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Agronomy Series

Geomorphology and Soils of the Northeastern United States and Pennsylvania:

A Series of Reprints

by

Edward J. Ciolkosz
and
Nelson C. Thurman

Agronomy Series Number 116

May 1992

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Chapter 2. GEOLOGY, PHYSIOGRAPHY, VEGETATION AND CLIMATE*

Edward J. Ciolkosz, Thomas W. Gardner, and John C. Sencindiver

The northeastern part of the United States encompasses a relatively small geographic area, about 62,232,000 hectares (153,779,000 acres) or about 7 percent of the United States area. Although relatively small, this area has a diverse array of geology, physiography, vegetation, and climate. The following sections of this chapter will briefly discuss these topics.

GEOLOGY

The geologic evolution of the northeastern United States is an extremely complex series of events (Cook et al., 1980; Eardley, 1951; King, 1969, 1977; King and Beikman, 1974; Isachsen, 1980). Most of these events can be best understood in terms of plate tectonics or, as it was called in its early stages of development, continental drift.

A complete and comprehensive discussion of this subject is available in the excellent summary texts by Wilson (1976) and Bird (1980). Briefly, the concept indicates that the earth's continental landmasses are not stationary, either with respect to a fixed reference point or with respect to each other. The continental areas are thought to be rigid blocks or plates (the lithosphere) of relatively light, sialic material (silica and aluminum-rich) that "float" and move over a denser mafic (iron and magnesium-rich), less rigid mantle material (the asthenosphere).

Figure 1 illustrates a general cross-section of the earth showing upwelling zones and relative motion of the plates. Material rises from the asthenosphere, cools and forms new lithosphere. The zone in which new lithosphere is created is an active "rift zone" and the Mid-Atlantic Ridge is a classic example of one such zone. Iceland is located astride the Mid-Atlantic Ridge and its pronounced volcanic activity results from its location on a rift zone. The lithosphere descends back into the asthenosphere along subduction zones in other areas of the globe, such as along the Pacific coast of North America. Numerous earthquakes and volcanic activity occur along these zones.

Plate tectonics provides a useful model to help explain the geologic history of the northeastern United States. For example, the periods of mountain building in the Northeast can be related directly to the collision of lithospheric plates. The sequential tectonic development of the Northeast is illustrated in Figures 2, 3, and 4. These figures should be referred to during the following discussion.

The geologic history of the northeastern United States begins with the Precambrian Grenville rocks (Figure 2). The major Grenville orogenic period ended about 950 million years ago. Igneous and metamorphic rocks of Grenville age are visible in the Adirondack Mountains of northern New York state and in the anticlinorial uplifts of the Green Mountains of Vermont, the Hudson and Jersey Highlands, the Reading Prong in eastern Pennsylvania,

