain near Jacksonville, Ala., and Dugger Mountain, south of Piedmont, Ala. Both are underlain by the Weisner Quartzite. The maximum relief in these areas is about 1,000–1,200 feet and is judged by the writer to be comparable with and only slightly less than the relief in the Paine Run area described herein. Areas of scree exist, but on a very small scale, and most of the blocks scattered on the ground surface are subrounded boulders that are apparently much weathered.

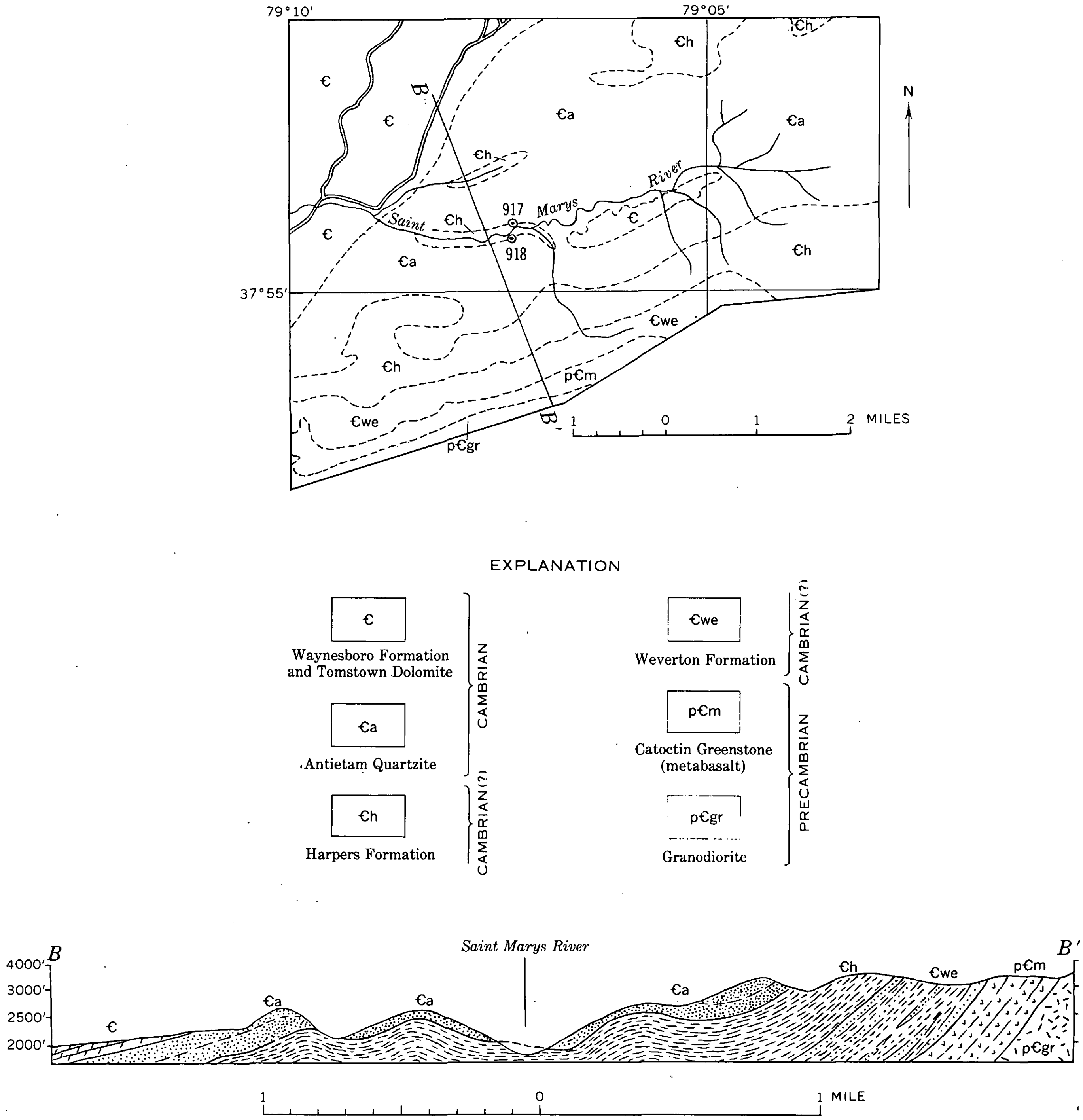

FIGURE 17.—Simplified geologic map and section of the St. Marys River valley showing the localities described in the text. Redrawn from Knechtel (1943, pl. 29).

more abundant in the vegetation-covered areas. Rock fragments in the vegetation-covered areas are more rounded and weathered in appearance than in the bare areas. There is no evidence at this locality of any recent movement of the scree or of recent rock falls or slides.

The St. Marys River, in its bed and flood plain, has coarse blocks of quartzite, much of which may be derived from the reworking of scree. The channel shown at the base of the scree in figure 18 contains blocks having an average size of 140 mm, which is only slightly smaller than the size of fragments of the scree above, and inspection of other reaches of the river show that this sample is typical of the riverbed. About 10 percent of the fragments in the riverbed is more than 500 mm in

TABLE 3.—*Comparison of size of fragments in samples of scree from vegetation-covered and bare areas*

[Individual samples measured along separate traverses]

Area	Geometric-mean diameter of fragments (mm)						Total number of fragments in all samples
	Individual samples					All samples	
	1	2	3	4	5		
Vegetation-covered	98	140	130	85	------	115	79
Bare	200	160	210	200	174	190	100

diameter, which is larger than the diameter of fragments found in many, though not all, of the scree areas.

Investigations in this valley demonstrate the following: (1) Scree forms directly on or directly beneath cliffy slopes on the rock of which it is composed, (2) where vegetation covered, the blocks of scree become rounded and reduced in size by weathering, (3) the principal stream has on its bed and, when in flood, presumably moves material of sizes as large as the blocks on the valley sides. The movement of this material by the river suggests that the screen areas are not static but are being eroded by present processes.

PAINE RUN AREA

An area of about 15 square miles in Shenandoah National Park containing a large outcrop area of massive quartzite and many large areas of scree was studied by the writer. Much of the scree is on the outcrop from which it is derived, but some has moved downslope, 1,000–1,500 feet. A geologic map and section of the area are shown in figure 19.

The oldest formation within the area is the Catoctin Greenstone, exposed on the east side of the Blue Ridge about 500 feet below the crest. This rock consists of massive metabasalt much like the Catoctin Greenstone in the Elkton area (King, 1950). The Catoctin is overlain by a thick sequence of sedimentary rocks consisting of siltstone, sandstone, argillite, and some thin dark-colored quartzites; the sequence is not differentiated but comprises the Loudoun, Weverton, and Harpers Formations. The total thickness of these rocks is about 2,200

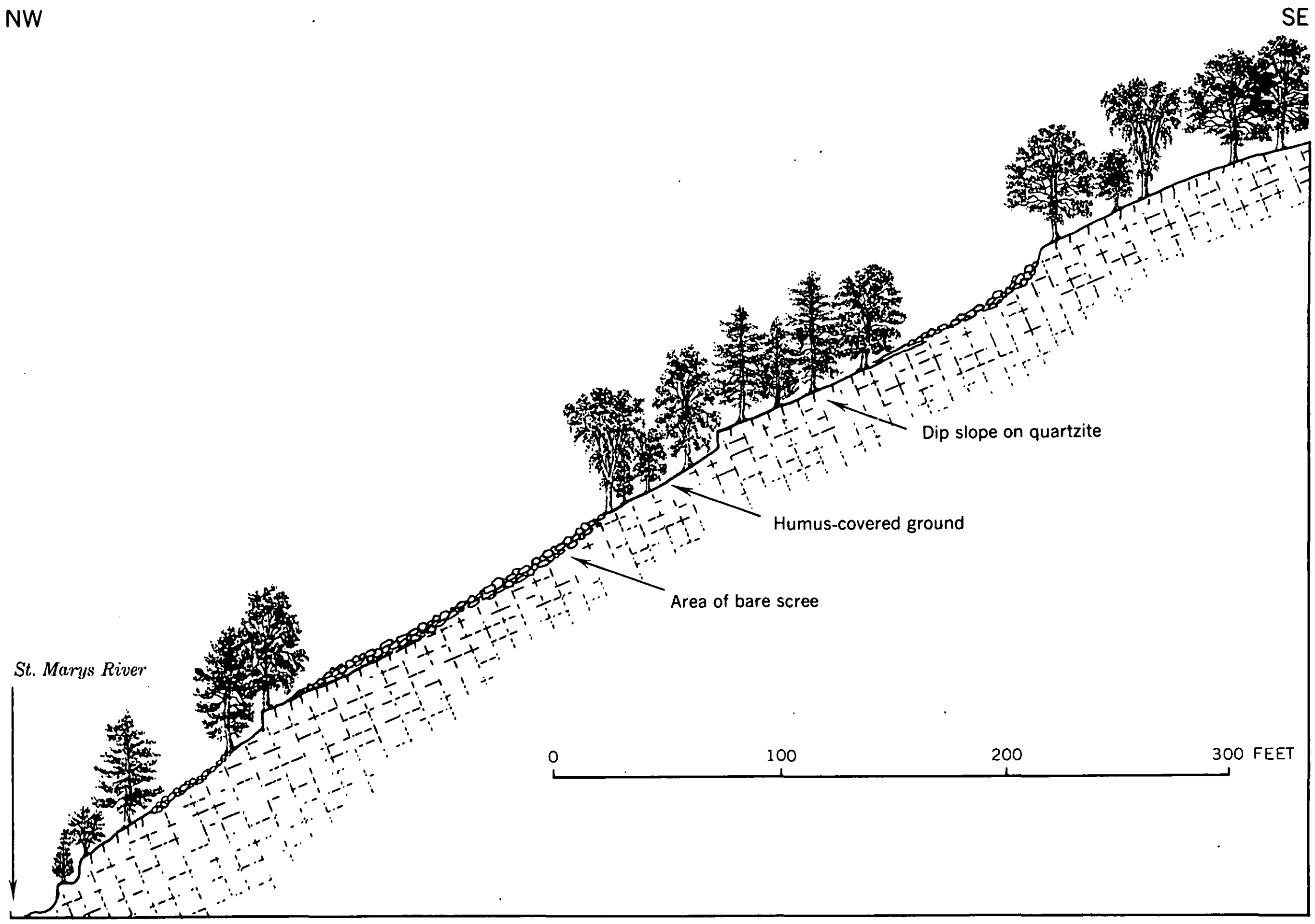

FIGURE 18.—Slope and scree on south side of St. Marys River valley (loc. 918 described in the text). Traverse made with tape and hand level.

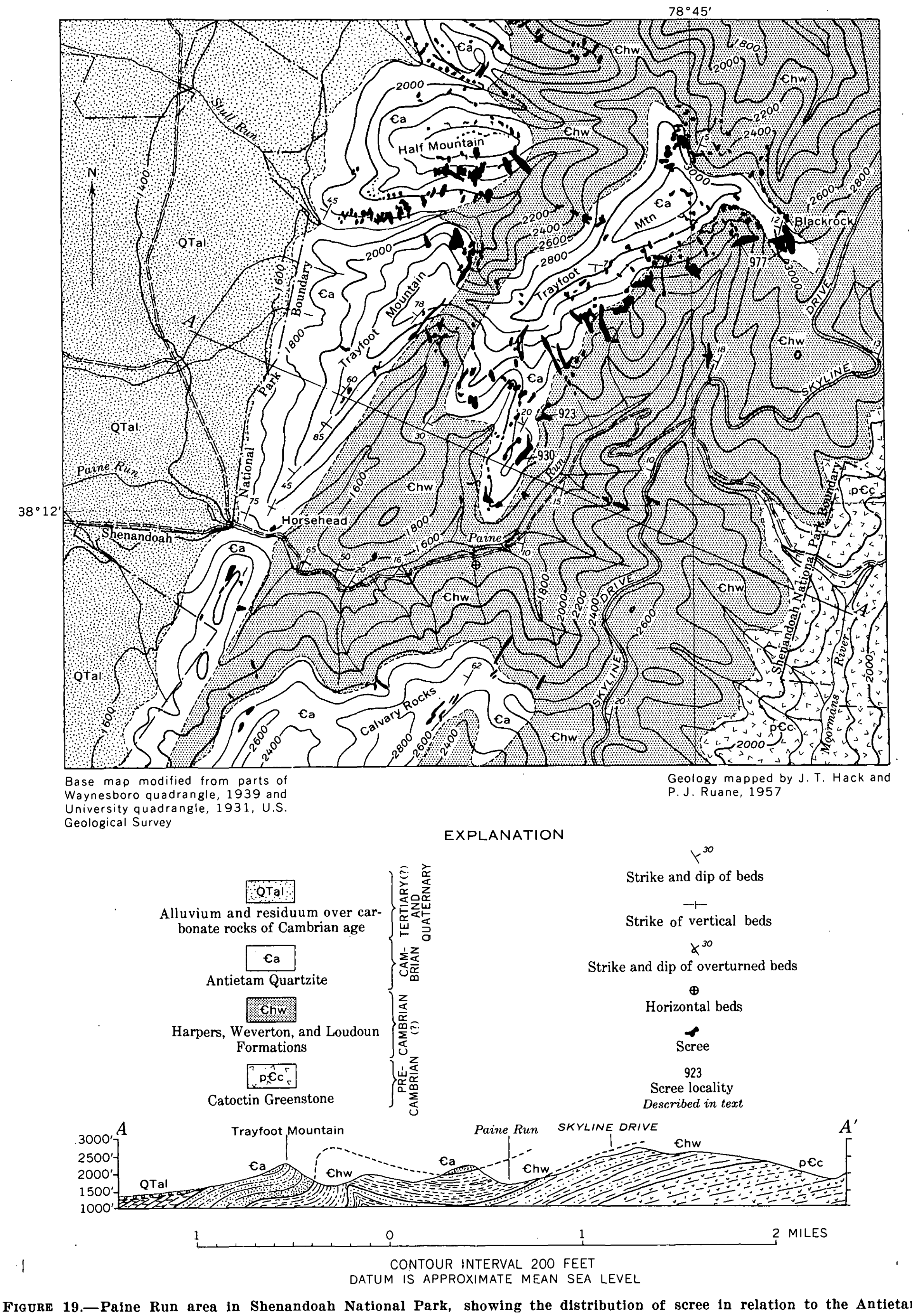

Figure 19.—Paine Run area in Shenandoah National Park, showing the distribution of scree in relation to the Antietam Quartzite and showing the scree localities described in the text. Datum is mean sea level.

feet. The upper 800 or 900 feet is primarily fine grained sandstone and siltstone that is tough but brittle and highly fractured. It contains a few thin bands of quartzite. The Antietam Quartzite overlies and contrasts sharply with the Harpers, as it is much tougher and less fractured.

Unfortunately, no complete section of the Antietam was found within this area, but fairly good exposures of much of the Antietam occur in the narrow gap at Horsehead, where Paine Run leaves the Blue Ridge. About 1,000–1,100 feet of quartzite is partially exposed on the mountainside south of Paine Run. The basal beds of the Antietam are of particular importance because they furnish scree-forming material throughout the area. They consist of about 100 feet of very pale orange to pinkish-gray extremely dense fine-grained massive quartzite. The basal beds are overlain by 800 feet of quartzite, most of which has a well-defined prismatic cleavage at right angles to the bedding planes. It ranges from dense to coarsely granular. Scolithus tubes are common at several horizons within the Antietam, especially 360–420 feet above the base, 650–700 feet above the base, and near the top of the formation 900–1,000 feet above the base. The scolithus tubes are at right angles to the bedding and parallel to the prismatic cleavage. The upper 100 feet of the formation exposed at this place is a friable well-bedded sandstone cemented by calcium carbonate. On the basis of the relation of the outcrops on ridge tops to the boundaries of the formation, it appears that the most important ridge-making beds in the formation are the upper scolithus unit and the dense massive beds directly above the Harpers.

Many of the scree areas shown in figure 19 were visited by the writer during the mapping of the bedrock geology, but their boundaries were delineated by photointerpretation. The map shows only the scree areas on which dense vegetation is absent. Nevertheless, these areas include most of the large deposits of coarse scree. Others exist, and in places vegetation grows on the border of or in patches on the open scree. Many other scree areas have become weathered and are partially soil covered and overgrown with forest. The map (fig. 19) probably shows the areas that have thick and highly porous deposits of blocks on which vegetation cannot take hold. Ground traverses over the entire area revealed that forested areas not mapped as scree for the most part have much finer grained sandy material between the blocks and a rather smooth ground surface on which footing is good. The soil might be called a very coarse stony loam rather than a block field.

The map brings out several facts relating to the origin of the scree. Bare scree is more extensive on south-facing than on north-facing slopes. Note, for example, the two sides of the valley of Stull Run as well as the valley to the north of it. Trayfoot Mountain has extensive scree on the south slopes and much less on the north slopes. The prevalence of scree on south-facing slopes is confirmed by examination of aerial photographs of other areas in the Blue Ridge. The reason for the relation are not known, but south-facing slopes are generally drier than north-facing slopes (Hack and Goodlett, 1960). A possible explanation is that the breakup of the rock because of chemical weathering or frost action may be more effective on the wetter and colder north-facing slopes and scree is therefore not as readily preserved. On the other hand vegetation may grow among coarser blocks on north-facing slopes and scree occurs there but is no longer bare.

Scree is most abundant on the outcrops of the Antietam Quartzite itself; however, many areas of scree form at the base of the Antietam, and some scree has moved downslope onto the Harpers Formation—for example, on Trayfoot Mountain (fig. 19, loc. 923). These scree areas are commonly associated with hollows at valley heads, where runoff is concentrated (fig. 20).

Several examples of activity on the scree areas were noted. A scree near the south end of Trayfoot Mountain (fig. 19, loc. 930) shows evidence of recent movement. This scree is formed on the Antietam Quartzite itself. A section, drawn to scale, of the scree measured and sketched in the field shows the relations at the site (fig. 21). A distinctive sorting of the boulders on the ground can plainly be seen. The scree at the crest of the hill is finer grained that the scree further down; a very slight scarp or sharp drop separates them. At the base of the scree, a narrow ridge parallels the contour of the hill, and several large trees growing in the ridge are tilted toward the scree (fig. 22). The tilted trees are evidence of a downhill movement of the blocks. The ridge is separated from the main body of the scree by a row of oval depressions in the boulder field. The kind of motion that occurred at this locality is not clearly evident, but the bottom part of the scree apparently slid downward as a mass or sheet and forced up a ridge at the tree-covered lower edge. The tilted trees on the low ridge are judged by the writer to be more than 50 years old. The lichen cover on the boulders is less than 75 percent and is considerably less complete than the cover on many other areas of scree. These facts suggest that the last movement of the scree occurred more than 30 or 40 years ago but considerably less than 100 years ago.

A scree on the Harpers Formation was studied at locality 923 on Trayfoot Mountain (fig. 23). This

FIGURE 20.—The Paine Run area showing two kinds of scree. The large scree area on the right (about 200 ft across) is locality 923 (described in the text) and is in a hollow.

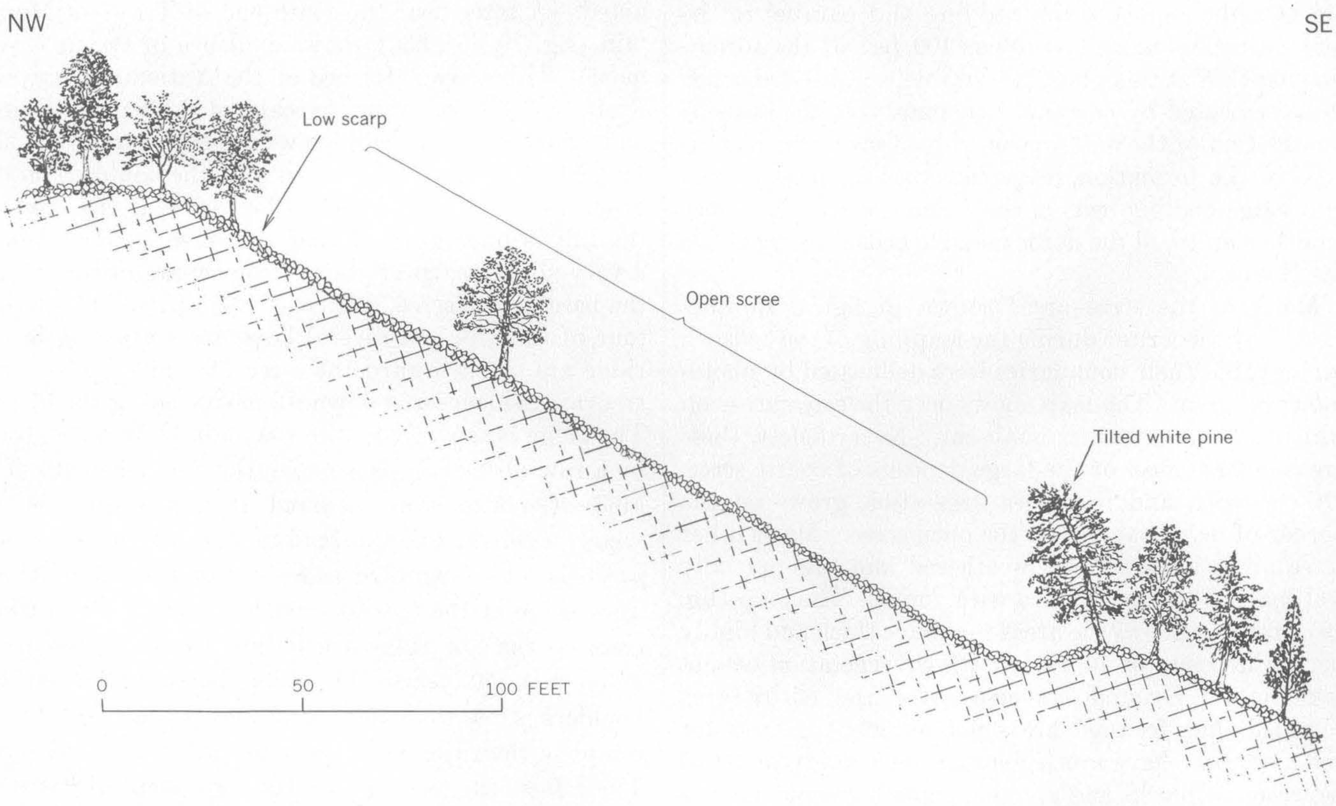

FIGURE 21.—Scree on southeast slope of Trayfoot Mountain at locality 930 (described in the text), showing evidence of recent movement of the scree.