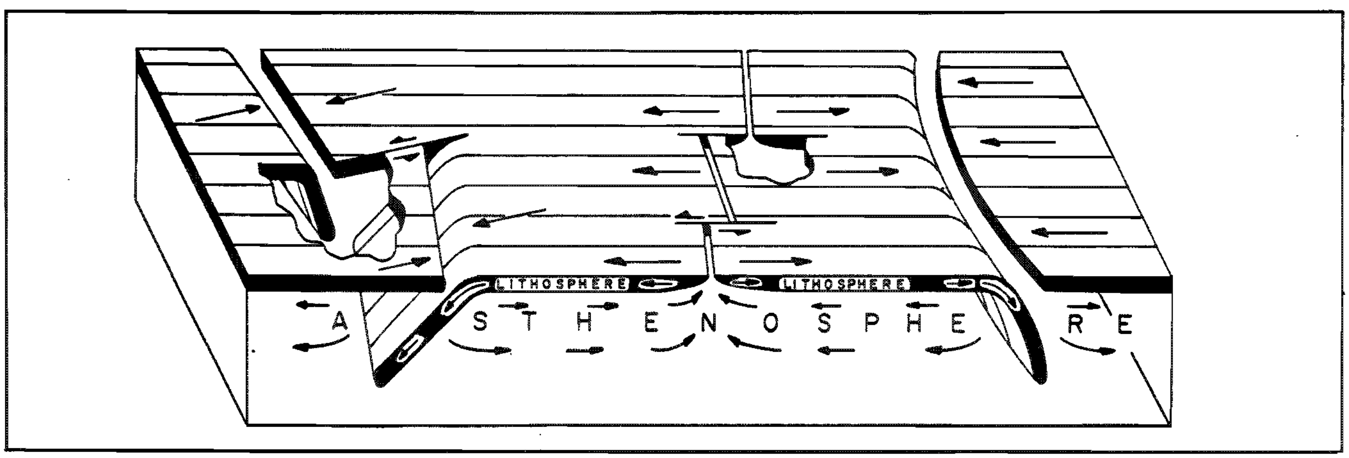

Figure 1. Generalized plate tectonic model illustrating the configuration, role, and sense of motion of the lithosphere and asthenosphere. Arrows on lithosphere indicate relative motion of adjoining blocks. Arrows in the asthenosphere show upwelling of new crust along a rift zone or active spreading center. From Isacks, Oliver and Sikes, Fig. 1, in Plate Tectonics, ed. J. M. Bird, 1980, reprinted with permission of the publisher.

 *Reprinted from R. L. Cunningham and E. J. Ciolkosz (editors) Soils of the Northeastern United States. PA Ag. Expt. Sta. Bull. 848. 1984.

and are present as basement rocks beneath Paleozoic sediments throughout the area farther west. In Maryland and Virginia, Grenville age and late Precambrian rocks (extending to about 600 million years ago) are exposed in the Blue Ridge and Piedmont Provinces. They consist of metamorphic schists, gneisses, and intrusive igneous rocks of granitic to gabbroic composition. The Blue Ridge is composed of resistant granitic and gneissic rocks with a minor component of metamorphosed volcanic rocks (greenstones) and stands topographically above the Piedmont. In the Piedmont, a significant amount of less resistant metamorphosed Paleozoic rocks is contained within the gneissic basement. The Baltimore gneiss domes in the Piedmont of Maryland are a good example of a metamorphic basement complex upon which later Precambrian and Paleozoic sediments were deposited and subsequently metamorphosed during later orogenic events. The metamorphic and igneous rocks of the Piedmont have very complex structures indicating numerous periods of deformation.

Together, the rocks of the Piedmont and Blue Ridge record a long history which probably included repeated epochs of sedimentation, deformation, metamorphism, and intrusion. Several areas of these rocks produce economic deposits of metals. An example is the Franklin Mine in northern New Jersey which has been a major zinc producer.

The next major event in the late Precambrian history (about 750 million years ago) of the Northeast was the splitting of the continent that contained the Grenville belt, producing first a rift zone and then the proto-Atlantic ocean basin, with North America on its northwest side (Figure 3a). On the margin of the North American continent, sediments accumulated in the Appalachian geosyncline. The geosyncline is made up of a miogeosyncline which forms the continental shelf and a eugeosyncline which forms the ocean floor itself (Figure 3b). Sediments accumulated in the basin throughout Cambrian and Lower Ordovician time producing a thick accumulation of fossiliferous marine limestones and