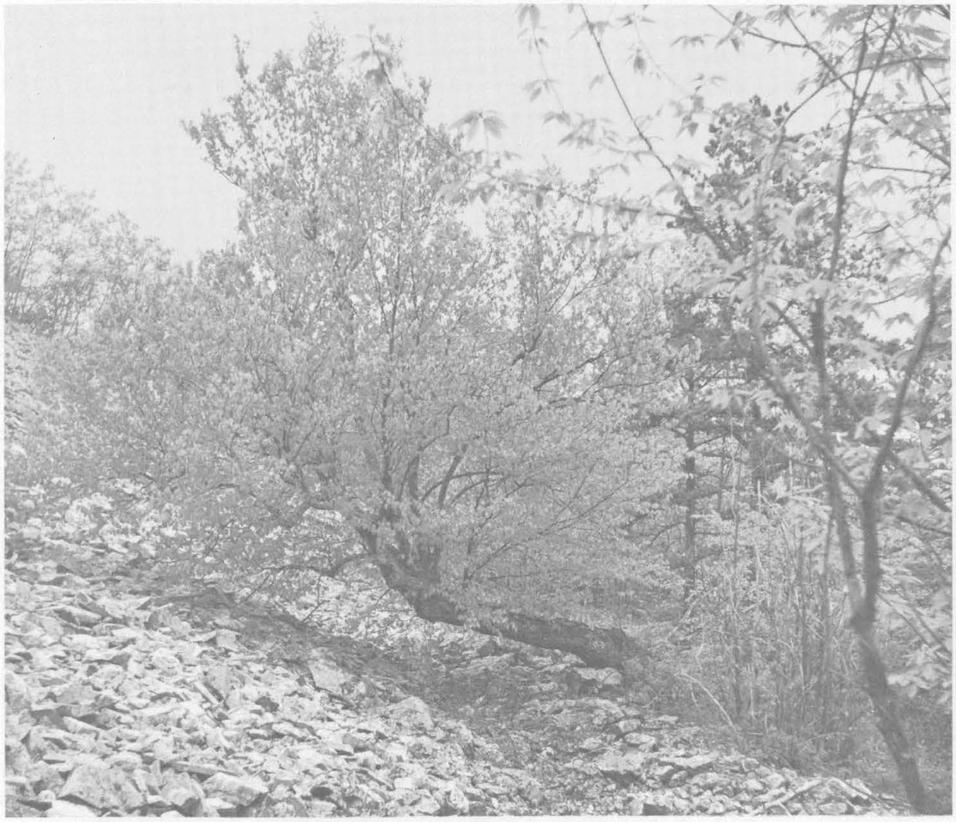

FIGURE 22.—Tree at base of scree tilted by sliding.

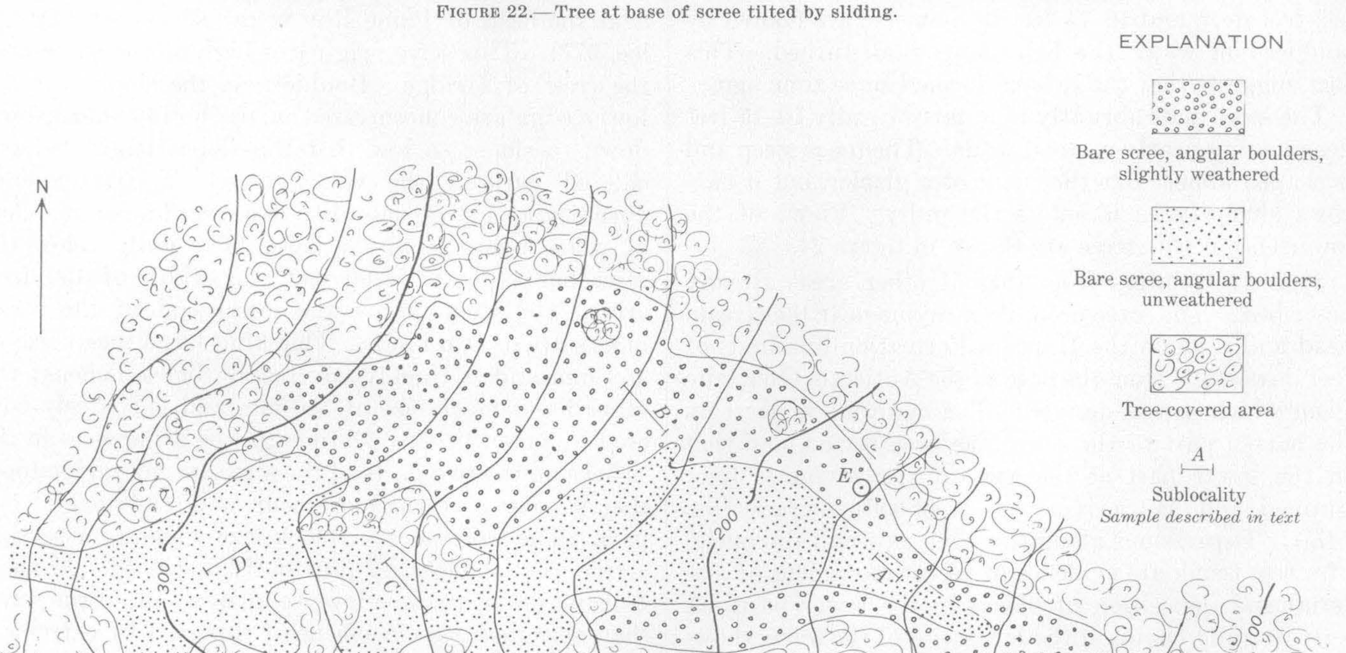

FIGURE 23.—Scree on west side of Paine Run at locality 923 (described in text) showing sample sublocalities.

scree is composed entirely of blocks of dense quartzite typical of the basal beds of the Antietam that crop out upslope from it. The bedrock, as shown by an outcrop 50 feet south of the scree, is a different rock consisting of greenish-gray shale or siltstone of the Harpers Formation. The blocks in the scree are fresh and lichens are sparse, especially at the lower end of the scree. The scree consists of two parts (fig. 21): One area of scree occupies the bottom of the ravine and has a slope of 17° at the lower end and 19° or 20° at the upper. The geometric-mean diameter of this material is 90 mm at sublocality A, 100 mm at sublocality C, and 170 mm at sublocality D. The other area, on the north slope of the ravine, is somewhat steeper and is composed of distinctly coarser material. A sample at sublocality B has a mean size of 190 mm. The lower half of the scree is almost completely devoid of vegetation of any kind, and the rock is fresh and angular. Green lichens cover about 70–75 percent of the deposit. A sassafras tree at sublocality B provides evidence of recent movement of the scree. The trunk of the tree is buried in the rubble of cobbles to a depth of half a foot. Several small rootlets have sprouted from the trunk, indicating that the rubble has moved enough to bury the lower part of the trunk within the recent life of the tree. Several small depressions between sublocalities C and D 1–2 feet deep and 10–15 feet in diameter are floored by boulders on which the lichens are undisturbed. This fact suggests that the lichens formed some time ago.

The scree ends abruptly in a narrow gully 10–15 feet deep having nearly vertical walls. The toe is steep and is shaped almost like the snout of a glacier, but it narrows abruptly as it enters the gully. Views of the lower end of this scree are shown in figure 24.

This scree differs from that of other areas already described. The scree occupies a ravine near the stream head and rests on the Harpers Formation hundreds of feet downslope from the base of the Antietam Quartzite from which it was derived. The material, at least on the barren part of the scree, has been sorted, and that in the lower part at the axis of the ravine is finer grained and has a remarkably smooth surface (fig. 24*B*). Depressions along the scree axis and the burial of a tree trunk are evidence of recent movement of material and deposition of blocks in the lower part. A narrow gully begins at the lower end of the scree, showing either that the scree has slumped down the ravine into the gully, that large flows of water issuing from the scree have cut the gully, or both.

The writer believes that the scree is genetically related to the debris avalanches common in the Appalachians, such as those produced by the 1949 flood of the Little River west of Harrisonburg, Va. (Hack and Goodlett, 1960). During this flood large masses of cobbly and blocky rubble were dislodged from the upper slopes, carried down into hollows and channelways, and redeposited downstream. In some areas, however, the rubble was arrested at the stream head or in the channelway, a condition producing fields of blocks or boulders in the floor of the ravine (Hack and Goodlett, 1960, p. 44 and pl. 5*A*). The scree at locality 923 may be a lag deposit of blocks that have moved downslope into the ravine during periods of excessively heavy runoff. During such periods the ravine would be deepened and the boulders sufficiently loosened to move downslope. With sufficient runoff, some sorting of the blocks would occur. This explanation would account for the location of the scree area hundreds of feet downslope from the source rock, the snoutlike appearance of the lower end of the scree that is advancing onto the gully, and also the hollows at the valley axis caused by erosion of the ravine bottom beneath the scree. Since the scree is not now connected with its source rock but lies far below, presumably the area of scree is not expanding at the present. Some boulders may be added during floods, but the net result of the process may be to move the scree further from its source.

Another area of scree similar to that in locality 923 is at the head of Paine Run below Blackrock (fig. 19, loc. 977). This scree originates high on the slope near the crest of a ridge. Boulders on the slope near the lower edge are concentrated in the hollow and extend down to slopes so low that the slopes might be considered part of the valley floor. The scree ends abruptly at a steep snoutlike slope, and a copious flow of water issues from the base. The gully below the scree has a depth of 60 feet and width of 150 feet. About 100 yards above the lower end of the scree, blocks are of two kinds.. One kind has a mean size of 450 mm and is thickly covered by black lichens; the other has a mean size of 150 mm and is sparsely covered by green lichens. Although the differences in the lichens on the two parts of the scree are not understood, they suggest a recent movement or disturbance of one part. The difference in block size suggests a sorting action as at the locality previously described. As these features occur near the lower end, or snout, of the scree, they may indicate movement of this part by extremely high flood runoff in the ravine.

MADISON RUN

Accumulation of boulders on the surface of a scree was observed just north of the Paine Run area on the north slope of the valley of Madison Run, east of Grottoes (pl. 2). This area was visited by the writer

A

B

FIGURE 24.—Scree at locality 923 (described in the text). *A*, View from the gully at the lower end of the scree; *B*, view from the surface of the scree downslope toward the distal, or lower, end.

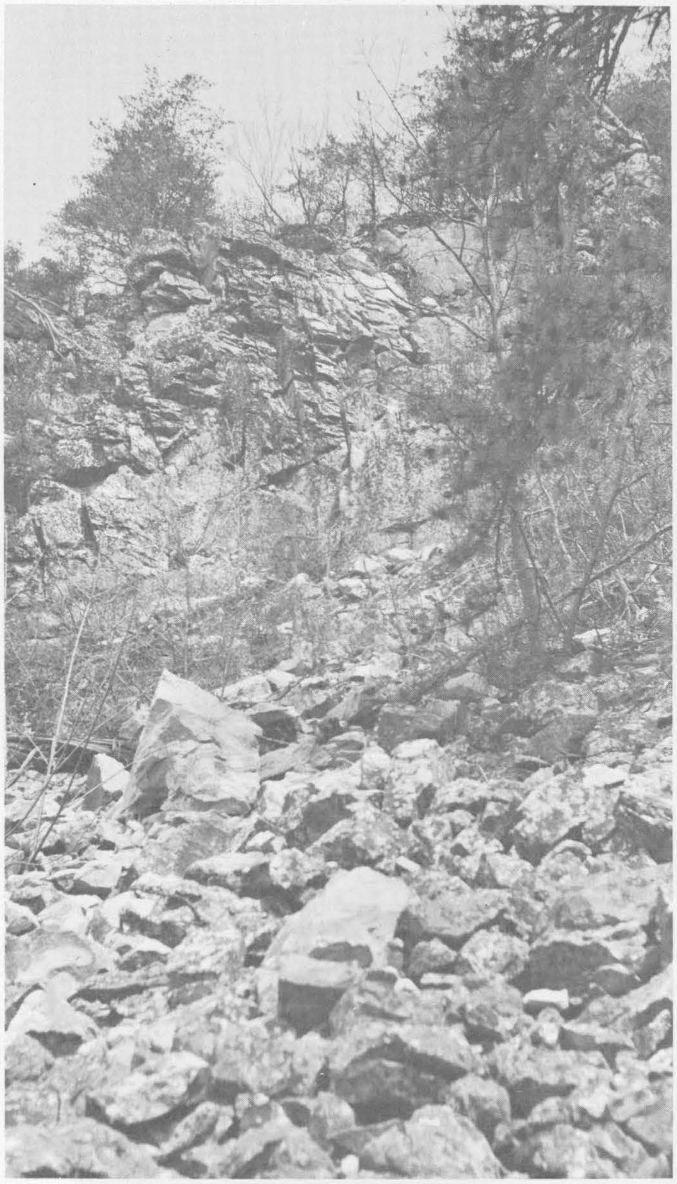

FIGURE 25.—A cliff and scree at its base, which includes fresh debris that fell during winter of 1957–58.

in the fall of 1957 and the spring of 1958. During the spring, a fresh rockfall was noted at the top of the slope. A block of fresh quartzite approximately 7,500 cubic feet in volume had fallen from a vertical cliff about 30 feet above a large scree having a slope of about 35°. Fresh debris on the scree surface was evidently derived from the rock fall and was scattered downslope for a distance of 200 feet. It was broken into blocks about the size of the older scattered blocks on the scree surface (figure 25).

INTERPRETATION OF SCREE

The data presented indicate some activity on the scree areas of the Blue Ridge at present. The activity includes processes that tend to accumulate scree as well as processes that move, weather, and erode it. The relative rates of these processes may have varied considerably in the past during times of different climate. Scree seems to accumulate today on slopes below steep cliffs. Since many bare scree areas composed of angular boulders lie immediately below steep cliffs, the rate of accumulation appears to be too rapid or was too rapid in the recent geologic past to permit vegetation to gain a foothold. Evidence in the Paine Run area at locality 930 (fig. 21) indicates that at present scree moves downslope by some kind of mass sliding. Evidence at localities 923 and 977 shows that flood runoff in the ravines is at times competent to move large masses of blocks of the size that composes the scree. Furthermore, streams like the St. Marys River and Paine Run transport material of this size, as is indicated by the blocks and boulders that line their beds.

On the other hand some areas of scree, like locality 923, have no present connection with any source area, and they are evidently moving downvalley away from the Antietam Quartzite from which they were derived. Many areas like those described in the St. Marys Valley are covered with vegetation and are undergoing weathering.

The present climate is cold enough so that frost may be an important agent in shattering exposed bedrock and could be a factor producing scree. In accordance with the suggestion of Troll (1944, 1958), the climatic records at two stations were examined to determine (1) the frost-free days (defined herein as the number of days per year in which the temperature does not drop to 32°F), (2) frost-alternation days (defined as the number of days per year in which minimum temperature is 32°F or below and the maximum temperature ture 32°F or above), and (3) the ice days (defined as the number of days in which the maximum temperature is 32°F or below). The results are shown in table 4. One of the stations, Luray, is in the Shenandoah Valley about 8 miles west of the crest of the Blue Ridge; the other, at Big Meadows, is on the crest of the Blue Ridge. The two stations are probably representative of the warmest and coldest parts of the Blue Ridge area.

Freeze-thaw frequencies in Canada were studied on a regional basis by Fraser (1959), who showed that the number of days that alternate between freezing and thawing increases to the south. Comparison of Fraser's data with the freeze-thaw maps of Visher (1954, maps 322, 323, and 325) indicates that the frequency of freeze-thaw days in eastern North America is at a maximum in the northern Appalachians and the Great Lakes region and decreases both to the north in Canada and to the south in the United States. A

TABLE 4.—*Occurrence of freezing temperatures at two stations in the Blue Ridge area, Virginia*

Year [1]	Frost-free days	Frost-alternation days	Ice days
Luray			
1955–56	239	124	2
1956–57	266	96	3
1957–58	229	123	13
Average	245	114	6
Big Meadows			
1955–56	228	94	43
1956–57	249	90	26
1957–58	215	107	43
Average	231	97	37

[1] July 1–June 30.

belt in which freeze-thaw days occur with relatively high frequency extends down the Appalachian Mountains, and the Blue Ridge area of Virginia must lie in one area of the belt. The frequency of occurrence is apparently somewhat higher in New England.

At present, the number of frost-alternation days in the Blue Ridge area is relatively high. On the other hand, the number of ice days, as indicated by table 4, is very small; so, the depth of frost penetration in the ground cannot be very great, and the materials affected by frost must be confined to a thin surface layer. Because of the large number of frost-alternation days, the climate may be well suited to the frost-shattering of exposed rock surfaces. Under an insulating cover of vegetation, soil, or humus, however, it is questionable whether frost would have any effect whatever, especially at lower altitudes.

The concept of dynamic equilibrium indicates that the area of the Blue Ridge containing scree is a system in equilibrium, in which the material eroded and carried out of the system is balanced by the material furnished by rock shattering. The balance is not complete and exists only if the erosion and rock shattering have occurred over long periods of time. If the problem is examined in this light, several factors must be as important as frost in producing and maintaining the scree: (1) the area must have beds of very tough chemically inert rock that is reduced to small sizes very slowly either mechanically or chemically, (2) the relief and erosional energy of the area must be great enough so that the resistant rock stands high and forms many steep cliffs, and bare areas must exist on which frost riving can occur, and (3) the climate must be cold enough so that the rocks are weathered mechanically. These factors are balanced by others that remove or destroy the scree, including creep, transportation by flood runoff, and rock weathering under forest cover.

The three scree-producing factors enumerated are important only relative to one another. The more resistant the rock or the more rugged the relief, the less necessary is an intensely cold climate. Conversely, the less rugged the relief or the less inert the rock, the more important is the cold climate. If the relief in the area were less, bare areas of scree could not long be maintained, for vegetation could gain a foothold on the more stable slopes, and weathering of the scree and reduction to finer sizes could proceed more effectively. Similarly, if the rock were less resistant to weathering, the rate of destruction of scree would be higher, even with the present rugged relief.

The effects of temperature are particularly important. If the temperature were drastically reduced but the rate of the destructive processes remained the same, the rate of production of scree on the existing steep slopes would be increased, with a resulting increase in the area of scree. For a time the system would be out of balance, and the rate of accumulation would exceed the rate of removal. A balance would eventually be reached when the volume of scree reached such a size that the amount removed would balance the amount supplied by frost riving. An increase in the amount of scree would probably tend to reduce the relief and the area of cliffy slopes and thus itself bring about a balance. An increase in temperature would probably have the opposite effect; it would decrease the production of scree, increase the rate of weathering, and accelerate removal of scree, except possibly in hollows where floods might concentrate fields of blocks. These factors are all difficult to assess, especially as any change would affect the vegetation, itself a factor in the system. Nevertheless it seems certain that changes in climate would produce changes in the factors that affect the equilibrium and result in changes in the area of scree on the slopes. That scree is more abundant on quartzite in Pennsylvania and less abundant on quartzite in Alabama suggests perhaps even more strongly than the foregoing argument that a colder climate is more favorable for the production of scree than a warmer climate and that a change to colder conditions would increase the area of scree.

In conclusion, changes in climate must affect the equilibrium conditions that determine the areas and amounts of scree on the slopes. The factors involved are complex, and without knowing exactly what equilibrium conditions are under different climatic conditions, it is not possible to predict what changes might accompany a particular change in climate or even to assess what areas of scree have been inherited from the past. Nevertheless, many large areas of scree are probably preserved from the Pleistocene, and the present blocky mantle is the product of processes that have operated over a long span of time in which several climatic fluctuations may have occurred.

RESIDUAL MANTLE IN SHALE AREAS

The Martinsburg Shale (map unit s, pl. 2) occupies the center of the Massanutten syncline, and in the southwestern and northeastern parts of the Shenandoah Valley, it forms an extensive lowland area having high drainage density and low but steep-sided hills. Shales of Devonian age occupy large areas in Massanutten Mountain and in the mountains on the northwest side of the Shenandoah Valley. The Devonian shales generally support more relief than the Martinsburg Shale because they are on the flanks or foothills of mountain ridges composed of sandstone and quartzite. The residual mantle on the shale is light colored and generally very thin and contains abundant flakes and slivers of shale. Much of the area, especially in the Martinsburg Shale, is in pasture, and the flakes of shale are visible in the sod between tufts of grass. The contrast between this kind of area and the areas of other units mapped is so striking that the shales form a very easily mapped unit; the area of Devonian Shales is distinctive even where no exposures or cuts occur. Weathered bedrock is generally within a few feet of the surface, and material that can be called true residuum is commonly measured in inches rather than feet. A typical roadcut in an area of Martinsburg Shale is described as follows:

Section exposed in road cut 7.5 miles south of Harrisonburg

[Section reads from top to bottom]

	Thickness (feet)
Silt loam, grayish-orange (10*YR* 7/4); contains shale fragments	0.7
Silt loam, dark-yellowish-orange (10*YR* 6/6); contains many grayish-orange (10*YR* 7/4) flat shale fragments of pebble size	1.0
Shale fragments, closely packed, and weathered shale bedrock; grayish orange (10*YR* 7/4) on dry surfaces; contains some patches of light-brown (5*YR* 5/6) clay loam	5.0
Total thickness	6.7

Exceptions to these common relations occur wherever the Martinsburg or other shale is covered by a protective mantle of resistant cobbles or blocks derived from outcrops upslope. In such places thick residuum forms on the bedrock surface. Exposures of residuum on the Martinsburg are common, for example, on the slopes of Massanutten Mountain, where sandstone blocks have crept downslope from Silurian sandstone outcrops above. In these places residuum may reach thicknesses of more than 40 feet (fig. 26).

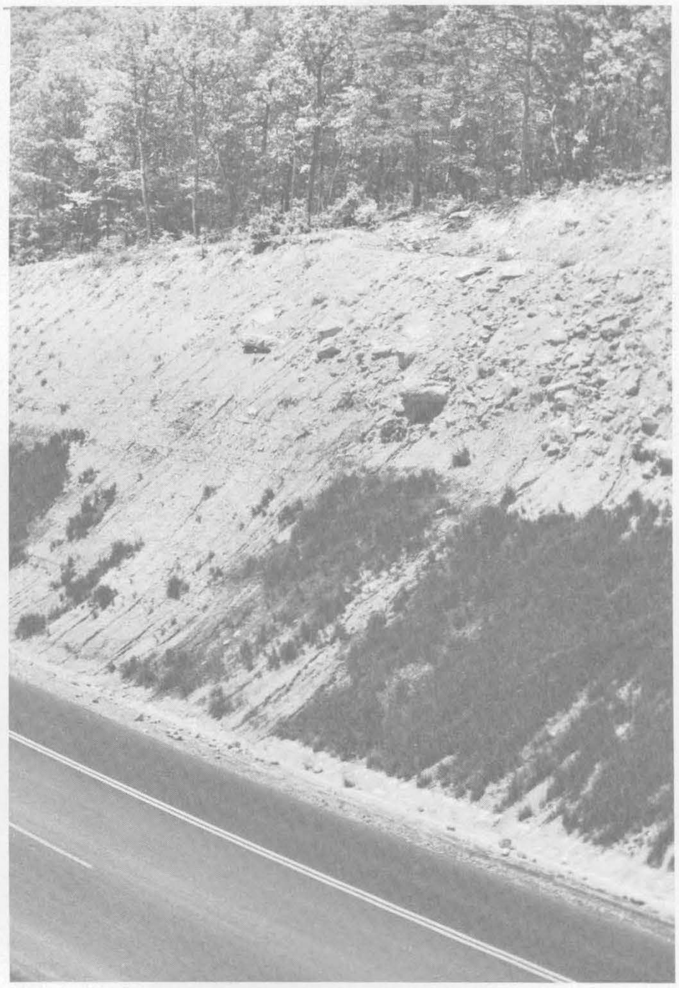

FIGURE 26.—Residual silt and clay derived from Martinsburg Shale under a cover of colluvial cobbles and boulders. Near Luray, Va., on the east slope of Massanutten Mountain.

RESIDUAL MANTLE ON CARBONATE ROCKS

Carbonate rocks underlie a large part of the Shenandoah Valley, forming a lowland whose average local relief is only 200–400 feet. The residual mantle in this area is extremely varied both in thickness and composition. Some large areas have almost no mantle, and bedrock outcrops are more abundant than in even rough mountainous terrain. In other areas the mantle is so thick that no bedrock is exposed even in the deepest cuts. In this report the residual mantle on the carbonate rocks is divided into two kinds on the basis of

FIGURE 27.—Ledges of limestone of Middle Ordovician age cropping out in pasture near Harrisonburg, Va.

thickness. The distinction between the two is readily made in the field.

THIN RESIDUAL MANTLE

The thin residuum is generally less than 5 feet thick and contains limestone fragments or core stones of the underlying rock (pl. 2, map unit c). The color is commonly, but not everywhere, brown rather than red or orange. Roadcuts in the area of thin residuum commonly have bedrock exposed in places to within a few inches of the surface. Rock outcrops are numerous and may be seen projecting above the soil or forming extensive rocky rows, as is shown in the characteristic scene in figure 27. The mantle is absent locally but elsewhere is as much as several feet thick. Because of the stony nature of the soil, the land is generally used as pasture.

The soil matrix is clay, silt, or silt loam. The color is determined in part by the source rock. For example, residuum on the Athens Limestone of Butts (Edinburg Formation of Cooper and Cooper, 1946) is markedly less red than it is on some other rocks. Samples taken from six localities on this rock have a 10*YR* hue and below the A horizon are mostly yellowish orange or yellowish brown. Soils on the Lenoir Limestone of Butts (Lincolnshire Limestone of Cooper and Cooper, 1946) are commonly more red (hue 5*YR* or 10*R*), ranging in the B and C horizons from light brown to very dusky red. In these soils, pH ranges from slightly acid (6.0) to slightly alkaline (7.5). The relatively high pH is probably caused by the abundant core stones of carbonate rock and the closeness of the bedrock surface.

In general, the thin residuum on carbonate rocks is coextensive with certain rock formations or parts of rock formations. It is particularly identified with the rocks of Middle Ordovician age and the upper part of the Beekmantown Dolomite, especially in the northern part of the valley. Patches, however, occur on all the carbonate rock formations, and some beds in the Conococheague and Elbrook Formations generally have thin residuum.

THICK RESIDUAL MANTLE

The areas of thick residual mantle (pl. 2, map unit t) are generally different in character. Outcrops, even in roadcuts, are few and are rarely seen on the slopes. Instead of core stones of limestone and dolomite, the thick residuum contains more siliceous residues derived from rocks like sandstone, chert, and siliceous oolite. The maximum thickness of this unit is not known because deep exposures are not common. Several road cuts more than 40 feet deep, however, do not penetrate bedrock. Like the thin residuum, the thick residuum varies considerably in its properties and reflects in its color and other characteristics the rock on which it occurs. It is probably thickest on the Beekmantown Dolomite, but great thicknesses also occur on the Conococheague Limestone. A thin cherty surface layer is common in mantles on the Beekmantown. The chert consists of large fragments of pebble or even cobble size that are spongelike and light gray to variegated in color. It generally overlies a brown or reddish-brown unctuous clay more intensely colored than the residuum on most other carbonate rocks.

The following are typical exposures:

Exposures of residuum on Beekmantown Dolomite

Harrisonburg bypass east of Harrisonburg

	Thickness (feet)
Silt loam, stony, grayish-orange (10*YR* 7/4); contains many pebbles and granules of light-gray chert; pH 7	1.5
Clay, sandy, light-brown to moderate-brown (5*YR* 4/6), very heavy textured; breaks into pelletlike aggregates on weathered surfaces; pH 5	3.5
Silt or clayey silt; mostly pale-yellowish-orange (10*YR* 8/6) to dark-yellowish-orange (10*YR* 6/6); contains some streaks or masses of light-brown (5*YR* 6/6) silt; pH 5; base concealed	39.0
Total thickness	44.0

Rockingham County, 1.2 miles southeast of Mount Crawford

Silt loam, pale-yellowish-brown (10*YR* 7/2); contains abundant chert fragments	1.3
Silt loam, grayish-orange (10*YR* 7/4); contains a few chert pebbles	1.0
Silt loam, clayey, dark-yellowish-orange (10*YR* 6/6); mottled near base with reddish-brown patches; contains very few chert pebbles	1.0
Clay, hard, blocky, moderate reddish-brown (10*YR* 4/6); contains streaks and mottles of yellowish orange; contains some angular chert pebbles; surface of cut weathers to clay pellets of coarse-sand size (1–2 mm); pH 4.5–5.0; base concealed	2.3
Total thickness	5.6

In the Conococheague Limestone belt, thick residium is less red and is sandy and silty rather than clayey. In places, the uppermost foot or more of the material contains chert fragments, but more commonly the material contains cobble-size fragments of porous sandstone derived from calcareous sandstone beds. Thick sections of this residuum are common in roadcuts. Remnants of structural features such as bedding planes and even joints are characteristic. The following is a typical exposure:

Exposure of residuum on Conococheague Limestone 6.5 miles northwest of Harrisonburg

	Thickness (feet)
Sand, loamy, fine, moderate yellowish-brown (10*YR* 5/4); contains many angular pebbles of chert and sandstone; pH 5	1.0
Sand, loamy, dark-yellowish-orange (10*YR* 6/6); contains pebbles on sandstone and siltstone, which are not as abundant as in layer above; becomes more loamy downward	1.0
Sandstone, weathered, porous, light-brown (5*YR* 5/6); bed inclined at angle from horizontal	.8
Clay loam, light-brown; contains a few pebbles; pH 5–6; base concealed	2.4
Total thickness	5.2

On the Elbrook Dolomite the residuum is commonly darker in color than the residuum of the Conococheague; however it is less red than the residuum on the Beekmantown and appears to be thinner. Areas of thin residuum (pl. 2, unit 3) are scattered in elongate patches. The following is a typical exposure:

Exposure of residuum on Elbrook Dolomite in road cut 11 miles north of Staunton, Va.

	Thickness (feet)
Silt loam, coarse, grayish-orange to dark-yellowish-orange (10*YR* 6/4); pH 6	2.0
Clay loam, light-brown (5*YR* 5/6); contains blocky structure; pH 6	1.6
Clayey loam, mottled light-brown (5*YR* 5/6) and dark-yellowish-orange (10*YR* 6/6); pH 5.5; base concealed	1.3
Total thickness	4.9

Except for a disturbed upper foot or two, structural features in the residuum on limestone commonly reflect structural features in the bedrock from which the residuum is derived. In some exposures, however, there is evidence of minor deformation that occurred during the weathering process. One such locality, illustrated in figure 28, is a roadcut 6–12 feet deep. It extends diagonally up a gentle hill exactly parallel to the regional strike of the Conococheague Limestone. The cut is entirely in dark-yellowish-orange residuum composed mostly of porous silt and clayey silt. On the southeast side of the road, a single bed of porous crumbly sandstone ½–1 foot thick is exposed for a distance of 50 feet in the sloping cut and in the shallow ditch. Exposures were sufficiently good in 1952, when the original sketch was made, so that the sandstone bed could be traced almost continuously and dips measured as shown in the plan view (fig. 28). Observations of many outcrops of bedrock in the Shenandoah Valley show that a sinuous pattern of such a short wavelength

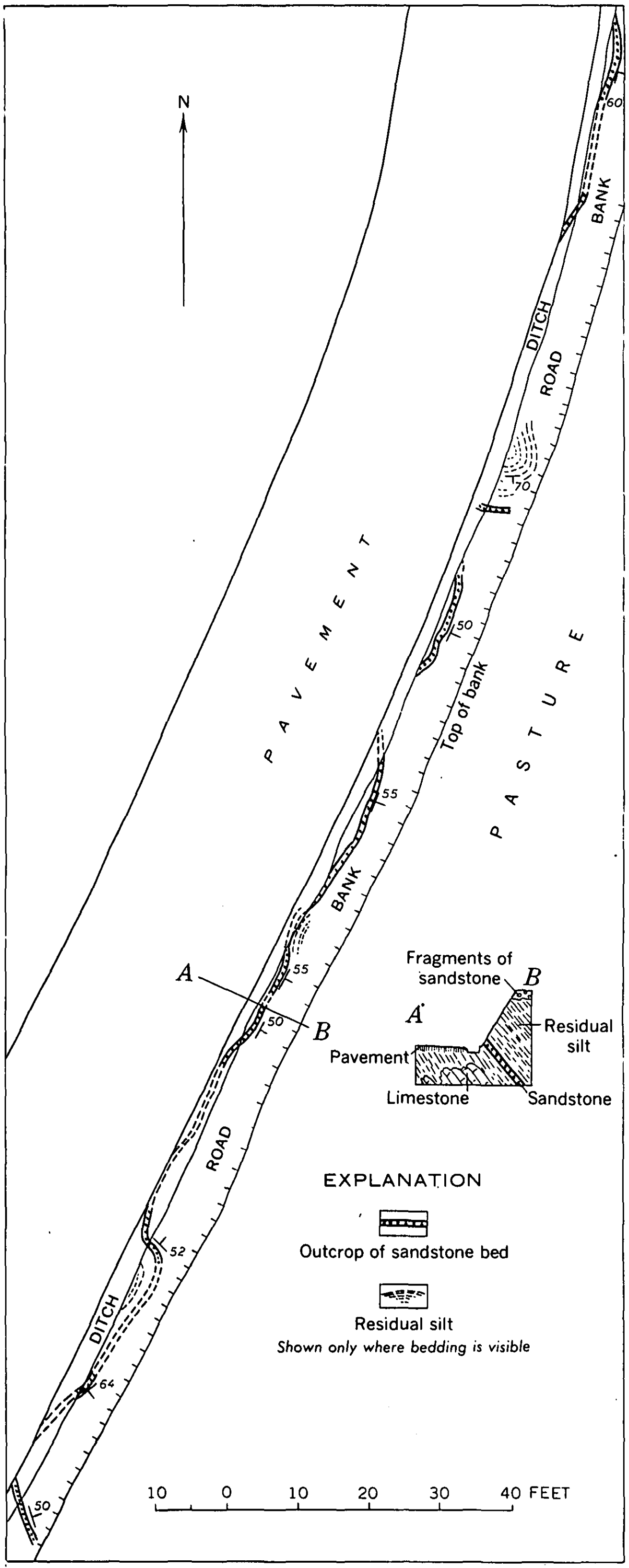

FIGURE 28.—East side of a roadcut, showing a sandstone bed cropping out in silty residuum on Conococheague Limestone. On Old Greenville Road 2.3 miles south of Staunton, Augusta County, Va.

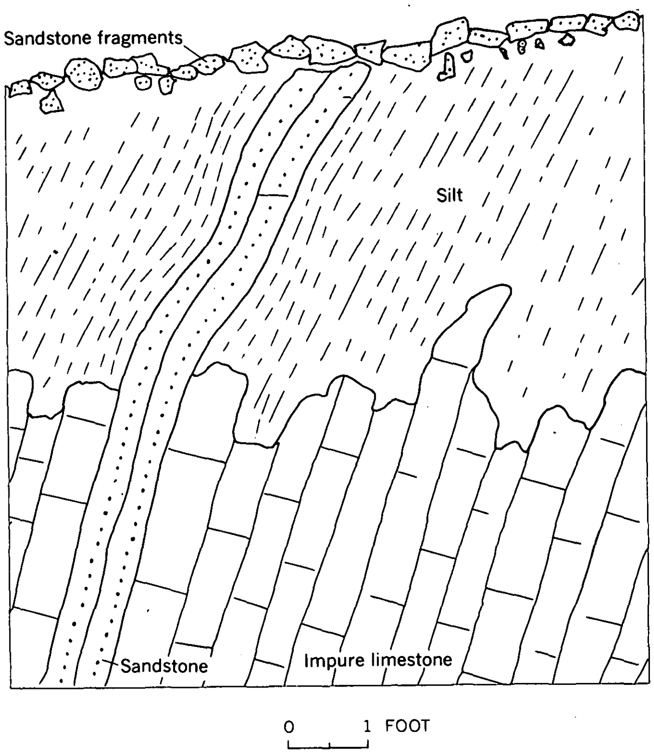

FIGURE 29.—Hypothetical deformation of a sandstone bed in association with loss of volume of surrounding material during weathering.

does not occur in unweathered rocks. Hence, the sinuosity may be presumed to have occurred as a result of differential shrinkage from loss of volume during the weathering of the limestone because of removal of carbonate. In shrinking, a steeply dipping competent bed would slump so as to flatten the dip, as shown in figure 29. If the shrinkage were uneven, then a sinuous or wavy pattern would form.

The wavy pattern cannot be the result of downhill creep, because the gentle 5° slope of the hill is oblique to the strike, whereas the wavy pattern of the sandstone bed is very nearly symmetrical with relation to the strike. Furthermore, it is unlikely that any prolonged creep could occur without destroying the continuity of the sandstone beds.

Areas of thick carbonate residuum are in some places overlain by small patches of alluvium or colluvium apparently transported only a short distance by running water. A few layers of stones a few feet below the surface were observed above which the unconsolidated material appears to be crudely stratified. The stone layers and the material over them resemble the carpedoliths described by Parizek and Woodruff (1957) in the Piedmont of Georgia. These localities are in small hollows several hundred feet from a ridge crest where runoff is concentrated on the slope but where no channel or gully has formed. Exposures in such places are obviously uncommon, for hollows are generally crossed by roads on low fills rather than in cuts. It is not known, therefore, whether the small alluvial deposits are generally associated with hollows. The writer's opinion is that the stone layers represent floors of former gullies or ravines that have been filled by crudely stratified colluvium. The pavement of stones beneath the colluvium is a lag deposit. The gullies in residuum formed at some time in the past and were possibly caused by cloudbursts that produced excessive runoff. Another cause might be a prolonged period of drought that removed the vegetative cover.

Thick residuum underlies alluvial terrace deposits in many places and is therefore included in the area mapped as alluvium (pl. 2, unit at). Beneath the extensive alluvial aprons at the west foot of the Blue Ridge (pl. 2), a residual mantle in places more than 200 feet thick (King, 1950, p. 54) completely blankets lower Cambrian carbonate rocks. This residuum is exposed in manganese and iron mines and prospect pits. In many places it contains relicts of the original bedding, but in other places slump structures indicative of deformation are found. Such structures on a large scale may be related to solution and caving of carbonate rocks beneath the residuum. Sinks of various sizes are common on the surface of gravel terraces in many parts of the valley, an occurrence indicating solution of underlying rocks. The residuum beneath the alluvial cover ranges in texture from waxy clay through silt to coarse sand and commonly contains chert. In the area west of the Blue Ridge, the rocks from which the residuum is derived commonly cannot be identified because exposures are scarce. The Tomstown, Waynesboro and Elbrook Formations are all represented, and the rocks of all these formations contain enough impurities to provide a considerable amount of residuum.

Residuum not only underlies many terrace areas but crops out in the flood plains of streams (King, 1950, pl. 1). South of Waynesboro, mining and drilling at the Lyndhurst mine on Back Creek revealed that residual clay extended to a depth of more than 100 feet beneath the level of the stream (Knechtel, 1943, p. 181–184).

RELATION OF RESIDUAL MANTLE TO TOPOGRAPHY AND KIND OF BEDROCK

The distribution of the residual mantle in the Shenandoah Valley is closely related to the kind of bedrock. The mantle is not related to any particular altitude or level, such as would be expected if it were formed on a peneplain or ancient surface, and its pattern of distribution does not suggest that it is a remnant of a once continuous deposit. Thus, in the limestone lowland the residual mantle is particularly thick and extensive in the southwestern part of the valley, where relief on the carbonate rocks is the greatest. The mantle is least widespread in the northeastern part of the valley, where the relief is more moderate and where the mantle should be found if it had formed on a surface of lower relief. The residuum may occur at any level and even extends down to 100 feet below the flood plains of some streams like Back Creek, a large tributary of the Middle River.

Comparison of plates 1 and 2 shows that areas of thick residuum are arranged in belts parallel to the strike. Residuum does not occur extensively on the rocks of Middle Ordovician age, and the trace of these rocks can be followed over the entire Shenandoah Valley, except in the alluvial areas, by the absence of an appreciable mantle over them. The mantle on these rocks is sparse on the drainage divides as well as in lower areas near the streams. However, a more detailed examination of the area discloses the following: Whereas the Athens and Chambersburg Formations, which constitute the greatest thickness of the Middle Ordovician sequence, almost universally lack any residual mantle except where covered by alluvium, the Lenoir Limestone, also a part of this sequence, does have a mantle of residuum, which, however, is too narrow to be shown on the map (pl. 2). Thick residuum is generally lacking on the upper beds of the Beekmantown Dolomite, but it is widespread on the lower beds of this formation as well as on all the Cambrian carbonate rocks. Residuum is thin or lacking on the Martinsburg Shale, except where the shale is covered by alluvium.

The generalization can be made that all the carbonate rocks mantled with thick residuum contain coarse siliceous impurities that are inert chemically. The residue from the Lenoir Limestone is a blocky black chert that occurs as thin bands in the rock. The Beekmantown Dolomite contains thick beds of massive gray chert at various horizons, especially in the lower part. The Conococheague Limestone of Cambrian age contains sandstone interbedded with the limestone and dolomite. The coarse inert constituents survive the weathering process; as downwasting of the hills occurs, they are concentrated on the ground surface and form an armor or blanket that protects the fine constituents beneath from erosion. Some of the hillslopes in these areas are almost paved with small chert fragments. In other places, especially on the Conococheague, the upper few inches of the mantle is a sand containing only scattered blocks of weathered sandstone.

The amount of inert residue in the rock is not the factor that controls the accumulation of residuum. The most important factors are the size, toughness, and rate at which the fragments of residue can be broken up.

The Athens Limestone, for example, the most conspicuous of the formations that lack residuum, is an impure limestone containing beds of shale in addition to clayey impurities in the limestone itself. The Martinsburg Shale also generally lacks residuum, even though most of the material in it is siltstone and claystone. On the other hand, Martinsburg Shale does break down into thick clay and silt where protected by a mantle of sandstone blocks, as downslope from an outcrop of sandstone or under an alluvial terrace deposit.

All the rocks are subject to both mechanical and chemical weathering, but the rates of the two kinds of weathering differ. In mountain areas like the Blue Ridge, quartzites are subject to chemical weathering, but the rate is slow and the rocks stand in rugged relief. As a result they are exposed to mechanical weathering, and much of the material in the mountains is carried off by creep and by streams in the form of coarse blocks. In the lowland areas, mainly because of the kind of rock present, chemical weathering is more rapid than mechanical weathering and produces fine-grained residues that can be carried off on gentle slopes. Small amounts of resistant material in the rock, like chert or sandstone, weather more slowly and, like the quartzite in the mountains, form blocks that are mixed with or blanket the fine-grained material.

ALLUVIUM

Alluvium in the Shenandoah Valley is widely distributed in piedmont aprons at the foot of the higher mountains and as flood plains and terraces in narrow belts along the streams. On the map (pl. 2) a distinction is made between alluvial bottom lands (unit ab) and terraces (unit at). This distinction is rather arbitrary and is based on the height above stream grade and on the degree of dissection of the surface. The two units have little stratigraphic significance and serve only to separate the lower terraces and bottom lands closely related to the present streams from higher ones that are more dissected and that may in some places have been deposited by different streams. The lower of the two units corresponds to King's (1950) lowest unit in the Elkton area, which he calls alluvium. The higher corresponds to all three of King's older units, which he refers to as gravel units.

Nearly all the alluvial deposits are graded so that the coarse material is at the bottom and the fine material at the top. The size difference may be considerable; for example, along the Middle River the deposits range from cobbles or boulders to silt. The graded character is typical of nearly all stream deposits. As interpreted and described by Mackin (1948, p. 472), Hack (1955, p. 34), and Wolman and Leopold (1957), the coarser material is deposited in the bed or channel as the stream shifts laterally. The finer material represents point-bar accretions and overbank-flood deposits. The total thickness of a typical graded alluvial sequence is comparable to the maximum depth of the stream that formed the deposit when in flood. If aggradation has occurred, however, the thickness of an alluvial deposit may be much greater. In the extensive alluvial aprons that occur between Massanutten Mountain and the Blue Ridge, thicknesses greater than 140 feet are common, as indicated in test holes, and at one place 260 feet of gravel was penetrated (King, 1950, p. 59). Generally, the alluvial deposits are much thinner, and a thickness of 25 feet appears to the writer to be the average maximum.

The coarse fractions in most places are gravels composed of quartzite and sandstone. They are not visible in the flood plains of modern streams, except at low water in some cuts in the stream banks. The stream beds are composed of the coarse fractions, however, and the coarse material is evident in exposures of terraces, especially in the highest ones where the fines have been eroded and the ground is littered with cobbles and boulders. In some areas where the most resistant source rocks are not sandstones, the alluvium has a different character, and alluvial deposits at the foot of ridges of cherty residuum are primarily composed of chert fragments and clay. Such deposits are generally not extensive enough to map, but at localities 4.8 miles west of Edinburg and 12 miles south of Harrisonburg, cherty alluvium overlaps relatively nonresistant rocks below unusually large outcrop areas of cherty beds.

The size of the bed material in many streams has been measured (Hack, 1957; Hack and Young, 1959). The average material in most streams is of cobble size (64–256 mm), but scattered boulders occur. Some streams in the limestone region, especially those of smaller size, are floored by pebble gravel rather than cobble gravel. The coarse parts of the terrace deposits seem to be similar in size to the bed material in the corresponding streams, and deposits of cobble gravel are abundant.