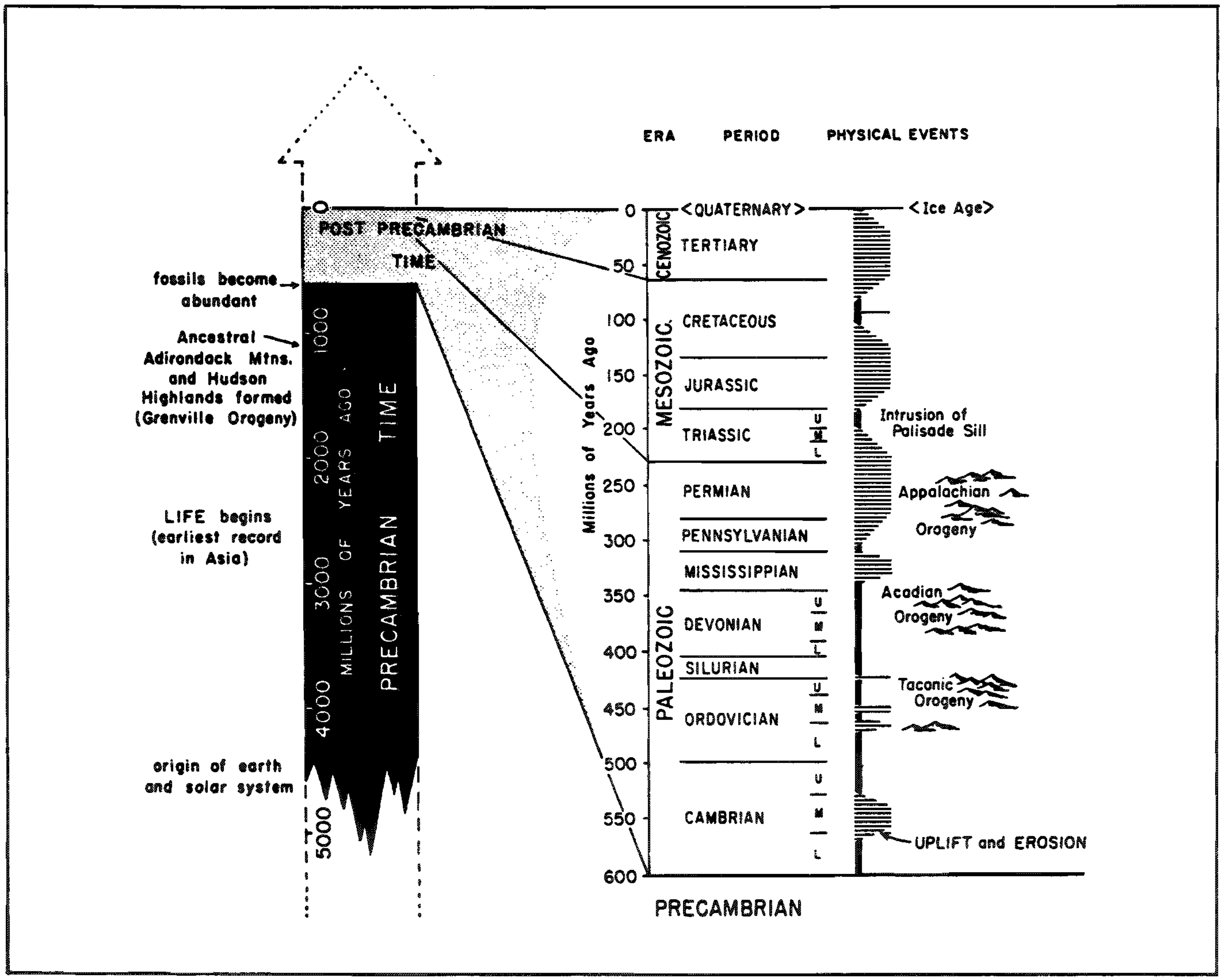

Figure 2. Geologic time scale showing major geologic events in the northeastern United States. From Boughton et al., 1966.