At the foot of the Blue Ridge, especially between Elkton and Front Royal, the alluvium contains metabasalt and granodiorite in addition to quartzite. These deposits show evidence of postdepositional differential weathering—that is, the quartzites and sandstones are more resistant than are the igneous rocks. Table 5 compares the modern alluvium with gravel in two terraces on Pass Run near Luray, a stream that drains a large area of granodiorite. Note that the proportion of quartzite pebbles in the terraces is considerably greater than in the modern alluvium. Since the matrix of the

FIGURE 30.—A small stream valley north of Harrisonburg, Va., that is typical of valleys that drain limestone areas and have no terraces.

terrace gravel is also much more clayey than is that of the flood-plain gravel and is darker in color, the difference between the two has probably resulted from decomposition of some of the igneous cobbles in the terrace gravel to form clay and sand, thus increasing the relative proportion of quartzite. Another high terrace near the South Fork, Shenandoah River, 200 feet above river level, is composed of quartzite boulders in a matrix of reddish-brown clay; this composition suggests that igneous boulders, once present, have been completely disintegrated by weathering.

TABLE 5.—*Cobbles in Pass Run, near Luray, Va.*

Location of sample	Number of pebbles larger than 2 mm in diameter	Proportion of quartzite pebbles (percent)
Bar in river bed	95	12
Terrace:		
20 feet above stream	71	17
40 feet above stream	76	24

The alluvium is commonly arranged in flights of terraces along the streams or in alluvial aprons at the mountain foot. However, terraces do not occur along all the streams and are restricted to the valleys where the coarse fractions of the alluvium are more resistant than the bedrock. For example, terraces do not occur along most small streams of the limestone lowland. Instead, such valleys generally have a single wide flood plain (fig. 30). The alluvium consists of cobbly fragments of carbonate rock overlain by silt and clay. Terraces are also lacking in many of the mountain areas where the valley walls and floor are composed of sandstone that is as hard and resistant as the cobbly alluvium of the stream channel. Thus terraces of the usual kind are virtually absent in areas of the Antietam Quartzite or along streams like the Little River in the Hampshire Formation, where the terraces that do occur are very low and are formed as a result of channel changes during floods (Hack and Goodlett, 1960, p. 48–55).

Terraces are most extensive wherever a stream carries hard resistant alluvium through a relatively soft rock area. Thus all the large streams entering the carbonate-rock area from the highlands underlain by sandstone or greenstone are bordered by terraces. The larger the drainage area in the resistant rocks, the larger the terrace areas downstream in the softer rocks. Christians Creek for example, a large stream whose drainage basin is almost wholly in the soft rocks, has a flight of rather narrow terraces, which are not as well formed as the terraces along streams originating in sandstone areas.

Christians Creek carries chert in its bed rather than sandstone, and the lower layers of the alluvium are composed of small chert fragments. Terraces are generally better formed where streams cross the Martinsburg Shale than they are in carbonate rocks. Thus the North River is bordered by broad terraces in the short reach west of Grottoes but by narrower terraces both upstream and downstream from the shale belt.

The terrace deposits show evidence of alteration related to age or drainage conditions and change progressively with height above the stream. The bottom-land deposits, which are occasionally flooded, approach closely the color of the A horizon of the soils and are less red than the terrace deposits. Pale yellowish brown (10*YR* 6/2) and dark yellowish brown (10*YR* 4/2) are common colors. In the carbonate rock areas, the pH ranges from slightly acid (6.5) to alkaline (7.5–8.0), probably because of occasional flooding by waters having a high lime content (Carroll, 1959).

Better drained terrace deposits above the valley bottom lands are almost invariably oxidized and are yellowish orange or reddish brown. B and C horizons of terrace soils, for example, are commonly moderate reddish brown (10*YR* 4/6) or dark reddish brown (10*R* 4/3). The pH ranges from less than 7 to as low as 5.0. The flood-plain deposits have relief features of depositional origin, such as flood channels and bars. Higher terraces are progressively more dissected, and the highest are in places evident only because of cobbles in the soil. The following is a section of a well-preserved but dissected terrace remnant typical of many of the exposures of older terraces:

Section of a weathered terrace deposit 75 feet above river level in a roadcut in Middle River valley, 6 miles north of Staunton, Va.

	Thickness (feet)
Silt, yellowish-orange	0.6
Clayey loam, moderate reddish-brown (10*YR* 4/4)	.3
Sandy loam, fine, pale-brown, very hard; broken by polygonal cracks 1–2 ft apart filled with clay	4.0
Loam and gravel containing pebbles of sandstone, pale-brown; base concealed	3.0
Total thickness	7.9

The Middle River valley serves as a sample area illustrating the disposition of alluvial terraces along medium-sized streams that cross a variety of rock types. The terraces were traced downstream from East Dry Branch, one of the principal tributaries in the mountains, to the junction of the North and Middle Rivers. In this part of the course, the Middle River and its tributary, East Dry Branch, are bordered by a well-formed, though narrow, flight of terraces. The lower ones are close to the river and can be called bottom-land surfaces. Thirty-nine cross profiles of the valley floor, averaging 1.6 miles apart, were measured with a hand level. They include 13 cross profiles along East Dry Branch in Devonian sandstone and shale, 19 along the Middle River in carbonate rocks, and 7 in the lower course of the river in the Martinsburg Shale. Representative examples of these cross profiles are shown in figure 31, and the terrace heights above stream grade are plotted on a graph in figure 32. The different spacing of the terraces in the different physiographic subdivisions of the valley are quite apparent in this graph.

Terrace remnants are fairly abundant up to about 75 feet above the stream. Still higher remnants occur at various altitudes up to 200 feet above the river, and at a few places sandstone cobbles are found in the soil of the upland above the river. Only the lowest terraces below about 30 feet above the river are sufficiently continuous to be traced for any great distance. The higher remnants are small and at such diverse elevations that they cannot be correlated except on some arbitrary basis. Soil profiles are too much alike to provide a means of correlation, although the degree of weathering of the boulders as well as the amount of dissection increases in the higher terrace remnants.

Below 30 feet however, three terrace surfaces (T_0, T_1, and T_2, figs. 31 and 32) are traceable for considerable distances along various reaches, though in places one or even two are missing. They all appear to be subject to at least occasional flooding. Records at the stream gage at Mount Meridian, near Grottoes, Va., indicate that the lowest surface, T_0, is flooded more than once a year on the average. The intermediate surface, T_1, is flooded every 2.3 years. The highest surface, T_2, is not exposed at Mount Meridian, but the statements of property owners at other localities along the river indicate that terrace T_2 was overtopped at many places during a flood that occurred in 1936, the highest flood on record at Mount Meridian. During this flood the gage height was 22 feet, or 18 feet above the low water level shown in figure 32.

The graph (fig. 32) indicates that none of the terraces is continuous along the entire river and that their heights are not constant. The writer made a careful attempt in the field to trace the terrace surfaces downstream and to identify them in each section but recognized that the only real basis for correlation is the three-fold sequence. It is doubtful whether a certain correlation for the entire river could be made even by continuous mapping. To correlate higher terraces such as T_3 would be quite impracticable.

Nevertheless, as a generalization it may be stated that the Middle River is bordered by three terraces or

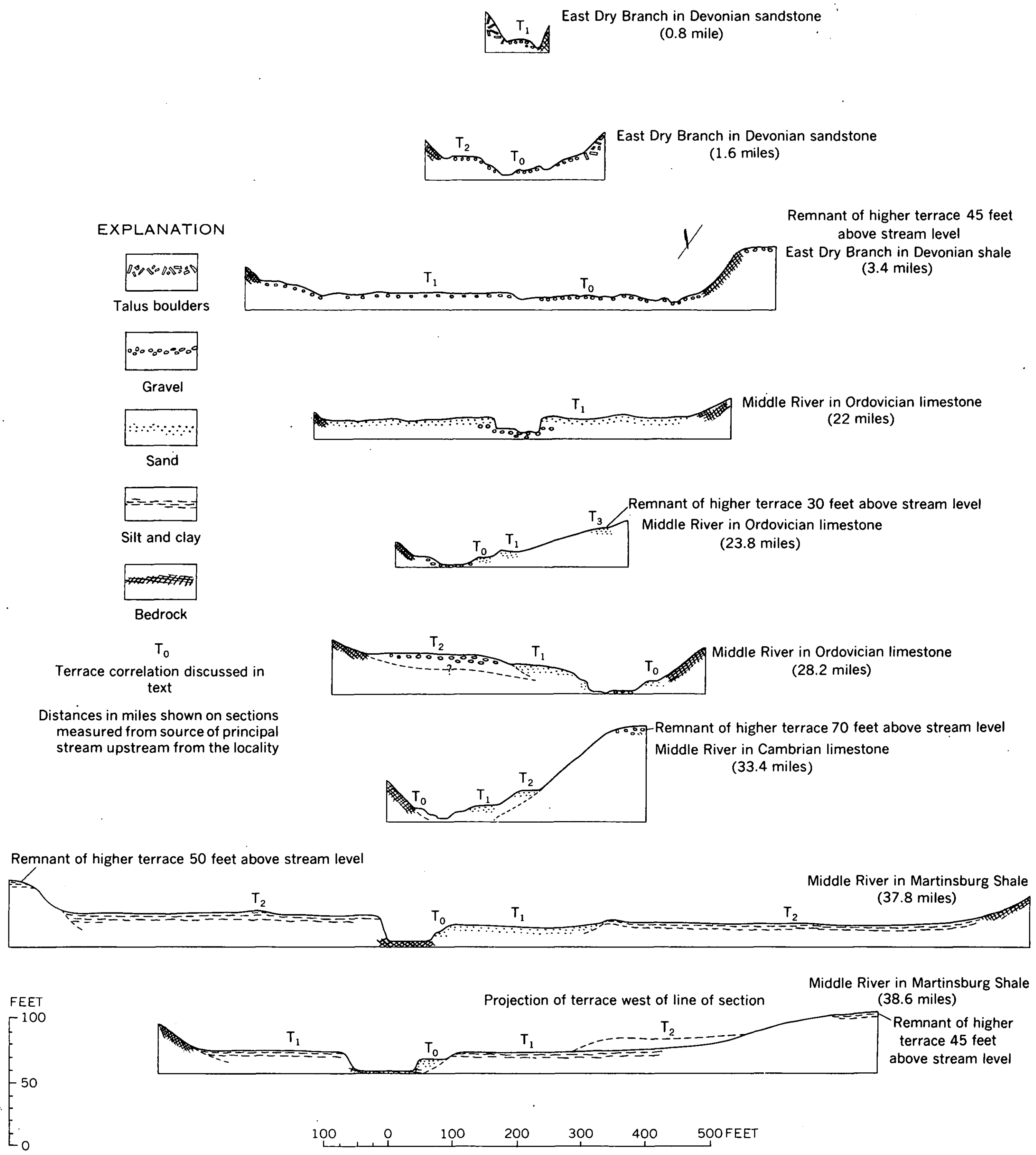

FIGURE 31.—Representative profiles of the valley floor of East Dry Branch and the Middle River.

bottom-land surfaces. The lowermost surface, flooded frequently, is underlain by sand and gravel in East Dry Branch and by sand downstream. Generally, it is narrow and almost a part of the channel, but in places it widens to 500 feet. Above this are two terraces that are generally much more extensive and underlain by finer grained material. In East Dry Branch their surface material is sand, whereas downstream it is gen-

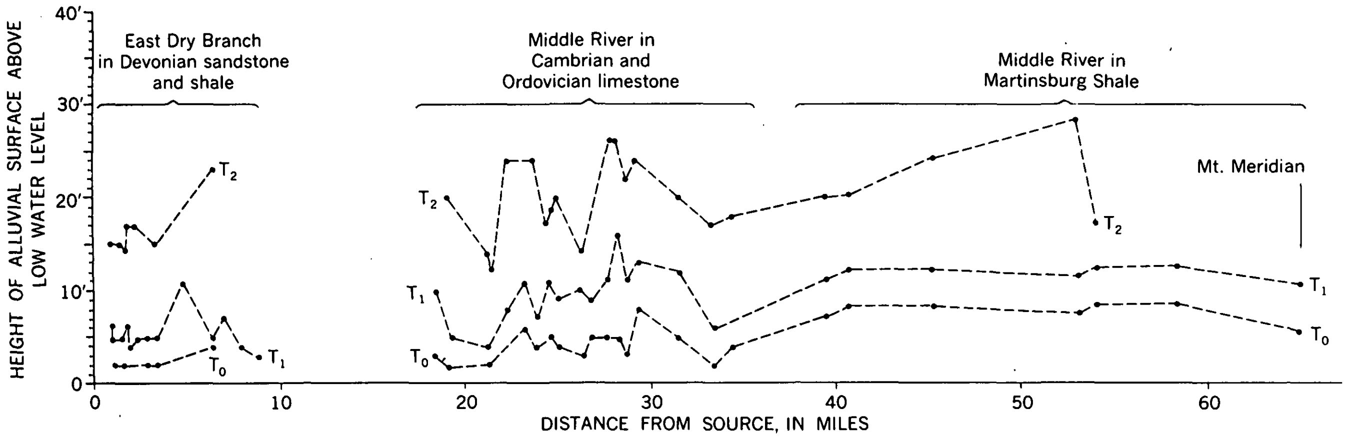

FIGURE 32.—Height above stream grade of the lower terraces or bottom lands along the Middle River valley at low water level.

erally silty loam. The material of the higher surface is commonly somewhat finer than that of the intermediate one.

INTERPRETATION OF THE REGOLITH

The surficial deposits of the Shenandoah Valley were mapped and studied for the first time by King (1949, 1950, p. 58–62) in the Elkton area. He subdivided the extensive gravels of that area into three units: the younger, intermediate, and older gravel units. In his 1950 report he regarded them as Pleistocene in age and thought they were deposited during the cold periods, when the upper parts of the Blue Ridge Mountains were probably above timberline and scree formation was active. The present writer believes that although the rate of formation of alluvial gravel may have varied considerably in the past and was affected by climatic changes, the Pleistocene cold was not the primary cause of gravel deposition. The deposition of gravel in the Shenandoah Valley is a continuous process and is related to contrasts in resistance of the rocks within a drainage basin. The distribution of the alluvial deposits, if considered throughout a large area, supports this interpretation. The reconnaissance map (pl. 2) brings out the close relation between the alluvial deposits and the resistant rocks that are their source. The principal rocks that supply gravelly alluvium are the granodiorite, Catoctin Greenstone, quartzites of the Chilhowee Group, Tuscarora and Massanutten Sandstones, and Pocono Sandstone. In addition, the sandstones of Devonian age are an important source of gravel even though they are mostly interbedded with shale.

Wherever the outcrop areas of these rocks is large and drained by large streams, fanlike aprons of alluvium are spread across the softer rocks that outcrop downstream. These deposits may be called piedmont alluvial aprons. They consist of typical river alluvium, such as cobble and boulder gravel, overlain by sandy loam and silty loam. The average size of the material decreases away from the mountain (Hack, 1957, p. 84). The alluvium is arranged in terraces, the higher of which are dissected and eroded so that only the gravelly part remains. The gravel is commonly underlain by residuum. As shown on plate 2, the size of a complex of alluvial aprons is roughly proportional to the size of the drainage basin in the mountains.

The principles that control the formation of the piedmont alluvial aprons have been discussed elsewhere (Hack, 1957, 1960a, p. 91–94) and are briefly summarized as follows: The average channel slope of a stream is inversely proportional to its discharge but directly proportional to a function of the size of the material it transports and the resistance of the bedrock or material that encloses the channel. Thus the piedmont streams draining areas of sandstone and quartzite have steeper longitudinal profiles than adjacent lowland streams despite their larger drainage area and discharge. The differences in the profiles are so great that streams like the North River, which are of mountain origin, cross the piedmont area at higher levels than their own downstream tributaries. The drainage system as a whole is in equilibrium, and the piedmont is transitional between the hard-rock area of high relief and the soft-rock area of low relief. The alluvial aprons exist because the debris shed by the high-relief area cannot be immediately carried off on the gradient of the lowland streams, and the debris is stored in the piedmont region until it is reduced by weathering and erosion to finer sizes. The area of the aprons is thus proportional to the area of the drainage system in the mountains and also to the difference in rock resistance and relief between the mountain and lowland areas.

A piedmont alluvial apron is really a kind of pediment, for it consists of a broad plain cut on bedrock and covered with a thin sheet of gravel arranged with other sheets of various ages in a fanlike complex of dissected and undissected plains at the mountain foot. As in typical pediments the gravel of the alluvial aprons rests on the eroded surface of the bedrock. In terms of the equilibrium theory, the origin is similar, though not identical with, the origin of pediments in the Henry Mountains as conceived by Hunt, Averitt, and Miller (1953, p. 189). The area as a whole is downwasting, and the net result of the processes in the piedmont area is to degrade the area. Alluviation at any one place is only temporary. The fragments shed by the mountain stream are deposited at the mountain foot in a fanlike thin deposit that tends to grow laterally because the mountain cobbles are more resistant to corrasion and corrosion in the stream than are the softer rocks of the valley. Thus the stream has a tendency to erode and migrate toward the valley side and to widen its valley. As the valley side is eroded, cobbles and gravel are deposited on the flood plain opposite to it. During the widening process low terraces may be formed as the channel shifts its position. The gradient of the alluvial surface is adjusted to the discharge and load of the mountain stream. Because of the size of the load, this gradient is generally steeper than the gradient of a smaller adjacent lowland stream, which erodes a narrower valley at the faster rate. Eventually a piracy occurs, and the lowland stream receives the load of coarse debris formerly deposited on the adjacent apron. The lowland valley is now alluviated but of course is never filled to the same level as the abandoned alluvial valley. As the process of degradation by the valley streams, piracy, and alluviation continues, the piedmont apron is gradually spread and becomes a complex of gravelly flood plains, terraces, and dissected terraces.

In the carbonate rock area the flood plain of a piedmont stream is an ideal site for the formation of underground caverns and the entrapment of residuum beneath the gravel. Since the waters of the mountain stream are low in alkalinity, they are capable of dissolving large amounts of carbonate rock. The overlying flood-plain gravel prevents the escape of the silt and the clay. The hydraulic gradient beneath such a deposit is high compared with other areas because nearby lowland streams are flowing at a lower level and ground water may flow laterally from the higher stream to the lower one.

The higher parts of the piedmont aprons that have been abandoned are eventually weathered and dissected until only gravel-capped hills are left. These disappear as erosion continues, and the former high area becomes a hilly lowland again and is at length reduced to altitudes below the more recent and originally lower parts of the alluvial aprons. The area of the alluvial complex is thus an exception to the general rule that all parts of the area are drownwasting at the same rate. The areas whose streams are entirely of lowland origin are downwasting more rapidly than the flood plains of the streams of mountain origin. Equilibrium is maintained only through the occurrence of piracies and shifts of the drainage from one piedmont valley to another.

Many areas in the Shenandoah Valley exemplify the effects of the processes just described. Probably, the most spectacular is in the headwaters of the North River west of Harrisonburg, briefly described in an earlier paper (Hack, 1960a). Part of this area is shown in figures 33 and 34, a topographic map and cross section which include the extensive piedmont aprons formed along the North River and Briery Branch. Dissected terraces at several levels are also shown. For example the upland plain north of Sangerville was once the flood plain of Briery Branch when it joined the North River near Sangerville. The dissected area south of Spring Creek and west of the village of Spring Creek is a still older terrace remnant and is also a former alluvial plain formed by Briery Branch. Exposures in the highway cuts south of Spring Creek village show that the gravel on this terrace is underlain by residuum. Mossy Creek in the southeastern part of the area (fig. 33) is an example of a stream in the carbonate rocks that has a gentler gradient than the North River, even though the North River, its master stream, has a much larger drainage area. Note that the road junction in Mount Solon has an altitude of 1,322 feet, about 50 feet lower than the flood plain of the North River less than half a mile away (fig. 33). During the extraordinary flood of June 17–18, 1949 (Mussey, 1950; Hack and Goodlett, 1960, p. 42), water actually spilled over the divide from the North River into Mossy Creek, flooding through the town of Mount Solon. During the same flood, large areas of sandstone cobbles and boulders were spread by the floodwaters on the low plain adjacent to the river.

The pond at Mount Solon is a large limestone spring that probably receives underflow derived from the North River. Considerable underground solution of the carbonate rocks has evidently gone on in the past between the North River and Mount Solon, as is indicated by a row of sinkholes on the hill north of town. The largest sinkhole is about 100 feet deep; its bottom is below the level of the North River flood plain. Underground solution is also evident on some of the terraces. The low terrace north and east of BM 1543, for example, shown in the southwest corner of figure 33, although covered with gravel, is pockmarked with undrained depressions. The extensive terrace plain be-

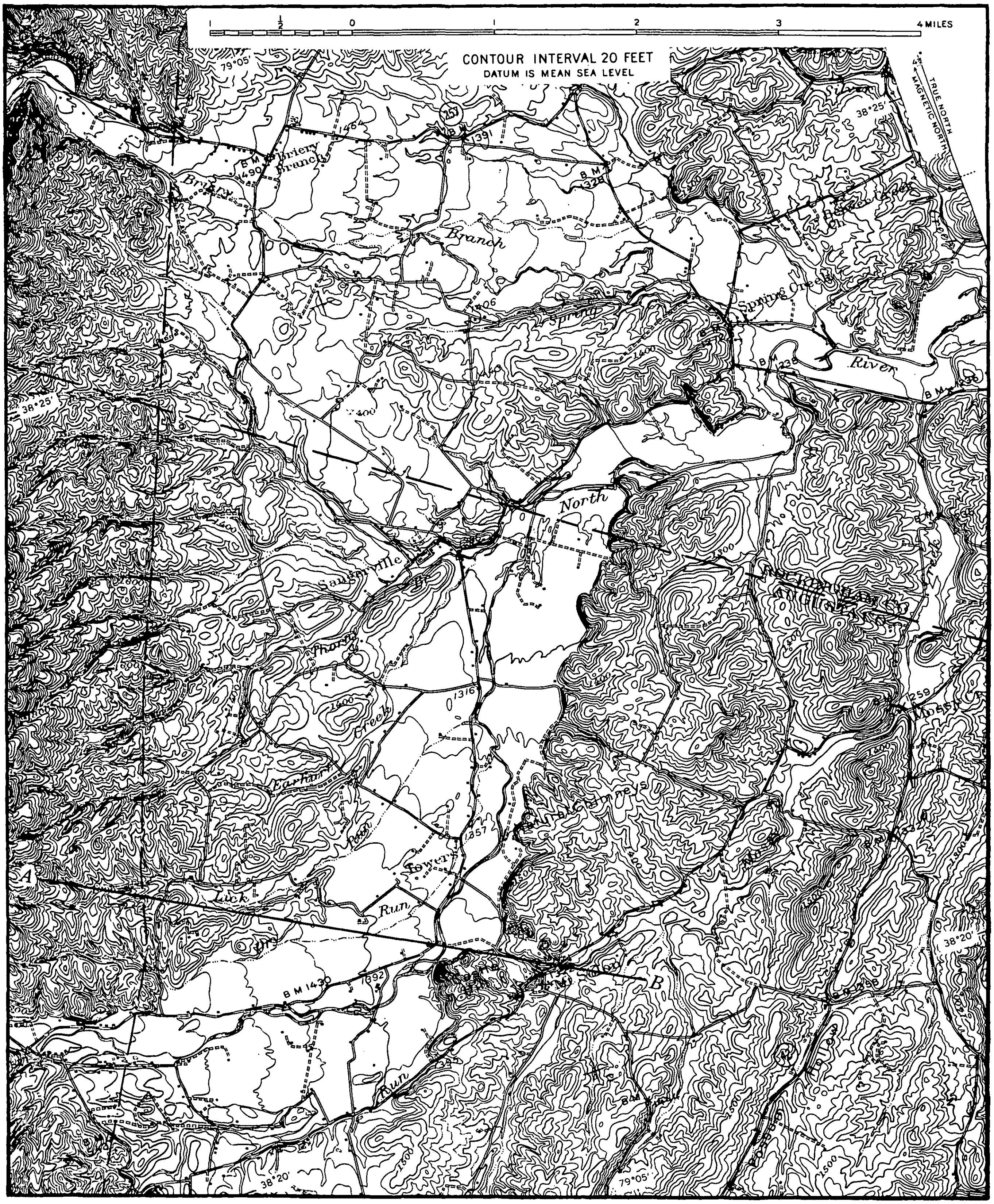

FIGURE 33.—Valleys of North River and Briery Branch at the point of entry into the carbonate rock lowland west of Harrisonburg, Va. Line *A–B* shows location of the geologic section described in text (fig. 34). Photographed from the Parnassus 15-minute quadrangle, Virginia and West Virginia: Topographic Division, U.S. Geol. Survey, 1944.

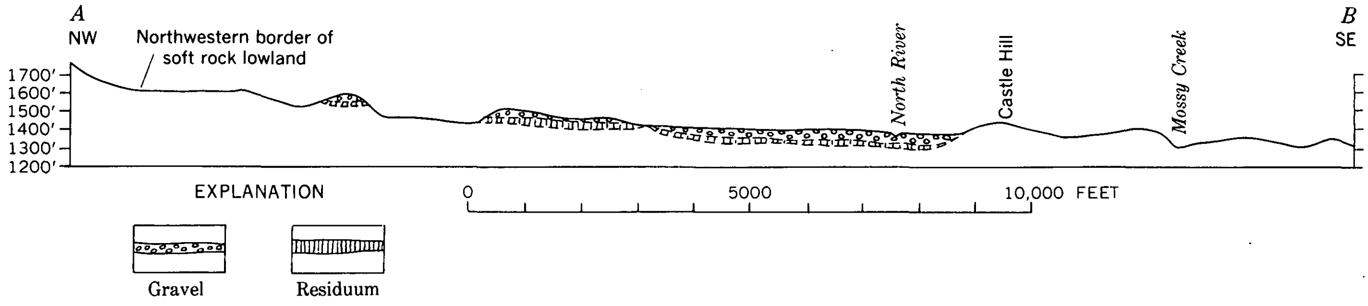

FIGURE 34.—Section (*A–B*, fig. 33) through piedmont alluvial aprons of the North River. The presence of the residium beneath the gravel is inferred.

tween Sangerville and the town of Briery Branch also contains many undrained depressions.

Many places in the Shenandoah Valley provide examples of terraces spread and dissected as the result of piracy. Some of them may be seen on the small-scale map (pl. 2). Several examples of abandoned alluvial aprons, for instance, are in the headwater area of the Middle River basin. Stony Creek, west of Edinburg, is a stream that is deeply intrenched beneath an older alluvial plain and along which a broad flood plain has not formed.

On the southeast side of the Shenandoah Valley, where gravel aprons are almost continuous, examples of piracies are less evident because the individual aprons are not so clearly marked. Many streams that drain the surface of the piedmont alluvial plain, however, are intrenched into it and flow at levels lower than larger streams of mountain origin. The streams of mountain origin because of their load of fresh quartzite cobbles do not develop gradients as low as the streams that originate in the piedmont plain itself, even though most of the area is gravel covered.

Alluvium also underlies terraces that border stream valleys like those of the Middle River, North Fork, Shenandoah River, Cedar Creek, and many other lowland streams. These terraces are also dependent on a contrast in resistance between the alluvium and the rock that encloses the valley. Flood plains occur on nearly all the streams regardless of the basin geology, but terraces occur only where the valley walls are of relatively soft rock, like limestone or shale, and the alluvium is material like sandstone or chert (p. 50). This is perhaps because terraces are flood plains that have been abandoned for one reason or another. If the material of the flood plain is less resistant than the valley walls, it is unlikely to survive as a terrace as downwasting proceeds. On the other hand, if it is very much more resistant, patches of it may remain as remnants on the valley walls and survive through a long period of downwasting until it even reaches a position on the drainage divide.

The distribution of the residuum in the lowland areas is explained in a manner similar to the distribution of the alluvium. Nearly all the limestone in the Shenandoah Valley contains some fine-grained impurities. As the rocks erode, the fine-grained insoluble residues are washed off the ground surface. Tough inert materials like chert cannot be broken down into small size at the same rate. They accumulate on the surface (fig. 35) and form a layer that protects the finer residue, also a constituent of the bedrock, from immediate erosion.

Since all elements of the landscape are downwasting at the same rate, the chert fragments above the cherty limestone must be eroding as rapidly as the ground surface on the nearby chert-free hills. The slopes on the cherty hill, therefore, must be steeper so that chert fragments can move downhill by creep and wash. As this erosion occurs, some of the fine-grained material escapes and is washed away. The materials are constantly replenished from below. It is the chert and the rate at which it breaks up, weathers, or moves downslope by creep and wash that determines the steepness of the hill. The mechanism explains why cherty carbonate rocks are commonly mantled by dozens of feet of highly oxidized silt and clay containing relatively little chert, whereas the surface layer, less than a foot or two thick, is commonly composed entirely of chert fragments.

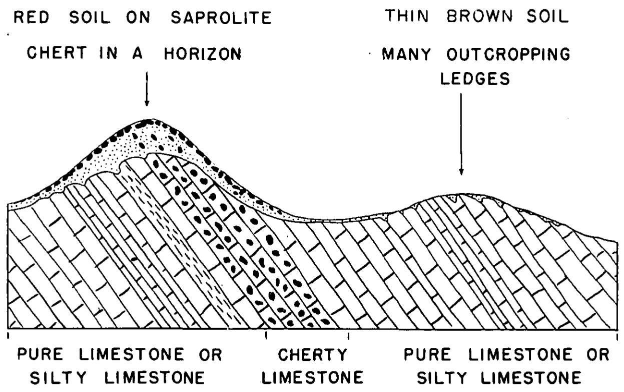

FIGURE 35.—Hilly area in limestone country showing the formation of residuum under a cover of coarse fragments of chert.

Thus, the cherty limestones and dolomites, especially the Beekmantown, form conspicuous ridges and ranges of hills that may be several hundreds of feet higher than their surroundings (fig. 11). The mechanism also explains why the residual mantle is more widespread in the southwestern part of the Shenandoah Valley than in the northeastern part. The difference is related to the character of the bedrock, particularly the Beekmantown Dolomite, which appears to be more cherty in the southwest (p. 13).

The equilibrium concept also explains why thick residuum collects beneath a cover of alluvium and is found even at substantial depths below the modern streams at the foot of the Blue Ridge. Coarse alluvium, like the fragments of chert, acts as a trap preventing the escape for a time of the fine residues. Residuum may form beneath some streams because the waters draining the Blue Ridge slopes and valleys, particularly within the outcrops area of the Chilhowee Group, are acid and have a suitable ionic state to dissolve limestone and dolomite. Because the water percolates copiously through the coarse silicious alluvium, it is an ideal agent for limestone solution (p. 72).

The oxidized state and reddish color of the residuum, especially where it is thick, suggests that the material has been through a period of time during which iron sequioxides were formed and concentrated by lateritic weathering. Denny (1959, p. 10) described red lateritic paleosols he believed to be pre-Wisconsin in age in Potter County, Pa., and Nikiforoff (1955, p. 53) suggested that red mottled subsoils of the Maryland Coastal Plain formed during a warm period in late Pleistocene time. This hypothesis may be applied to the red subsoils of the Shenandoah Valley and is not out of harmony with the equilibrium concept. It simply means that the rate of downwasting in the area is slow enough so that bodies of subsoil whose characteristics were partly determined by warm climates in pre-Wisconsin time have survived through the Wisconsin to the present. If, as is estimated on page 63, the rate of downwasting of the area is only about 1.3 feet in 10,000 years, then lateritic subsoils in areas of thick residuum could have survived from a time long before the Wisconsin, even if downwasting were continuous and the rate of erosion were greater than at present.

CONTINUITY OF TERRACES

In accord with the equilibrium concept of landscape, the alluvial areas bordering streams should be arranged in flights of terraces that diminish in size and increase in degree of dissection in proportion to their altitude above the present streams. These should be no particular grouping, in their arrangement if they are presumed to form as result of a continuous process. Thus a piedmont alluvial apron will become a terrace whenever the flood plain of a stream of mountain origin is abandoned because of a piracy. A terrace bordering a stream in a narrow valley will form whenever the stream changes the direction of lateral migration of its channel and thereby undercuts the bank of the former flood plain.

On King's (1950) map of the Elkton area, however, the terraces of that area are grouped in a sequence of three. Close examination of King's map (1950, pl. 1) indicates, as King himself stated (1950, p. 58), that the classification of gravel deposits is somewhat arbitrary. For example, in some places the older gravel deposits are highly dissected and have considerable relief, as near the Ingham mine and in the area directly east of Elkton. At other places, as north of the Boyer mine, the older deposits have a smooth, terracelike surface that merges gradually with the intermediate surface. Nevertheless, if the assumption is made that the terraces do have continuity, then King's grouping is certainly the most reasonable one that could be made. The present writer did not find an obvious tripartite grouping of the alluvial terraces in the North River basin, and in the small-scale map of this area published earlier (Hack, 1960a, fig. 4), he grouped the terrace deposits older than the Recent flood plains into two units rather than three. Nevertheless, the general continuity of King's terraces in the Elkton area is impressive enough to indicate that the grouping of terraces may not be entirely the result of an arbitrary classification scheme.

The bottom lands along some parts of the course of the Middle River (p. 51) also are too continuous to have their origin explained by the shifting of river bends. This is especially true of the long reach in the Martinsburg Shale, where the bottom lands are at three distinct levels for a distance of 25 miles (fig. 32). Some process or processes must operate that cause fairly long segments of a river flood plain to be abandoned or become intrenched at very nearly the same time.

The explanation of the terraces suggested by King (1950, p. 60) involves the change in regimen that presumably occurred as a result of climatic change. During a change from a cold climate to a warm climate, the load supplied to the stream may have decreased while the peak flood discharge may have increased. Such a change would presumably increase the capacity of the stream and bring about a change in gradient by enabling the stream to carry the same load on a gentler slope. The result would be an intrenchment of the channel and the formation of a terrace.

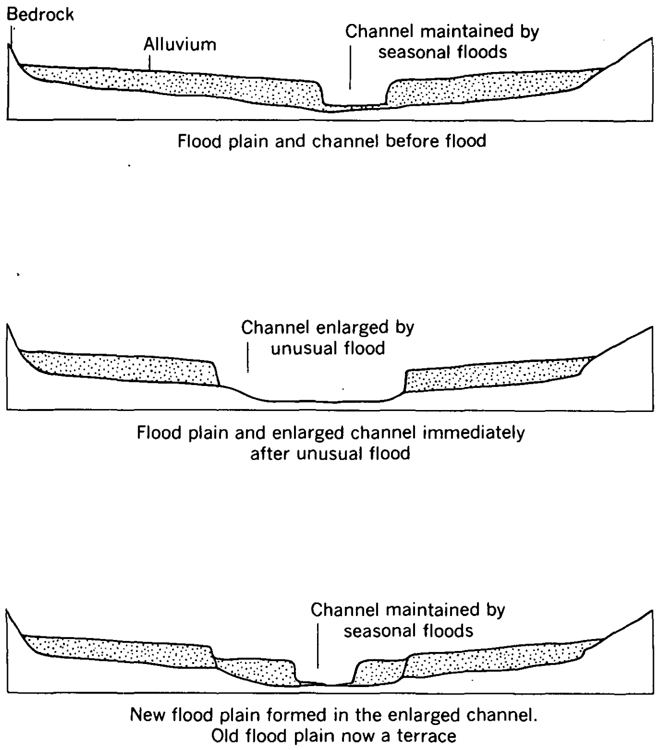

FIGURE 36.—Formation of a second bottom or terrace by the enlargement of the channel during a great flood.

This explanation might account for the apparent widespread arrangement of the alluvium in distinct but poorly defined levels having considerable lateral continuity. This mechanism would operate over large areas and cause widespread changes in regimen at the same time. Obviously, however, this is not the only mechanism that can cause the abandonment of the flood plain. Others operate that might form terraces having continuity only within a limited area. The terraced bottom lands of the Middle River, for example, might have been caused by severe floods large enough to destroy or modify stretches of the valley bottom (fig. 36). In the North River area the flood of June 1949 (Hack and Goodlett, 1960) caused the formation of terraces in many places.

Changes in regimen would also be caused by piracies in the piedmont area downstream from a drainage basin in resistant rocks. The piracies would affect the area of the capture itself and possibly the regimen downstream by altering the size of the drainage basin and, for a time at least, the load.

The evidence in the Shenandoah Valley, particularly the distribution of the alluvial deposits, indicates that the primary cause for the deposition of alluvium in terraces and piedmont aprons is a contrast in resistance between the rocks in different parts of a drainage basin. The lateral continuity of the flights of terraces may have several causes, only one of which is climatic change. Other factors may include the occurrence of violent floods and the lateral shift of the channel, as well as changes in regimen that result from piracies.

SOLUTION FEATURES IN CARBONATE ROCKS

Erosion through solution is an important process in the Shenandoah Valley. The present effectiveness of solution is evidenced by the high calcium and magnesium content of the waters issuing from the valley in the Shenandoah River. For example, analyses of the waters of the North Fork of the Shenandoah River at Strasburg average 36 ppm (parts per million) of calcium and 9.4 ppm of magnesium. Bicarbonate content is 138 ppm (Schroeder and Kapustka, 1957, p. 24). The evidence obtained in the present study seems to indicate that in the Shenandoah Valley most of the erosion by solution goes on at or near the surface, although a network of solution openings in the rocks extends to considerable depth.

UNDERGROUND OPENINGS

Evidence of underground openings in the carbonate rocks is available in records of water wells and in caverns that have been explored above the water table. The writer has no original information to offer, but a brief review of the published data on wells and springs will serve to show that openings in the rocks extend to considerable depth. In a study of the ground-water resources of the Virginia part of the Shenandoah Valley, Cady (1936, p. 34–42) showed that finding water-filled cavities in the carbonate rocks by drilling is a matter of chance. Over one-third of the 300 recorded wells obtained a flow of less than 5 gallons a minute. Only four wells obtained a flow of more than 100 gallons a minute. Cady's data seem to indicate that the probability of a well being successful does not change with depth of drilling from the surface to 300 feet. Below that depth, however, the probability increases, and the average yield of the 16 deep wells drilled is three times as great as that of the shallower wells. Of these wells, four are more than 1,000 feet deep, the deepest reaching 1,926 feet. A well 1,432 feet deep at Winchester, Va., has a yield of 150 gallons a minute. These data appear to be good evidence that openings in the carbonate rocks are not restricted to shallow depth but occur to depths at least as great as 2,000 feet. Cady concludes from his study that considerable artesian circulation of water occurs at depth along fault planes and bedding planes associated with the major synclinal structure of the valley.

Cady's hypothesis of deep circulation of water in the carbonate rocks is strengthened by the occurrence of thermal springs in the Shenandoah Valley. The springs have been studied by Reeves (1932). Fifteen springs classified as "warm" are included in the area of this report; most of them are on the northwest limb of the Massanutten syncline. The warm springs all have temperatures below 80°F, but their mean temperatures exceed the mean temperature of the air where the springs issue. The spring waters are mostly bicarbonate waters similar to those of the streams and cold springs of the carbonate rock area. However, the total content of dissolved solids is considerably higher than the average for cold springs. Discharges of these springs have a wide range, but many of them exceed 500 gallons a minute. The variation in discharge is low. The temperature of the warm springs remains constant and is unaffected by seasonal changes. Reeves believed that these facts support the hypothesis that the spring waters have been warmed by deep circulation in artesian basins where the temperature of the rock exceeds that at the surface.

Many springs of the Shenandoah Valley deposit large quantities of calcium carbonate on issuing from the ground. Falling Spring Creek, 4 miles north of Staunton, for example, heads in a warm spring and flows for several miles through a valley whose bottom lands are composed of marl. It enters the Middle River at a cascade about 50 feet high over banks of travertine built by the water. The large quantity of carbonate deposited downstream from the orifice of this spring attests to the presence of extensive solution cavities in the rocks somewhere below the orifice.

The evidence from study of the springs and of holes drilled for ground water in the carbonate rocks thus suggests considerable circulation of water at depths much greater than 1,000 feet. The openings in the rocks through which the circulation is channeled, however, may not be very large. Among the largest of the warm springs in the Shenandoah Valley is Quick Spring, which forms the source of Falling Spring Creek (Reeves, 1932, p. 47). The discharge of this spring is 1,000 gallons a minute, which corresponds to only 1.5 cubic feet per second, and is equivalent to the average discharge of a surface stream having a drainage area of only 1 or 2 square miles.

SINKS

The most obvious evidence of solution is the hundreds of surface sinks that occur in the carbonate rocks of the Shenandoah Valley. One of the most important factors controlling the distribution of sinks is the character of the bedrock. This factor has been analyzed by plotting all the sinks visible on aerial photograph in the southern part of the Shenandoah Valley (pl. 3). Examination of this map indicates that sinks are most abundant in the Ordovician rocks (table 6).

No explanation for the abundance of sinks in the Ordovician rock areas appears to be completely satisfactory. Sinks in general are more abundant in areas where residuum on the carbonate rock is thin or lacking. This fact can be verified by comparing plate 3 with plate 2. Sinks are especially abundant in the upper part of the Beekmantown and in the Middle Ordovician rocks, and these rocks are commonly covered only in thin residuum. On the other hand, sinks are formed in some places where residuum is 20–30 feet thick or more; along the foot of the Blue Ridge, they occur under a cover of alluvial cobbles and residuum that may be more than 100–200 feet thick.

Sinks commonly are heavily concentrated in fields or occur in clusters. These may be related to synclinal structures in many places—for example, the field of sinks northeast of Staunton. The sinks north of Mount Crawford are in a broad synclinal belt, and other similar areas of concentration occur.

TABLE 6.—*Distribution of sinks in the southern part of the Shenandoah Valley in relation to the kind of carbonate rock*

Rock	Area (square miles)	Total number sinks	Number of sinks per square mile
Covered by alluvial deposits	103	262	2.54
No alluvial cover:			
Ordovician age	149	636	4.27
Cambrian age	205	276	1.34

Caverns and sinks are apparently not more abundant in limestone than in dolomite. Thus, caverns and sinks are numerous in the dolomitic Beekmantown as well as in the calcareous Middle Ordovician rocks. Luray Caverns and the surrounding large field of sinks are formed in crystalline dolomite. Study of the rocks in the Caverns themselves indicates that solution is favored in the more coarsely crystalline beds, though structural features such as joints are obviously an important control (Hack and Durloo, 1962).

Sinks are more abundant along large streams than they are in interstream or in headwater areas. Among the reasons for this are (1) the inflow of ground water to the streams, (2) the steepening of the ground water gradient near large intrenched streams, and (3) the fact that many streams enter the carbonate rock from the mountains and are low in dissolved solids and alkalinity. This third factor has been discussed elsewhere (Hack, 1960b), and sinks were shown to be almost wholly lacking in the carbonate rock area where no streams enter the valley from noncarbonate rock areas. This is particularly true in the northern part of the valley north of

Massanutten Mountain, where sinks are scarce except along the main streams. The absence of sinks is noticeable in spite of the presence of broad areas of apparently suitable rock. A similar absence of sinks is evidence in the area shown on plate 3 at the southern end of the valley, where the streams all head in the carbonate rock area itself. Considerable water must circulate laterally in the ground some distance away from the streams, however, for as shown on plate 3, the sinks are by no means concentrated immediately adjacent to the main streams.

Sinks covered by alluvium occur in many areas (pl. 3). In some places, as along the North River and along East Dry Branch, low gravel terraces bordering the streams are almost wholly occupied by sinks and their intervening ridges, and very little flat area is left on the terrace.

Numerous sinks covered by alluvium and residuum occur along the northwest slope of the Blue Ridge on the Tomstown Dolomite and Waynesboro Formation. These sinks are discussed in connection with the origin of the manganese deposits (page 72).

The sinks appear to show that solution openings are formed in certain beds of rock and are concentrated in the area adjacent to large streams particularly where water of low alkalinity circulates. Great differences in hydrostatic head, such as might occur adjacent to the mountains that border the Shenandoah Valley, would favor underground solution. If this analysis of the solution openings is valid, many of the sinks and solution cavities in the valley must be inferred to have their origin at moderate depth and are not simply funnels opened by water draining the area immediately adjacent to them. The sinks may undergo considerable enlargement after they are exposed at the surface, even though they were originally formed and connected with the ground-water reservoir at some time in the distant past and at greater depth.

SOLUTION AT THE SURFACE

Although solution openings and sinks are conspicuous features in the Shenandoah Valley, they by no means occupy a large proportion of the area. The density of sinks, as shown by table 6, is really very low; the average is only about 3 per square mile. Many large areas of carbonate rocks exhibit no sign of solution pits and have a regularly ordered topography of ridges separated by open ravines. Even in areas occupied by many sinks, the parabolic form of slopes charactistic of interstream areas (p. 20) is the dominant type. This regularity and similarity to areas underlain by noncarbonate rocks indicates that here subsurface solution is not important.

If the dominant erosion were by solution underground, we might expect not only the form of the hills to differ from that of noncarbonate areas but also the drainage density to be considerably less. Drainage densities of small sample areas, averaging 1 square mile in size in different rock types, have been measured and are shown in table 7. The drainage density in the limestone area is indeed the lowest, but it is only slightly lower than that of the quartzite areas of the Blue Ridge, where mechanical erosion is of paramount importance. The streams in the limestone area, like those in areas of other rocks, are floored with bed material of angular cobbles and boulders generally locally derived, a fact indicating that the channels are cut at least partly by mechanical processes (Hack, 1957, p. 76–81).