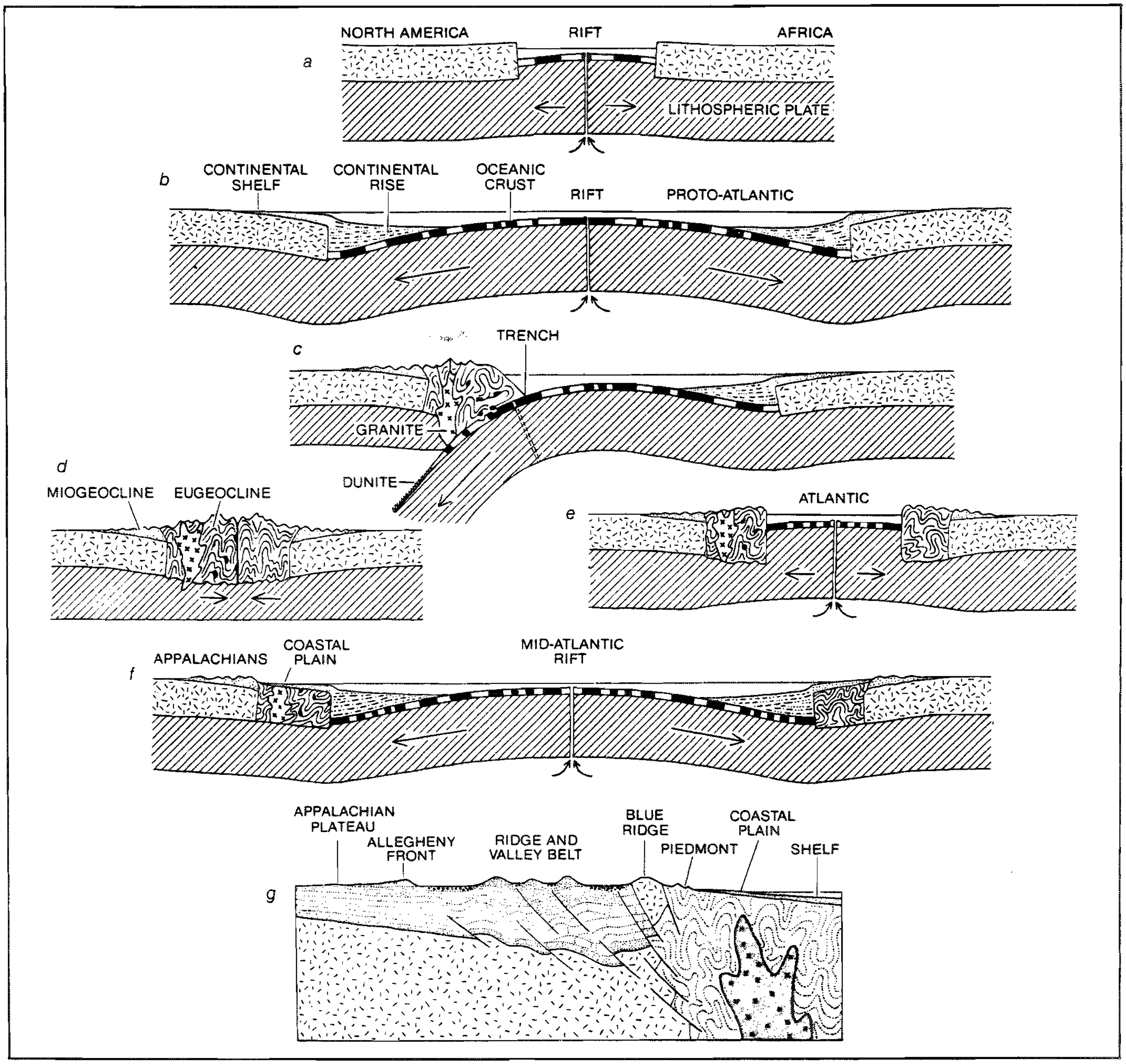

Figure 3. Mechanisms of crumpling that produced the Appalachian foldbelt based upon the hypothesis that the Atlantic Ocean has opened, closed, and reopened. In the late Precambrian (a), North America and Africa are split apart by a spreading rift, which inserts a new ocean basin. By the process of sea-floor spreading (b) the ancestral Atlantic Ocean opens. New oceanic crust is created as the plates on each side move apart. As the crust cools, its direction of magnetization takes the sign of the earth's magnetic field; the field periodically reverses, and the reversals are represented by the striped pattern. On the margin of each continent, sediments produce the geosynclinal couplet: miogeocline on the continental shelf, eugeocline on the ocean floor itself. The ancestral Atlantic now begins to close (c). The lithosphere breaks, forming a new plate boundary along the eastern portion of the present United States, and a trench is produced as the lithosphere descends into the earth's mantle and is resorbed. The consequent underthrusting collapses the eugeocline, creating the ancient Appalachians. The eugeocline is intruded with ascending magmas that create plutons of granite and volcanic mountains of andesite. The proto-Atlantic is now fully closed (d). The opposing continental masses, each carrying a geosyncline couplet, are sutured together, leaving only a transform fault (vertical black line). Sediments eroded from the mountain foldbelt create deltas and fluvial deposits collectively called molasse. North America and Africa were apparently joined in this way between 350 million and 225 million years ago (Mississippian through Permean periods). About 180 million years ago (e) the present Atlantic reopened near the old suture line (Triassic and Jurassic). Today (f) the central North Atlantic is opening at the rate of 3 centimeters per year, creating new geosynclines. An enlargement (g) of the left portion of (f) shows how the lithology and structure control the boundaries and position of the major physiographic provinces. From Wilson, 1976, reprinted with permission of the publisher.

shales, which pinch out against the stable interior platform to the west.

In Middle Ordovician time the ocean began to close and deformation occurred (Figure 3c). The deformation, called the Taconic orogeny (Figure 2), was expressed as folding, thrust faulting, uplift, gravity sliding, metamorphism, and granodioritic intrusion. The areas affected by mountain building were northern Maine, western New England and adjacent New York, northern New Jersey, and southeastern Pennsylvania. To the west in a shallow inland sea, a great delta (Queenston delta) was built reaching beyond Niagara Falls. Farther southeast in Maryland and Virginia, however, deposition was probably never interrupted. Thousands of feet of shallow-water limestones were deposited on a slowly subsiding continental shelf. These fossiliferous limestones and dolomites later would fold and erode to form much of the Great Valley of Pennsylvania, Maryland, and Virginia.