TABLE 7.—*Density of channels in areas of different kinds of rock in the Shenandoah Valley, as measured from aerial photographs*

	Rock	Area (square miles)	Drainage density (miles of channel per square mile)
1	Athens Limestone	1. 45	2. 4
2	Quartzite and phyllite of the Chilhowee Group	1. 35	3. 2
3	Granodiorite	1. 28	3. 4
4	Sandstone and shale of the Chemung Formation	1. 43	6. 6
5	Martinsburg Shale	1. 17	7. 6

NOTE.—The following are the locations of the areas:
1. Tributary of Middle River, 5.2 miles northwest of Staunton, Va.
2. Rip Rap Hollow, 9.5 miles northeast of Waynesboro, Va.
3. Headwaters of Hawksbill Creek, 7.5 miles southeast of Luray, Va.
4. Headwaters of East Dry Branch, 12.8 miles west of Staunton, Va.
5. Drainage basin of Christians Creek, east of Staunton, Va.

In the writer's opinion, topography in most of the limestone region is graded for the transportation of material along the surface. Although the carbonate rock itself must go into solution, the insoluble residue present in the limestone beds in small to large amounts remains on the surface and determines the grade and form of the slopes. Probably the most effective solution of the limestone takes place at the very surface of the rock, for most of the water that penetrates below the surface except in periods of heavy rainfall may already be saturated in carbonate. Solution-pits and characteristic lapies on the surface of the limestone bedrock are seen whenever the bedrock surface is stripped of soil for quarrying or other engineering operations. Core stones of limestone are commonly found in the thinner residual soils (p. 45). These are like the core stones of igneous rock areas and result from differential weathering of the bedrock. In some areas, mostly small in extent, the effects of underground solution dominate the landscape, and the drainage is underground.

TRANSPORTATION OF ERODED MATERIAL

One of the corollaries of the equilibrium theory of erosional landscapes is that all elements of the topography in a single dynamic system are downwasting at the same rate or at a rate proportional to the average slope (p. 9). Differences in relief within the system are the result of differences in rock resistance which affect the forms of slopes and stream channels. In accord with a cyclic theory of landscape development, however, the highland areas are presumed to be relics of highlands that were once more extensive and are preserved in the landscape because they are more resistant to erosion or have not been reached by agents of renewed activity and are thus downwasting at a slower rate than the lowlands. It is of interest, therefore, to examine the content of the streams draining the Shenandoah Valley in order to judge whether the different rock types are in fact being eroded at the same rate. Unfortunately, records that show the sediment content and chemical character of the water do not cover a long enough period nor a wide enough range of conditions to make possible a quantitative analysis of the problem. However, the scant records indicate that the rate of transport of materials of different kinds is of the right order of magnitude to be in harmony with the equilibrium theory.

The most complete record of sediment transport in a large stream in the Shenandoah Valley is that for the South Fork at Front Royal (Schroeder and Kapustka, 1957). The drainage basin of the river contains contrasting rock types and topography; about half the area is underlain by carbonate rocks and half by clastic rocks (table 8).

TABLE 8.—*Area and percent of basin underlain by principal rock types in drainage basin of the South Fork Shenandoah River above Front Royal, Va.*

Principal rock type	Area (square miles)	Percent of drainage basin
Greenstone and granodiorite	151	9
Quartzite and argillite of the Chilhowee Group	170	11
Shale	278	17
Sandstone and shale (interbedded)	264	16
Carbonate rocks	783	47
Total	1, 646	100

The concentration of dissolved solids in stream waters is closely controlled by the geology of the drainage basins; streams draining areas of limestone have a concentration of dissolved solids commonly several times higher than do streams in clastic rocks. If the material eroded from the carbonate rock area is mostly transported in solution and if the material eroded from the noncarbonate area is transported largely in suspension and along the bed, then, according to a cyclical theory of erosion, the quantity of material transported in solution should be several times greater than the quantity transported in suspension. This conclusion would follow from the presumption that the carbonate-rock area is eroding more rapidly than the more resistant clastic-rock areas.

According to the equilibrium theory, however, the rate of erosion in the carbonate area is no different from that in the clastic rock area. Therefore, the amount of material transported in solution should be equaled or exceeded by the amount transported in suspension, because the carbonate rocks contain clay, sand, and chert that may constitute 10–20 percent of the total thickness; although some of this material is dissolved, much of it must be carried off in suspension.

The transportation of material in the South Fork in suspension probably does exceed the amount transported in solution. Measurements of the concentration of both suspended sediment and dissolved solids are available at Front Royal for the period April 1953 to September 1954 (Schroeder and Kapustka, 1957), and as the river is flowing on shale bedrock, nearly all the material eroded from its drainage basin may be assumed to have passed this point in the stream. The only complete year of record is the water year October 1, 1953–September 31, 1954. During this year, according to the estimate of Schroeder and Kapustka (1957, p. 25), the transport of material in suspension amounted to 77,129 tons. Using their chemical data, the present writer calculated that the transport of dissolved solids was 155,000 tons, approximately twice as great. However, the water year 1953–54 was not a typical year, for the average discharge was considerably less than normal during this year, and the maximum discharge was only 17,000 cfs. The maximum recorded discharge is 130,000 cfs, and the 17,000 cfs maximum that occurred in 1954 has been exceeded in 15 out of the 24 years of recorded flows.

The concentration of suspended material increases at a high rate with increasing discharge, whereas the concentration of dissolved solids decreases. Presumably, therefore, a longer period of sampling would indicate that a somewhat higher proportion of the material is carried in suspension.

This presumption can be supported by further calculation and extrapolation by use of records of mean daily discharge. Based on the period during which sediment analyses are available, curves showing the concentration of sediment and dissolved solids are constructed as shown in figure 37. As shown by this

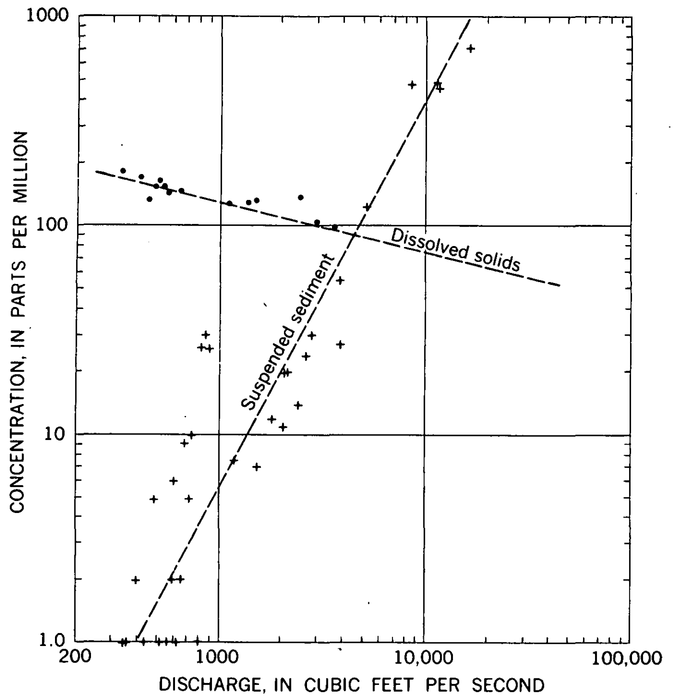

FIGURE 37.—Concentration of dissolved solids and suspended material at different discharges in the South Fork Shenandoah River at Front Royal, Va. Data from Schroeder and Kapustka (1957).

graph, the concentration of suspended sediment is roughly proportional to the 1.75 power of the discharge, a very high rate of increase. It appears that during one large flood of very short duration more sediment may be carried in suspension than during a whole year of average flow. The rate of transport in suspension begins to exceed the rate in solution at a discharge of 4,000–5,000 cfs. The writer has calculated that about 25 percent of the total water flow of the South Fork is normally discharged in flows that exceed this amount.

Although the water year 1953–54 is atypical, the data shown in figure 37 can be extrapolated and applied to the discharges characteristic of a larger period of record. For this purpose, the period 1931–42 has been chosen for analysis. This period includes the maximum recorded flow as well as flow during several abnormally dry years. The results have been summarized in table 9. This table shows that the high flows and the high sediment concentrations that must be associated with them are sufficient to raise the amount of material transported in suspension to a value larger than that transported in solution. Thus a flow of 15,000 cfs occurring on 2.8 days per year transports almost as much suspended material as is transported in solution during the entire year. The totals given do not include the amount of load transported during approximately 1 day in which the average discharge exceeds 20,000 cfs because this value is beyond the range

TABLE 9.—*Computation of annual load, in tons, carried by the South Fork Shenandoah River at Front Royal, Va., based on the average daily discharges for the period 1931–42*

[Concentrations are estimated from the graph (fig. 37)]

Discharge (cfs)	Assumed concentration		Average number of days per year	Annual load	
	Suspended sediment (ppm)	Dissolved solids (ppm)		Suspended sediment (tons)	Dissolved solids (tons)
0–200	0.08	240	2.2	----------	427
200–400	.6	180	49.2	23	7,173
400–600	1.5	160	70.6	166	15,249
600–800	2.8	145	45.2	405	12,387
800–1,000	4.3	135	35.3	774	11,580
1,000–2,000	11	128	95.7	5,037	49,610
2,000–4,000	39	100	42.1	18,367	34,182
4,000–6,000	100	90	12.1	34,702	14,701
6,000–8,000	190	83	5.8	55,529	9,098
8,000–10,000	300	78	2.7	75,213	5,117
10,000–20,000	660	70	2.8	150,056	7,938
20,000–40,000	----------	----------	.9	----------	----------
40,000–60,000	----------	----------	.3	----------	----------
60,000–80,000	----------	----------	.0	----------	----------
80,000–100,000	----------	----------	.1	----------	----------
Total	----------	----------	365	340,272	167,462

of the data in figure 37. If this unknown but large load were included, obviously the amount of material transported in suspension would exceed the dissolved load by a very large factor.

Unfortunately, the foregoing analysis ignores several factors. For example, the load transported along the bed has not been measured, and we have no way of judging its amount. Gravel, cobbles, and boulders on the stream beds are probably moved only in flows that have a very low recurrence interval. In severe floods, however, the amount of coarse material transported may be quite large. Some carbonate rocks as well as sandstone and quartzite may be included, although studies of the bed material in carbonate-rock areas indicate that carbonate rocks are not transported very far in the form of gravel (Hack, 1957).

Another important factor that the analysis ignores is the condition of the watershed. As the carbonate-rock area is heavily populated and is crossed by a dense network of roads and farm trails, the silt and clay cover may erode more rapidly at present than it has in the past. This factor might increase somewhat the amount of sediment transported in floods. Soil erosion, however, is at present certainly not severe in this area. The carbonate-rock area is largely in pasture and well covered with sod.

A small error is introduced into the calculation of the material removed from the watershed in solution by neglecting the dissolved solids contained in rainwater. So far as the writer knows, no analyses of rainwater have been made in this area. The content must be very small, however, and amount to only a few parts per million. This error could not affect appreciably the results of the foregoing analysis.

Another qualitative test of the theory is provided by a comparison of the loads carried by two streams draining areas of diverse rocks. For this purpose the drainage basin of the South Fork at Front Royal, which is half in carbonate rock, is compared with that of the Hazel River at Rixeyville. The Hazel River drains the east side of the Blue Ridge outside the area shown on plates 1 and 2, and its basin adjoins that of the South Fork along the crest of the range. Its drainage basin contains only crystalline rock, and the content of dissolved solids in the stream is only one-third as great as that in the South Fork, averaging about 38 ppm (Schroeder and Kapustka, 1957).

Table 10 shows the results of this comparison. Because the Hazel River is relatively small compared with the South Fork, the data must be expressed in terms of tons per square mile. Obviously, the Hazel River transports for each square mile about half as much material in solution as does the South Fork. This amount is compensated, however, by a much greater suspended load, and the total loads per unit area of the two streams are about equal. As in the previous analysis, the data do not include bed load and they do not take into consideration the possible effects of differences in the condition of the watershed.

Although all factors cannot be considered, records of the sediment load and the chemical quality of water in the region do not in any way suggest great differences in the rates of erosion between streams draining areas of different rock. Critical differences may exist, but they are not large enough to be detected in the data available.

TABLE 10.—*Comparison of the loads discharged by two streams in different geologic environments, October 1953–September 1954*

Stream	Average concentration of dissolved solids (ppm)	Total yearly discharge (cfs-days per square mile)	Transported load (tons per square mile)		
			Dissolved solids	Suspended sediment	Total load
South Fork Shenandoah River at Front Royal, Va. (Drainage area, 1,638 square miles)	160	221	98	47	145
Hazel River at Rixeyville, Va. (Drainage area, 286 square miles)	38	204	52	78	130

The average rate of erosion in the Shenandoah Valley can be calculated by use of the estimates shown in table 9 of the suspended and dissolved solids discharged past Front Royal. All discharges greater than 20,000 cfs are not included in these estimates, even though larger discharges on the average probably occur more than 1 day a year. Obviously, such an estimate would be too low. The sediment carried past the station during 1 day of extraordinary discharge may even equal or exceed the total amount carried during the entire year by lower discharges, and a calculation is likely to be something less than one-half the correct figure. If 340,272 tons is the annual discharge of suspended sediment, 167,462 tons is the annual discharge of dissolved solids, and the average density of the rocks is estimated to be 2.8, the average rate of denudation of the basin of the South Fork is 0.00013 feet per year, or approximately 1 foot in 7,700 years.

If the area had been denuded continuously at this average rate, more than 13,000 feet of rock would have been removed in the 100 million years since Cretaceous time and 23,000 feet of rock in the 180 million years since the end of Triassic time. Because the rate of denudation may have been more than twice as great, there is no difficulty in accounting for the removal of a thickness equal to the entire Paleozoic sequence, generally estimated at 30,000 feet.

INTERPRETATION OF PHYSIOGRAPHIC HISTORY

The interpretation of the landscape in the present report emphasizes the continuity of geomorphic processes. Almost no reference has been made to stratigraphy, and little attempt has been made to arrange any of the surficial deposits in a chronologic sequence. The interpretation is based on interrelations of processes, rocks, and other features that exist today as though they were almost time independent. This does not mean that the deposits could not be arranged in a sequence or that there have not been any changes in the landscape. Any or all of the surficial deposits could be arranged in a time sequence by comparing their height above the stream, degree of dissection, degree of weathering, and other characteristics that are useful in stratigraphy.

The purpose of a study of the equilibrium relations among all the features of a landscape is to assess those that are not in equilibrium and that therefore may reflect historical changes in the development of a landscape. Stratigraphic studies of surficial deposits within the valley are not useful without an understanding of the equilibrium conditions. Individual bodies of gravel in a piedmont alluvial apron, for example, are arranged in flights of terraces, and some of the units have unconformities between them, which indicate that deposition ceased and erosion occurred at that place. Such relations do not mean that deposition alternated with erosion simultaneously throughout the region. Both erosion and deposition have gone on continuously, but the activity has shifted locally from place to place.

As we can account for most, if not all, of the geomorphic features of the Shenandoah Valley on the basis of an interrelation between present processes and rocks, we need not postulate any major changes in the past. If deposits or forms were found that were completely out of harmony with the processes of the present, then we would conclude that there had been a change in the past, and the deposit or form in question would immediately become a sort of stratigraphic landmark. If, for example, a terrace in the lower valley were found composed completely of clay different from the graded gravelly flood plains that form along the streams at present, we would be forced to the conclusion that a change had occurred in the processes that operated in the valley.

Analysis of the Shenandoah Valley as though it were in a time-independent state does explain most of its major features. Minor ones, such as the scree deposits of the Blue Ridge, appear to be out of harmony with the explanation, for the scree seems far too extensive for all of it to have been deposited under conditions like the present and some large areas of scree appear to be separated from any source area. Some of the alluvial terraces are also not fully explained. Major features, however, like the form of the hills, the variation in relief, the piedmont alluvial aprons, and many others could have formed in an environment like the present and can be explained as though they were time-independent.

Nothing seems to compel a belief in any important recent diastrophic movements or other important recent environmental change within the Shenandoah Valley itself that caused an evolution of the landscape from one form to another. Such changes must have happened but so long ago that the evidence is not preserved in the present landscape. A problem remains, however, in interpreting the development and history of the landscape during the time before it was brought near to its present condition, for obviously the landscape must have been very different near the close of the Triassic, and many changes, including those brought about by diastrophism, must have occurred.

Very little evidence has been found within the Appalachian Mountains themselves to aid one in speculating on the early evolution of the range. Erosion has probably been continuous ever since the initial uplift of the Appalachians. Direction of the original drainage is unknown, because as the region rose above the sea the drainage may have been superposed from deposits of Permian, Triassic, or Jurassic age that no longer exist. The uplift that followed the initial rise and folding may have required a very long span of time and may have persisted through the Jurassic period. Slower movements of various parts or all of the mountain system may have persisted through the Teritary. During the Mesozoic and Tertiary, about 30,000 feet (King, 1959, p. 57) of sedimentary rock was probably removed from some areas, and major changes in the drainage could have taken place as the rivers became more completely adjusted to varying kinds of rocks exposed on the erosion surface. No basis exists for thinking that the relief was lower than the present at any time during the Tertiary, and it may well have been higher.

Superposition of some of the streams may have taken place during the downwasting of the region and caused maladjustments. As the amount of alluvium deposited in the lowland areas is a function of the contrast in relief between hard and soft rock areas, the gravel deposits in the lowlands must have been more extensive in the past, when the relief was greater than now. Locally streams may have become superposed on underlying structures as the gravelly mantle was reduced in extent during erosion. Local superposition could also have occurred during the downwasting of the area where a stream eroded through a thrust sheet or limb of a fold and became enclosed in the different rocks on the underside, as postulated in the theory of Meyerhoff and Olmsted (1936) and by Bloomer (1951).

It seems unrewarding to speculate further on the basis of evidence now available, especially as many untapped sources of evidence may enable more fruitful speculation after further work. The Coastal Plain, for example, now very imperfectly understood, is underlain by thousands of feet of sediments shed by the Appalachian Highlands since Cretaceous time that may yield valuable clues. A better knowledge of the structural development of the Appalachians, particularly of the Piedmont region, is also needed. The detailed geologic mapping of the last 30 or 40 years has already changed some of the concepts of Davis' time that formed part of the basis for his thinking. These changes include the abandonment of the concept of Appalachia as well as a growth in appreciation of the great complexity of Appalachian structure.

The theory of the geographic cycle as propounded by Davis provided the basis for an involved and detailed theory of Appalachian history that traced its development through the entire Mesozoic and Tertiary. Basically, the theory depended on the mistaken belief that the present topographic surface could be subdivided into several different erosion surfaces, each of which was the remnant of an ancient topographic surface and therefore could be treated as though it were a stratigraphic horizon. The theory of Davis was greatly elaborated by later workers. The classic treatments are those by Davis (1889, 1890a) and by Johnson (1931). Concise summaries have been written by many authors and include those of Bryan, Cleaves, and Smith (1932a), Fenneman (1938, p. 195–278), and Thornbury (1954, p. 230–240).

A few of the most important problems dealt with by the theory of the geographic cycle as applied to the Appalachians are discussed very briefly in the following paragraphs. They serve to point up the differences between the classic theory and the theory of the equilibrium landscape. Some of them have already been mentioned earlier in the report.

REVERSAL OF DRAINAGE

The hypothesis of drainage reversal was conceived by Davis (1889, p. 219–224) because the major streams of the central Appalachians cross structural features from younger rocks to older, and presumably a much greater sequence and thickness of rocks must have been eroded from the southeastern border of the range, where the streams leave to enter the Piedmont, than from the northwestern border in the headwater areas. Johnson (1931), concerned with the obvious difficulty of reversing an already established drainage, postulated that the reversal occurred after a marine invasion in the Cretaceous covered the Appalachians with sediments from which the drainage later became superposed.

As the kind, age, or thickness of the rocks on which the Appalachian drainage originally developed is not known, it does not seem to the writer that Davis' hypothesis of drainage reversal need be taken very seriously. Modern concepts of the structural development of the Appalachians are more sophisticated than the concepts on which Davis based his hypothesis and suggest that the general direction of drainage might have been like that of the present ever since Triassic time (Thornbury, 1954, p. 239). Studies of heavy minerals in the Coastal Plain of Delaware suggest that the source of sediment in that area has been to the northwest since at least Cretaceous time. If a reversal of drainage took place, it must have been by headward migration of divides and was well advanced by late Cretaceous time (Groot, 1955, p. 123). The concept of superposition from a cover of Cretaceous sediments is not supported by any evidence in the Shenandoah Valley or the finding of any marine deposits anywhere in the Appalachian Highlands. The adjustment of most of the streams to rock structure is remarkably close. It is hard to see how the course of the Potomac, for example, could have been established by superposition and yet cross both North Mountain and the Blue Ridge at the narrowest and weakest places.

THE SCHOOLEY PENEPLAIN

The concept of the Schooley peneplain was postulated by Davis (1889, 1890b) to explain the supposed accordance of summits of large areas of mountains. No real accordance of summits exists, but altitudes of ridge summits are related to the character of the rocks that form the crests (Thompson, 1941, and Edmundson, 1940). As shown in the preceding pages, the altitudes of the higher mountains vary considerably, and their heights are directly related to the widths of the outcrop belts of the most resistant rocks as well as their distance from the principal streams. The relation is particularly well illustrated by the Blue Ridge (p. 28).

LATER PENEPLAINS AND LOWLAND TOPOGRAPHY

Davis (1889, p. 199) in his original theory postulated only one erosion surface of low relief lower than the Schooley peneplain, and he believed that it was of Tertiary age. This surface was later called the Harrisburg peneplain by Campbell (1903). It supposedly is preserved on the smooth rounded hilltops of carbonate rock and on shale lowlands in large parts of the Appalachians, and in places it is presumably covered with residuum and gravel. There has been little agreement on the interpretation of the lowland belts, and the opinions of geomorphologists have ranged from the idea that the lowland topography is entirely the result of differential erosion to the idea that at least five erosion levels exist, each one a partial peneplain controlled by a former base level. The most common view is that the lowland forms result from the dissection of only one peneplain, the Harrisburg (Thornbury, 1954, p. 239–240).

Few geomorphologists working in the central Appalachians have done any extensive mapping of the surficial deposits, and it was not until the work of King (1950) in the Shenandoah Valley that a comprehensive study and map were made of the deposits on the lowland surface. Although King accepted the existence of the Harrisburg peneplain, he pointed out a serious difficulty in the concept. The middle Shenandoah Valley above Front Royal has far too steep a gradient to have been inherited from a peneplain formed when the streams were controlled by base level as according to older concepts. The area, therefore, must have been warped since the formation of the peneplain. The gravel deposits, however, that mantle large areas in the middle Valley are graded to the present drainage. Thus, the gravel deposits cannot be related to the peneplain but must have been deposited after the warping. King (1950, p. 12) regarded the residuum beneath the gravels as inherited from Tertiary time and probably formed on the peneplain.

In the present report, another step is taken, for the steep gradient of the middle Shenandoah Valley is shown to be related closely to the sinuosity of the North and South Forks of the Shenandoah River (p. 29). The steep gradient is therefore related to the processes of the present, not only to those of the past, as also are the gravel deposits. The peneplain is not expressed in the topography of the present valley, and there is no reason to believe that it ever existed.

If the residuum is mapped throughout a large area, it is found to occur notably on carbonate rocks wherever they are mantled by resistant rock fragments of either alluvial or residual origin. In addition, erosion by solution occurs beneath the protective cover of fragments, and residuum has formed at depths as great as 100 feet below present streams. This subject is further discussed on page 72, but the evidence seems to indicate that present conditions in places are suitable for the formation and preservation of residuum. Presumably residuum has always been present in the Shenandoah Valley and always will be present as long as an appreciable amount of relief exists. It is simply an inherent part of an erosional system in which some rock constituents weather and are removed more readily than others.

The extensive gravel deposits in the Shenandoah Valley are analogous to the residuum, in that they too are lag deposits of the more resistant materials. They consistently occur in soft-rock areas along streams whose headwaters cross either resistant rocks or soft rocks that contain resistant components. The coarse resistant fragments are not carried off in the stream channels on the low gradients that prevail in the softer rocks. They are therefore in effect stored in the landscape in terraces and piedmont aprons while weathering reduces them to sizes or components that can be removed.

DEPOSITS INHERITED FROM COLD PERIODS

Many of the gravel aprons, terraces, and talus deposits are obviously of Pleistocene age, and conditions during the cold intervals of the Pleistocene probably favored their formation (Smith and Smith, 1945; King, 1950). Analysis of the deposits suggests that although this relation is probably true other factors are equally important and that gravel deposits would exist and talus might exist even if the climate had always remained the same as it is now. Formation of talus requires not only cold but also steep slopes and a relatively high rate of erosion at the base. It must also be composed of rock that is resistant to weathering. The effect of intense cold is probably to cause the accumulation of talus on gentler slopes and in areas of lesser relief than in times of warm climate. The same might be said of terrace deposits, though the importance of the effect of a cold climate on such deposits is more difficult to evaluate. The modern streams are able to transport to the lowlands material just as coarse as that in the terraces and in the talus (page 34 and Hack, 1957); deposition of such material is still going on in many places on flood plains in the lowland areas.

DEPOSITS INHERITED FROM WARM PERIODS

No basis has been found for another postulate: residuum formed mainly during warmer climates, presumably in the late Tertiary, and that present conditions are not favorable for its formation (Hewett 1917; King, 1950 p. 55). Residuum in the Shenandoah Valley is derived mostly from the weathering by solution of carbonate rocks. These rocks are more soluble in cold solutions than in warm. Furthermore, the carbonate and bicarbonate content of the stream waters is high and indicates that much rock is now going into solution. Sinks occur on many of the low gravel terraces, and residuum extends to levels below the present streams. The evidence seems to show that residuum forms at present and will accumulate wherever it is protected from erosion by a lag deposit of either gravel or cherty residues.

SUPERGENE DEPOSITS OF IRON AND MANGANESE OXIDE

Deposits of supergene manganese and iron oxides are abundant in the Shenandoah Valley. They are associated with and commonly enclosed by a residual or alluvial cover on certain carbonate rocks. The supergene deposits have received considerable study over many years and are of particular interest to the geomorphologist because they are thought by some geologists to have formed in early Tertiary time on a peneplain.

Reexamined in terms of the concept of an equilibrium landscape, the deposits of supergene oxides can be explained in a different way. Rather than as relics of processes and conditions that existed long ago, they may be interpreted as having been formed under conditions like those of the present through the action of processes that can be observed today.

The Shenandoah Valley has been an area of active manganese production for about 100 years, though the rate of production has fluctuated widely. The peak periods were in 1880–92 and during the two world wars. Total production of manganese in Virginia beginning in 1838 is estimated at 870,000 tons (Sears, 1957). Much of this production was from the Shenandoah Valley. The Crimora mine north of Waynesboro was once the largest producer in the United States, producing 160,000 tons in 1867–1917. In 1958 the total production in Virginia was 8,184 tons, valued at $648,000 (U.S. Bur. of Mines, 1959, p. 982). The largest producing mine was in the Shenandoah Valley.

During the 19th century the Shenandoah Valley was also an important source of iron ore. Iron occurs in deposits similar to and in places associated with the manganese deposits, but little mining has been carried on since about 1905.

An extensive literature dealing with the manganese deposits, their geology, and origin has recently been reviewed and summarized by Pegau (1958). The mines are associated with rocks of three different ages (fig. 38). The writer has obtained his knowledge of the deposits from the following principal sources: For the deposits associated with rocks of Cambrian age the principal references are Hewett (1917), Stose and others (1919), Knechtel (1943), and King (1943, 1950); the few small deposits associated with Ordovician rocks are described by Stose and Miser (1922); for the deposits associated with Devonian rocks, Stose and Miser (1922) and Monroe (1942) have been the sources of information. The distribution of the deposits as indicated in these reports is shown in figure 38.

DEPOSITS ASSOCIATED WITH THE TOMSTOWN DOLOMITE AND WAYNESBORO FORMATION

The principal group of mines and prospects forms a long row along the northwest foot of the Blue Ridge that stretches from the southern end of the valley northeast to Front Royal. Gaps in the row occur north of Waynesboro and near Luray. Most deposits are in clayey residuum close to the foot of the mountains at a horizon that corresponds to a bed or beds in the lowermost part of the Tomstown Dolomite. Deposits also occur higher in the section in residuum that is clearly derived from the Waynesboro Formation. The deposits include iron as well as manganese oxides. Iron seems to predominate in the deposits at the higher horizon.

The principal manganese minerals that have been described are psilomelane, which constitutes about 75 percent of the ores, manganite, pyrolusite, and wad. The principal iron minerals are limonite and goethite. In many deposits the iron and manganese oxides occur as mixtures in which iron generally predominates. The ore minerals form nodules, pods, and stringers in variegated clayey residuum. In many deposits the nodules have replaced the clay in which they are embedded (Stose and others, 1919, p. 41–45).

The stratigraphic and structural position of the supergene deposits is of several kinds. These are illustrated in figure 5 of Stose and others (1919, p. 50) and are described briefly as follows:

1. The commonest deposits are in the clayey residuum immediately northwest of the outcrop of the Antietam Quartzite, where the dip of the beds is monoclinal northwestward toward the Shenandoah Valley. The residuum and ore are comonly overlain by angular scree derived from the quartzite hills or by cobbly quartzite alluvium. In other places the residuum is not covered but is protected from erosion by areas of alluvium that surround it. This type of deposit is illustrated by the Watson tract (fig. 39).
2. Another type of deposit, in which the Crimora mine is located lies in open shallow synclines immediately above the Antietam Quartzite. The residuum may have a cover of alluvial cobbles, as in the St. Mary's tract described by Knechtel (1943), or it may be uncovered.
3. The manganese may occur in veins in the Antietam Quartzite. One such deposit described by Knechtel (1943, p. 197), called the Hogpen Hollow prospect, occurs in quartzite and is approximately 1,000 feet stratigraphically below the Tomstown Dolomite (Knechtel, oral commun., 1958).
4. At Dargan, Md., north of the Shenandoah Valley, deposits of manganese oxide are associated with a fault that brings the Harpers Formation into contact with residuum derived from the Tomstown Dolomite.
5. The manganese oxides in some places are in gravelly clay that overlies residuum or are in channels cutting residuum of the Tomstown Dolomite.

THEORIES OF ORIGIN

The question of origin of the deposits may be handled as two problems: (1) the original source of the manganese, and (2) the processes that have concentrated or preserved it in deposits of mineable grade. With regard to the first problem, two ideas have been held in the recent past. Most geologists who have written on the problem appear to believe that the iron and manganese are syngenetic and that below the zone of weathering they occur either in the basal beds of the Tomstown Dolomite or in the Waynesboro Formation in the form of iron and manganese carbonate disseminated through the rock in very low concentration. In northeast Tennessee analyses of fresh Shady Dolomite (the stratigraphic equivalent of the Tomstown) have shown that beds 100–200 feet above the top of the Cambrian quartzite contain MnO, probably originally carbonate, in amounts as much as 0.6 percent. The manganese content of the rock appears to be greater in beds unaffected by hydrothermal alteration than in beds that have been so altered (Rodgers, 1945; King, 1950, p. 65–66). Other geologists have thought that the manganese, although syngenetic, was originally in beds either stratigraphically higher or lower than the present position of the deposits. (See King 1950, p. 65, and references therein.)

In contrast, some geologists believe that the original source of the manganese is hydrothermal and that the present deposits are true gossans in which the manganese has been concentrated by the weathering of more thinly disseminated sulfide deposits. This view has

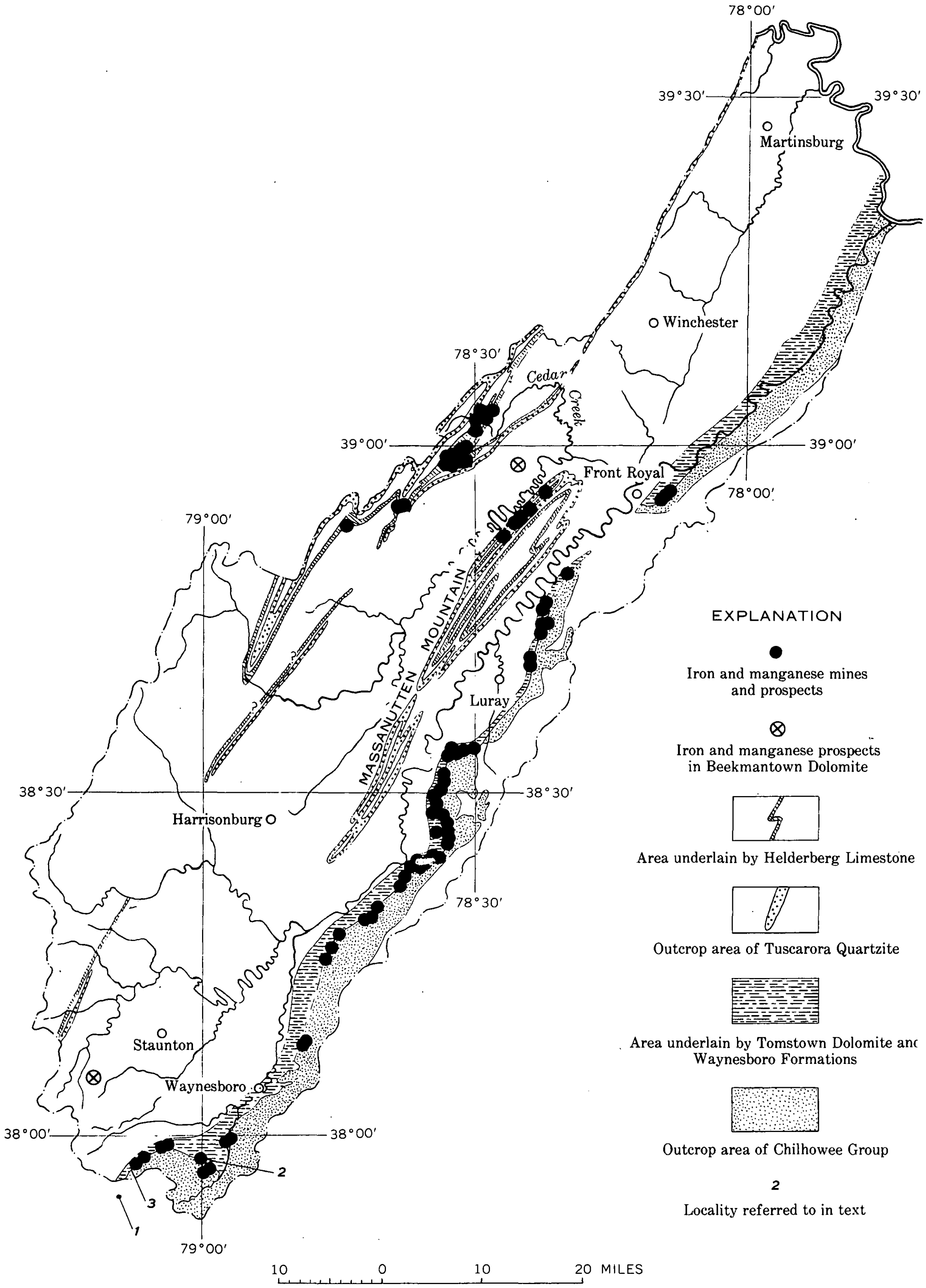

FIGURE 38.—Distribution of principal mines and prospects of iron and manganese oxides in the Shenandoah Valley.

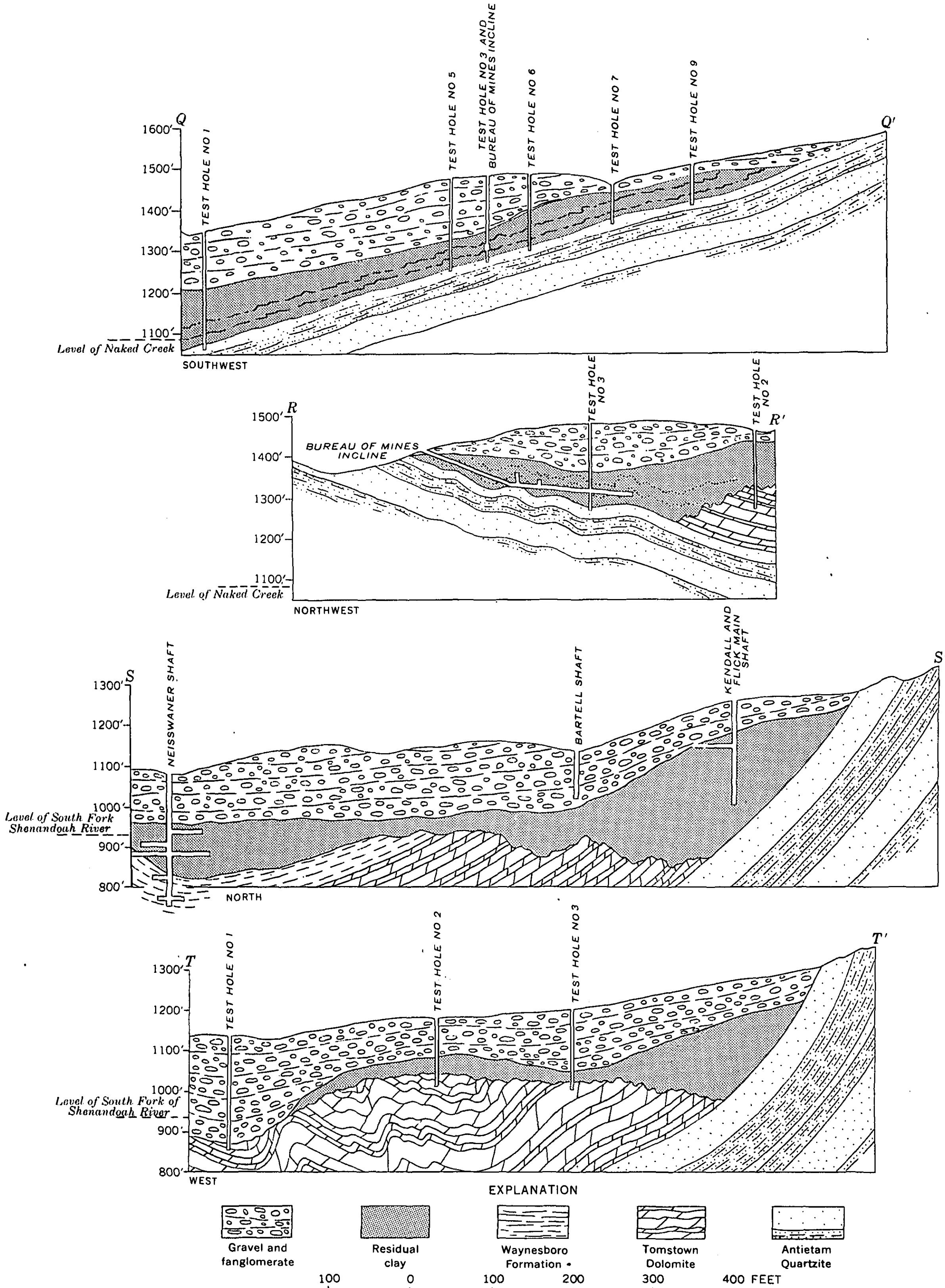

FIGURE 39.—Inferred relations of gravel and residual clay to bedrock at the Watson tract. Reprinted from P. B. King (1950), fig. 16).

been strongly upheld by Kesler (1950, p. 54–57), who worked in deposits of similar age and type in Georgia. Kesler cited the association of the iron and manganese with sulfides and with primary barite. He also noted the occurrence of manganese deposits along faults and the presence of the deposits at more than one horizon in the Cambrian rocks.

The present writer has no evidence bearing on this problem. That the origin is syngenetic is perhaps a more appealing idea, because the known deposits in the Shenandoah Valley are mainly associated with a few beds of carbonate rock in the Tomstown, Waynesboro, Beekmantown, and Helderberg Formations, although the entire sequence of carbonate rocks is thousands of feet thick. Furthermore, the lateral extent of the stratigraphic units that contain manganese is enormous—for example, the principal Lower Cambrian manganiferous unit extends all the way from Maryland to Georgia.

The second problem concerns the processes that have concentrated the iron and manganese in the form of oxides and hydroxides in the clayey residuum. That the manganese was originally sparsely disseminated in the source rock (whether of hydrothermal or syngenetic origin) was apparently first suggested by Harder (1910). Hewett (1917, p. 43) later related the formation of the residuum to the Harrisburg peneplain formed on the Cambrian carbonate rocks at the level of the hilltops. This idea has been followed by Stose and others (1917) and Stose and Miser (1922). These writers believed that conditions during the formation of the peneplain favored deep circulation of ground water and that the oxides were deposited in and replaced the residuum at that time. The deposits that exist today, according to this theory, are remnants of a former more widespread mantle of clayey residuum that have been preserved from erosion. This idea was modified by King (1950), who showed that the Harrisburg peneplain in the Elkton area is not represented by any topographic features now preserved in the Shenandoah Valley. He agreed, however, that the ore deposits and the residuum probably formed in the Tertiary at the time of formation of the peneplain and that the Tertiary residuum is preserved where it became covered by Pleistocene gravels.

The present writer can find no evidence that links the deposits of supergene oxides to the level of a peneplain. The metallic oxides occur in residuum on carbonate rocks at all possible levels, limited only by the outcrop belts of the Waynesboro and Tomstown Formations. Oxides occur below river level as well as on the highest divides. Most of the deposits occur in residuum that overlies basal beds of the Tomstown Dolomite, and these beds seem to be the principal control, though more deposits occur on the upland interfluves than at lower altitudes. The close relation of the deposits to the base of the Tomstown is shown in figure 40. Note that many exploited deposits are in areas of residuum that are not covered by gravel. Such exposures are generally on interstream divides. This relation is expectable as the areas not covered are the ones in which the ores are most easily discovered. Deposits have been mined, however, from clays 100 feet below the flood plain of Back Creek, a major stream (Lyndhurst Mine, Knechtel, 1943, p. 181).

The modification of the peneplain idea suggested by King would permit the formation of residuum at considerable depth below the inferred Tertiary land surface and is in better harmony with the facts as they are now known. The concept of the equilibrium landscape, however, views the development of the valley differently and does not include the concept of a Tertiary peneplain. In the following pages an alternative explanation is offered in which the concept of the peneplain is unnecessary and the concentration of manganese ores is considered to be a more or less continuous process.

STATEMENT OF THE EQUILIBRIUM THEORY OF ORIGIN

The essential element of the equilibrium theory of origin is that the environment at the foot of the Blue Ridge constitutes a mechanical and chemical trap from which the supergene oxides of manganese and iron do not escape immediately but are preserved while other constituents of the original rock are dissolved or otherwise carried away. The oxides are carried away only by mechanical erosion of the residuum from the oldest terraces in the piedmont zone. The mechanical trap is provided by the gravel cover over the residuum in the piedmont alluvial aprons. The chemical trap is provided by the reactions that take place as the limestone is dissolved and the pH of the ground water as well as that of the surface water rises. The process is regarded as continuous, although conditions might have been more favorable at times in the past.

The geochemistry of iron and manganese at low temperatures is fairly well understood, and the following brief qualitative discussion is based on articles by Krumbein and Garrels (1952), Krauskopf (1957), and Hem and Cropper (1959). Water enters the outcrop area of the Tomstown Dolomite either as rainwater or as runoff in the streams that drain the Blue Ridge. Even after percolating through the soil, rainwater has relatively low alkalinity and a pH of about 5–6, and except for dissolved carbon dioxide, it is nearly pure. The same is true of the stream waters that enter the area

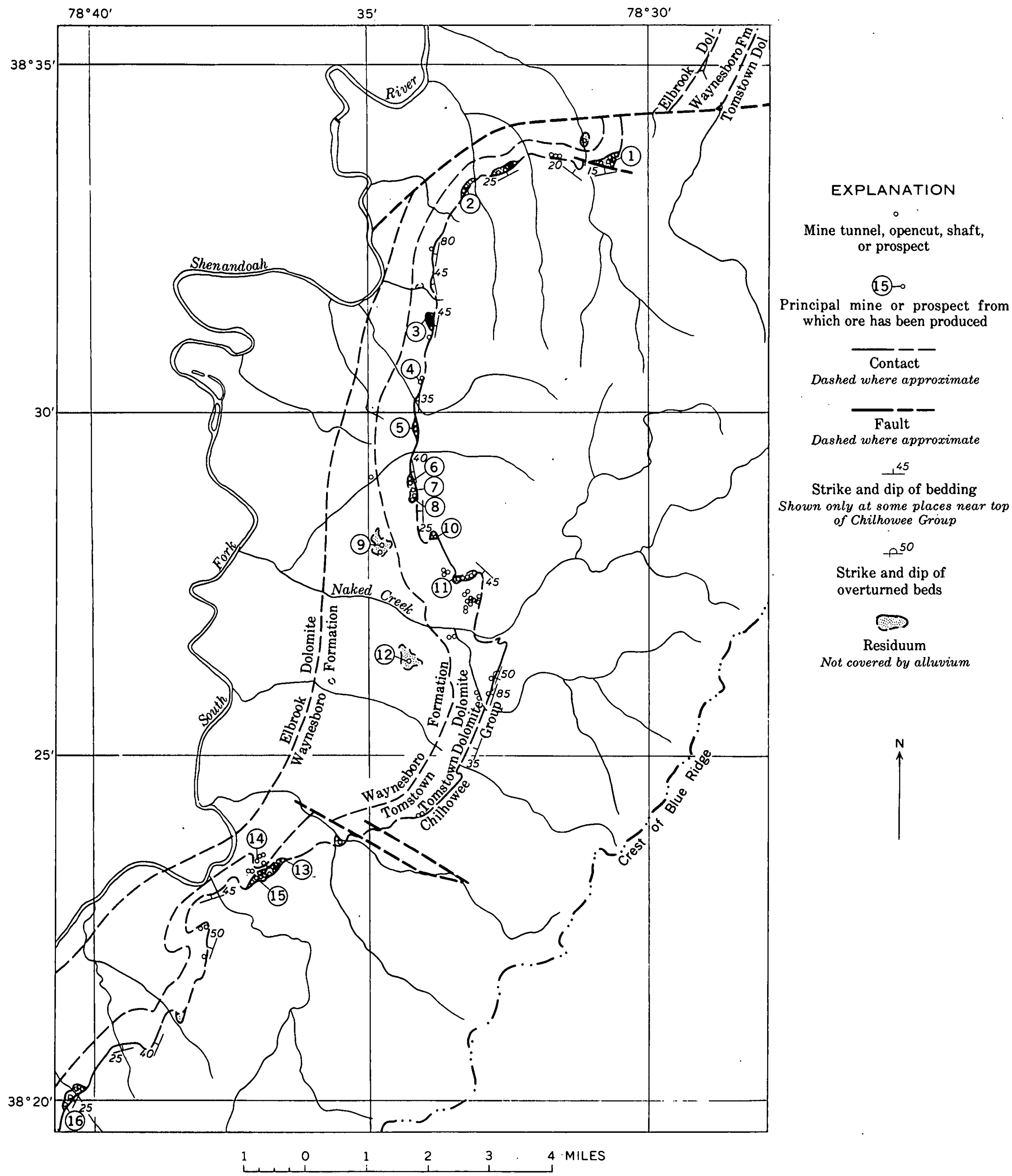

FIGURE 40.—Distribution of deposits of supergene oxides in relation to the geologic contacts between the Waynesboro Formation, Tomstown Dolomite, and Chilhowee Group, as shown by the abandoned mines and prospects in the Elkton area. (Redrawn from P. B. King (1950, pl. 1).

from the Blue Ridge, where the rocks are predominantly quartzite and nearly inert chemically. These waters dissolve the calcium and magnesium carbonates of the Tomstown, carry it away by percolation through the ground, and eventually leave the area through the main streams.

The alkalinity and the pH of the water are increased as it reacts with the limestone and dolomite. Under the prevailing Eh conditions, the iron and manganese are oxidized and remain, with other impurities, in the residuum though they may be moved locally for short distances. As indicated by the diagrams and discussions of Krauskopf (1957), the oxides and hydroxides of both iron and manganese become less soluble as pH increases. Solubility also decreases with increasing Eh (Krauskopf, 1957, figs. 1–4). According to Hem and Cropper (1959, p. 10), water having an Eh of 0.3 and a pH of 5 could contain ferrous iron in solution in a concentration of 100 ppm. At a pH of 7 and an Eh of 0.3, however, the concentration would be less than 0.001 ppm. Thus, the waters having a pH of 5–6.5 and an Eh of 0.3–0.5 entering the alluvial area at the foot of the Blue Ridge would be capable of dissolving and transporting some iron, and as the manganese compounds are somewhat more soluble, the waters could also transport manganese. Water whose pH has been raised to 7 or greater and whose Eh has been lowered slightly because of the reaction with carbonate rocks or with constituents of the residuum is unable to transport these metals.

In the interstream areas, water probably percolates downward through the residuum. As the residuum and weathered bedrock extend in places to depths more than 100 feet below the streams, this percolation must extend below stream level, and the movement of water must be temporarily reversed so as to rise toward the streams in the area near the channels. Because of the chemical equilibrium changes that take place in this percolating water, some solution of iron and manganese probably occurs in the upper part of the residuum. Some may escape the area by running off the surface, but it is precipitated in the streambeds, from which it is eventually carried off mechanically. Some of it, however, is carried downward until a small change in pH causes the iron to be precipitated, thereby enriching the residuum.

In accord with this theory, the manganese of the supergene deposits is concentrated from disseminated material in the original rock, whatever its ultimate source. On the basis of the analyses reported by Rodgers (1945) from a dolomite in Tennessee similar to the Tomstown, ample manganese is present to provide large deposits of the oxide. Rodgers sampled 238 feet of rock in the Shady Dolomite and reported an average MnO content of 0.38 percent. If this content were concentrated only by removal of the other constituents, a bed of oxide 0.81 feet thick would result. This concentration under an acre of ground would yield nearly 7,000 tons of pure MnO. Further concentration might result from chemical migration as residuum is ultimately removed by erosion at the surface. The effective thickness of the manganese-rich rocks would be greatly increased if the rocks were steeply tilted or vertical.

Physical and chemical conditions at the foot of the Blue Ridge are discussed in the following paragraphs to determine whether the conditions that exist at present in that area are consistent with the foregoing concept.

SOLUTION OF CARBONATE ROCKS

Ample evidence indicates that solution is going on under the alluvium and residuum at the foot of the Blue Ridge. One kind of evidence is gained from the analyses of stream waters. A sample of water collected by the writer from Canada Run, where it leaves the quartzite area of the Blue Ridge and enters the outcrop belt of Tomstown Dolomite (pl. 3), showed the analysis in table 11. Water collected downstream in the South River at Waynesboro, which has a large drainage basin in the outcrop area of carbonate rock, has more than 10 times the normal concentration of dissolved solids. A sample of water collected from a limestone sink in the alluvial area had a pH of 8.0, which probably indicates a high content of dissolved solids (p. 75).

TABLE 11.—*Comparison of an analysis of a water sample from a stream above the limestone area with that of a sample from a stream below the limestone area*

[Except for pH, quantities are expressed in parts per million]

Constituent	Canada Run [1]	South River at Waynesboro [2]
Silica	7.2	7.7
Aluminum	.1	----------
Calcium	.8	24
Magnesium	.4	15
Sodium and potassium	1.1	3.2
Bicarbonate (HCO_3)	5.8	136.0
Sulfate (SO_4)	.6	12
Dissolved solids (residue on evaporation at 180°C)	14.0	160.0
Alkalinity as $CaCO_3$	4.8	122.0
pH	6.4	7.5

[1] Location shown on pl. 3; drainage basin in quartzite; water collected by J. T. Hack on July 18, 1958; analysis by J. W. Barnhart and W. B. Hurlburt.
[2] Location shown on pl. 3; reported in Schroeder and Kapustka, 1957, p. 27; collected Oct. 6, 1952.

More striking evidence of solution is the occurrence of many sinks in the gravels overlying residuum, many of which are filled with water. The sinks are abundant in the alluvial area at the foot of the Blue Ridge (pl. 3),

FIGURE 41.—Exposure in wall of clay pit at the foot of the Blue Ridge at Cold Spring clay pit southeast of Greenville, Va. The dark layer in the center of the photograph and the lighter area in the upper left are a colluvial deposit consisting of angular blocks of quartzite. The light-colored layer beneath the dark-colored layer is clay. The talus in the foreground is derived from the colluvial layer and consists of quartzite blocks.

where depth to bedrock in places may be as much as 100 feet. The sinks are most abundant on terraces but also occur beneath the flood plains of small streams or across their channels. This relation indicates that solution is now occurring, or the sinks would otherwise be filled with alluvium. In places, subsidence is suggested by the occurrence of multiple channels and by swamps along the streams.

ENTRAPMENT OF THE RESIDUUM

The conditions at the foot of the Blue Ridge favor the loss of magnesium and calcium carbonate by solution but provide a trap for the large quantities of silt and clay, which are also constituents of the bedrock. The occurrence, formation, and preservation of this residuum are discussed on pages 44–49 but are briefly reviewed here with respect to the occurrence of manganese. As shown on page 12, both the Tomstown and Waynesboro formations contain silt and clay, in addition to carbonates, and the Tomstown alone in places probably contains enough to furnish deposits dozens of feet thick. Quartzite cobbles shed either as talus or alluvium from the hills of the Blue Ridge form an almost complete cover over the Tomstown and Waynesboro outcrop belt (fig. 41). Uncovered areas of residuum exist and form islands surrounded by channels and terraces composed of quartzite cobbles. The cobbles form an effective local base level that inhibits the erosion of the island of residuum.

It should be emphasized that the cover of alluvial cobbles and the residuum beneath the cover are a part of an erosional system in equilibrium in which the cobbles and the residuum are being removed from the inter-

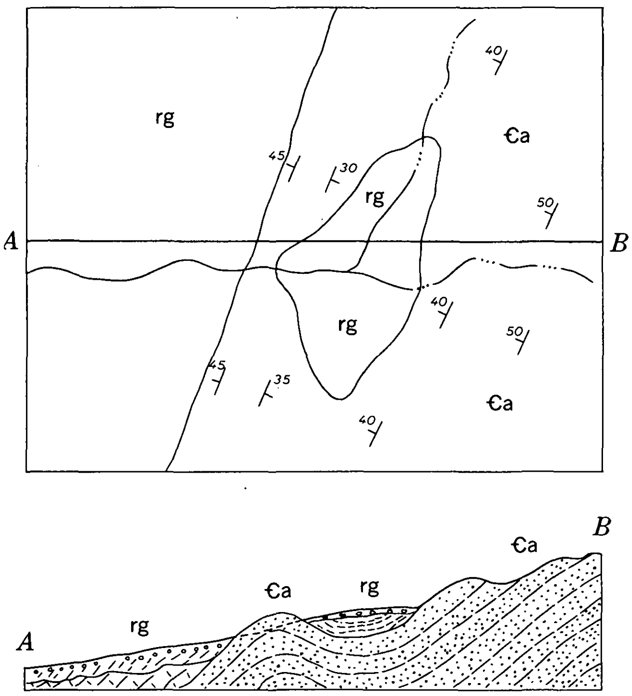

FIGURE 42.—A hypothetical erosional trap formed by a synclinal structure. rg, residuum derived from carbonate rock and covered in places by quartzite cobbles; Ꞓa, Antietam Quartzite.

stream divides at an overall rate equal to the rate of solution of the carbonate rocks below. The residuum is being added to from beneath and is being eroded from the oldest and most dissected of the terrace deposits. Were it not for the gravel, the relief of the land surface on the carbonate rocks would be much less, and the residuum would be thin or lacking. Evidence of this may be seen north of Front Royal where the Shenandoah River crosses the Tomstown Dolomite and erodes rocks of the Chilhowee Group. The river receives the debris eroded from the Blue Ridge directly and carries it off; therefore, the debris cannot blanket the silt and clay on the other side of the river. North of Front Royal, therefore, the relief on the Tomstown Dolomite is low (pl. 2), no manganese deposits occur (fig. 38), and although residuum is protected in places by terrace gravels of the main river, it is thin, and exposures of fresh rock occur in many ravines that lead down to the river. Near Harpers Ferry, where the river is still farther east of the belt of Tomstown Dolomite, terrace gravels are lacking entirely, the residuum on the Tomstown is quite thin, many bedrock exposures occur, and the rock is quarried for industrial purposes at several places.

Structural traps that preserve considerable thicknesses of residuum are also important in controlling the formation of ore deposits. The commonest of these traps are gentle synclinal structures along the margin of the Blue Ridge, as exemplified by the conditions at the Crimora mine (Stose and others, 1919, p. 83–95) and the St. Marys River tract (Knechtel, 1943, p. 190–194). At both of these localities, the manganese ore is in residuum that rests directly on the Antietam Quartzite. The quartzite is folded in a shallow basin or trough whose axis is transverse to the drainage. A hypothetical diagram of such a situation is shown in figure 42. The drainage from the synclinal basin passes over the quartzite in narrow ravines commonly floored with cobbles and boulders. This drainage exit forms a sort of sill behind which is trapped the residuum derived from the carbonate rocks that originally occupied the basin.

In terms of the equilibrium concept, the trap is only temporary. The residuum will eventually disappear as downward erosion continues and stream profiles in the basin are lowered below the original carbonate horizon. The residuum is now in the basin because it is mantled by the cobbles in the stream channel and flood plain and is protected, therefore, from mechanical erosion. It is not, however, protected from solution by ground water that circulates in the basin. This kind of trap probably is ideal for the accumulation of residual weathering products and especially the oxides of iron and manganese. The Eh and pH of the circulating waters can be expected to be such that the escape of iron and manganese in the streams would be very slow.

CHEMICAL EQUILIBRIUM IN THE RESIDUUM

The equilibrium theory requires that the ground waters circulating in the residuum or in the rocks beneath favor the accumulation and preservation of the metals as oxides and hydroxides. The metals must migrate in solution in order to concentrate in pods, lenses, veins, and nodules in the clay. On the other hand, as stated on page 72, the pH of the waters leaving the area must be high enough to prevent the escape of the metal in solution.

The hydrogen-ion concentration and redox potential of stream waters in the Shenandoah Valley have been studied by the writer with the help of Dorothy Carroll. Sampling trips were made in July 1958, during a period when the streams were low, and in April 1959, when the streams were high. A Beckman pH meter having platinum, calomel, and glass electrodes was used so that Eh could be measured as well as pH. The measurements were made immediately after collecting the sample.

The Eh measurements were probably not significant for the present purpose. The total range of Eh was between 0.375 and 0.505 volts. This is the range stated by Hem and Cropper (1959, p. 10) to characterize the waters of most natural streams. The temperature of the water samples ranged from 32°C. to 14°C., a fact that probably accounts for much of the range in Eh, as temperature and Eh vary inversely.

The range in pH, however, was from 5.5 to 8.9 and is significant, for the values are related to the geologic conditions in the drainage basins of the streams measured (Hack, 1960b). In streams draining quartzite areas the values obtained at the low-water period ranged from 5.5 to 6.4. In streams draining areas that contain outcrops of granodiorite and greenstone, the pH was between 7.1 and 7.8, and in limestone and dolomite areas the range was from 7.6 to 8.3. At the high-water period in April, the values tended to have a smaller range. The differences are related to the content of dissolved solids in the water as determined by the chemical activity of the rocks through which the stream flows. This can be verified by examination of the water analyses published by Schroeder and Kapustka (1957) and those obtained by the writer (fig. 43). Water in streams in the Potomac basin has a pH roughly proportional to the logarithm of the content of dissolved solids. The samples from these three geologic environments fall into distinct groups; the nearly pure water collected in streams issuing from the quartzite areas has the lowest pH.

Extreme differences were found in pH measurements made near the manganese deposits during the low period

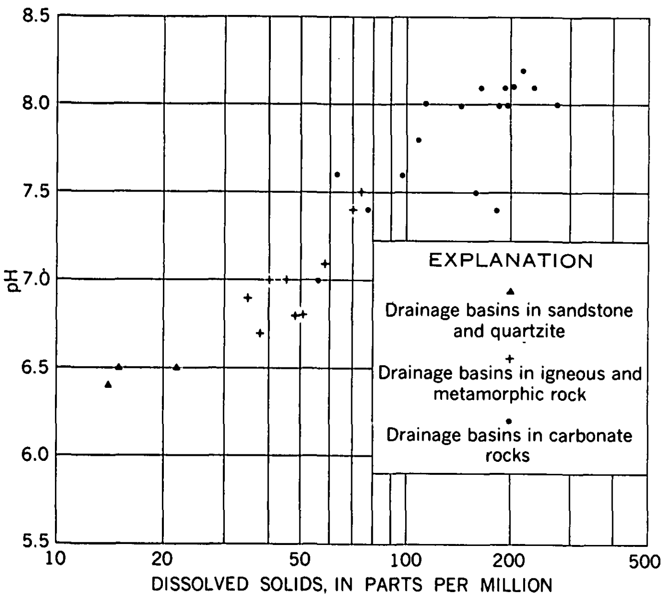

FIGURE 43.—Comparison of pH and content of dissolved solids in stream waters in the Potomac Basin. Data are from Schroeder and Kapustka (1957, p. 26–29). Analyses showing a high content of sulfate are not included because of the probability that they are artificially contaminated. The data include five samples collected by the writer and analyzed by J. W. Barnhart and W. B. Hurlburt.

of July 1958; measurements ranged from 5.5 above the manganiferous residuum to 8.0 further downstream below the area where the water passes through alluvium and residuum (table 12). In April 1959, when the flow was higher, the differences were less marked; nevertheless pH increased downstream from the quartzite. These measurements support the concept that present conditions are suitable for the concentration of manganese deposits. The values of pH, with the Eh conditions that prevail, lie within the critical range for both iron and manganese compounds. Both of these metals must be in the form of oxides rather than carbonates if they are to be in equilibrium with waters such as occur in the streams of the region (Krauskopf, 1957). The stability fields and solubility relations are such that with an oxidation potential of 0.3–0.5, oxides of iron would be slightly soluble under conditions of low pH but nearly insoluble with a pH as high as 7.5–8.5. At depth, beneath the residuum where the materials are not already oxidized, the manganese and iron may exist in the form of the carbonate because of a lower Eh. Some transportation and circulation of iron and manganese are therefore permitted by the conditions that prevail both in the bedrock and in the upper layers of the residuum, which are more leached and more acid.

TABLE 12.—*pH of stream-water samples collected above and below outcrop belt of Tomstown Dolomite and Waynesboro Formation*

[Samples measured in field with Beckman pH meter]

Locality	Date of sampling	
	July 1958	April 1959
Collected where streams enter outcrop area of Tomstown Dolomite		
Loves Run	5.5	5.9
Canada Run	6.0	6.0
Paine Run at Shenandoah Park boundary	6.4	6.1
Collected below outcrow area of Tomstown Dolomite and Waynesboro Formation		
Sink near Sherando, Va	8.0	6.4
Paine Run at Harriston, Va	7.0	6.4

The migration of some iron and manganese in the area of the ore deposits is indicated by deposits of metallic oxides on cobbles in some stream beds. Black coatings may be seen on cobbles in Canada Run downstream from the Kennedy mine. The coatings apparently form only at periods of low flow, for they are observed on cobbles that line the bottom of the channel but not on the sides or tops of the larger boulders. Deposition of manganese minerals in the small rapids of a stream near Harpers Ferry has been described by

Howe (1932, p. 62) and is associated with the action of algae.

Composition of clay minerals in the residuum is also most likely related to the equilibrium conditions that prevail. Samples of clay were collected in June 1957 by John Hathaway accompanied by Dorothy Carroll, M. M. Knechtel, and the writer. Hathaway sampled clay from the Vesuvius and Kennedy mines and Cold Spring clay pit (fig. 38, locs. 1, 2, 3). The clay in all these places was found by X-ray methods to be rich in iron and alumina. Kaolin constituted half of the clay content in each sample, and goethite, 10–20 percent (Hathaway, written commun., 1958). Knechtel (oral commun., 1957) stated that some bauxite was obtained from the Cold Spring clay pit during operations there. The high alumina and iron content of these clays is expected in the environment that now exists in this area. As shown by Carroll and Starkey (1959), under conditions of high pH such as must occur at the surface of the carbonate rocks or downstream in the residuum derived from them and in waters containing oxygen, silica is soluble but iron and alumina are insoluble, and enrichment in these substances might form kaolinite, goethite, and even bauxite.

DISCUSSION OF THEORIES OF ORIGIN

The writer believes the evidence in the field shows that present conditions favor the concentration of oxides and hydroxides of iron and manganese in the clays leached from carbonate rocks at the foot of the Blue Ridge, and it is therefore unnecessary to postulate concentration of the manganese on a peneplain surface. A considerably lower relief than now exists might indeed favor the preservation of the residuum, but the conditions that exist today also favor preservation in spite of the high relief. Furthermore, the same processes would operate under conditions of even higher relief to provide traps for the clay and its contained oxides. The existence of a warmer climate does not appear to be necessary. Limestone and dolomite are generally thought to be more soluble at low temperatures than they are at high temperatures. A different temperature might alter the rate at which the process of concentration of the metals operates but should not prevent its operation.

Both Hewett (1917) and King (1950, p. 57) mentioned highly weathered alluvial gravel that contains manganese and underlies terrace gravels as stratigraphic evidence that the period of manganese deposition occurred at some time during the distant past, perhaps in the Tertiary. King found one occurrence of such gravel in the Elkton area at the Stanley mine. The present writer visited this mine but did not see any gravel because of the deterioration of the exposures. The deposit described by King contains nodules and masses of manganese oxide that have impregnated or replaced sandy sediment. Manganiferous gravel of this kind is evidence that manganese accumulated and was concentrated at times in the past in the gravels as well as in the residuum. The evidence proves only that some of the concentration antedated some of the older gravels. It does not prove that all the manganese deposition occurred at some remote time in the distant past. In fact, migration of manganese and redeposition in gravel deposits is going on now, as is evidenced by the coated cobbles in Canada Run.

IRON AND MANGANESE DEPOSITS AT OTHER GEOLOGIC HORIZONS

Deposits of supergene oxides occur both on the Beekmantown Dolomite and on the Helderberg Limestone (fig. 38). These deposits were described by Stose and Miser (1922) and Monroe (1942), and the reader is referred to these sources for a description of individual mines. The principles of concentration of the metals at these horizons are similar to those of the deposits at the Cambrian horizons, though the situation may be quite different. The Beekmantown Dolomite is a formation that locally contains iron-rich beds and massive chert beds. Presumably at places it also contains manganese in small concentration. John Hathaway and the writer collected clay from a solution tube in the Beekmantown Dolomite exposed in a quarry that proved to be nearly pure goethite (Hathaway, written commun. 1958). The Beekmantown residuum is characteristically dark red, probably an indication of its iron content. Therefore many sites along the outcrop belt of the Beekmantown are favorable for the concentration of metallic oxides. As described on page 45, chert forms a surface deposit and provides a trap that prevents the erosion of the clay and whatever metallic oxides have been concentrated in the clay.

Several large deposits of supergene oxides have been found in residuum that overlies the limestones of Devonian age. The few deposits examined by the writer are apparently preserved beneath the same kind of protective blanket that preserved the Cambrian deposits. Nearly all the deposits are stratigraphically close to sandstone beds such as Oriskany, Clinton, and Tuscarora, which provide a source of cobbles and boulders that blanket the outcrops of limestone residuum and prevent their erosion.

As shown by the map (fig. 38), the deposits related to Devonian bedrock are especially numerous in areas where the limestones are repeated in synclinal folds as on Massanutten Mountain.

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