As the mountains wore down, the sea readvanced, reestablishing the continental shelf and slope in New England, probably somewhat farther east than before. During Silurian time in Pennsylvania, the Bloomsburg delta was constructed. The Bloomsburg delta is well known for its classic redbed sequence.

The Tuscarora Sandstone, which formed many prominent ridges in the Ridge and Valley area from Pennsylvania to Virginia, was deposited at the base of this thick deltaic sequence. It records a period of major tectonic uplift and clastic sandstone deposition. It is predominantly riverine in New Jersey and Pennsylvania, but becomes thin and more marine in Virginia. As with the Ordovician Taconic orogeny, this Silurian mountain-building episode was most active in the Northeast and decreased in intensity toward Virginia.

In Middle Devonian time, all of New England and the edge of New York were intensely deformed by the Acadian orogeny (Figure 2). There was extensive metamorphism and granitic intrusion, and the sea retreated. Again mountains were produced and the Catskill delta was built to the west beyond the limits of the deformed area. The sea returned to western Pennsylvania, Virginia, and Maryland for a short time during the Mississippian period; but deformation was renewed in a belt passing from southernmost New England and southwestward through eastern New York, New Jersey, easternmost Pennsylvania, and Delaware. The nearshore portion of these deltaic deposits, in eastern New York and Pennsylvania, is composed of coarse grained sandstones and conglomerates. These quartz-rich rocks are very resistant to erosion and are the cause of the high relief in the Catskill Mountains and Pocono Plateau. In Maryland and Virginia this period was marked by pronounced intrusion of deep-seated igneous rocks and metamorphism in the Piedmont. West of the coarse deltaic sediments, deposition of finegrained sandstone, shales, and limestones continued throughout the Devonian time.

During the Pennsylvanian period, sedimentation rates varied markedly on the deltas. Periodic marine transgressions and regressions occurred and extensive peat swamps developed on the deltaic deposits. Much of the change in sea level that caused the transgressions was due to glaciation on adjacent land masses. The Carboniferous period takes its name from the numerous coal beds that formed from the peat swamps. Erosion has removed much of the Pennsylvanian record, but large areas of coal still occur in Pennsylvania, West Virginia, western Virginia, and Maryland, and to a lesser extent in Massachusetts and Rhode Island.

A major deformation occurred about the end of the Pennsylvanian time and in the Permian (Appalachian or Allegheny orogeny), producing the "typical Appalachian" folds of Pennsylvania, Virginia, and Maryland. The Allegheny orogeny also caused folding, metamorphism, and granite intrusion in New England. At this time the proto-Atlantic was completely closed (Figure 3d) by the collision of the North American, European, and African continents which were then welded (sutured) together to form the supercontinent, Pangea (Figure 4). This continental collision caused the igneous and metamorphic rocks of the Blue Ridge to be thrust westward up and over the sedimentary rocks of the Valley and Ridge. This area currently is called the Eastern Overthrust Belt and is a prime target for oil and gas exploration today.

The next well-documented event was one of mild warping and faulting accompanied by fluvial, alluvial fan, dark lacustrine sedimentation, and some volcanism in rift valleys from the Connecticut River basin in Massachusetts and Connecticut to the Culpepper and Richmond basins in Virginia (Figure 5). Tension in the earth's crust that resulted from the pulling apart of the North American, African, and European continents created the faults which bound these basins. Volcanic rocks were intruded along these tensional features. Though these deposits have long been called "Triassic," they are now known to range in age through Late Triassic and Early Jurassic.

At the same time or, more likely a little later in the Jurassic, the present Atlantic Ocean began to open (Figures 3e and 4). A large group of volcanic calderas, centered in the present area of the White Mountains of New Hampshire but with outliers in adjacent states, also were associated with the opening of the Atlantic. The associated igneous rocks, both intrusive and extrusive, are markedly alkalic. Marine waters eventually transgressed the eastern margin of the North American continent, and the present continental shelf, slope, and rise developed. The oldest dated rocks at the bottom of the pile of shelf sediments are Jurassic, and are a thick sequence of evaporated beds of salt. However, the oldest visible rocks on the present coastal plain are lower Cretaceous in age. Southern New Jersey, Delaware, eastern Maryland and eastern Virginia are underlain by undeformed, gently seaward-dipping, marine Cretaceous and Cenozoic shelf deposits (Figures 3f, 3g, and 7). In Cretaceous and early Tertiary (Eocene) time the present Coastal Plain area was

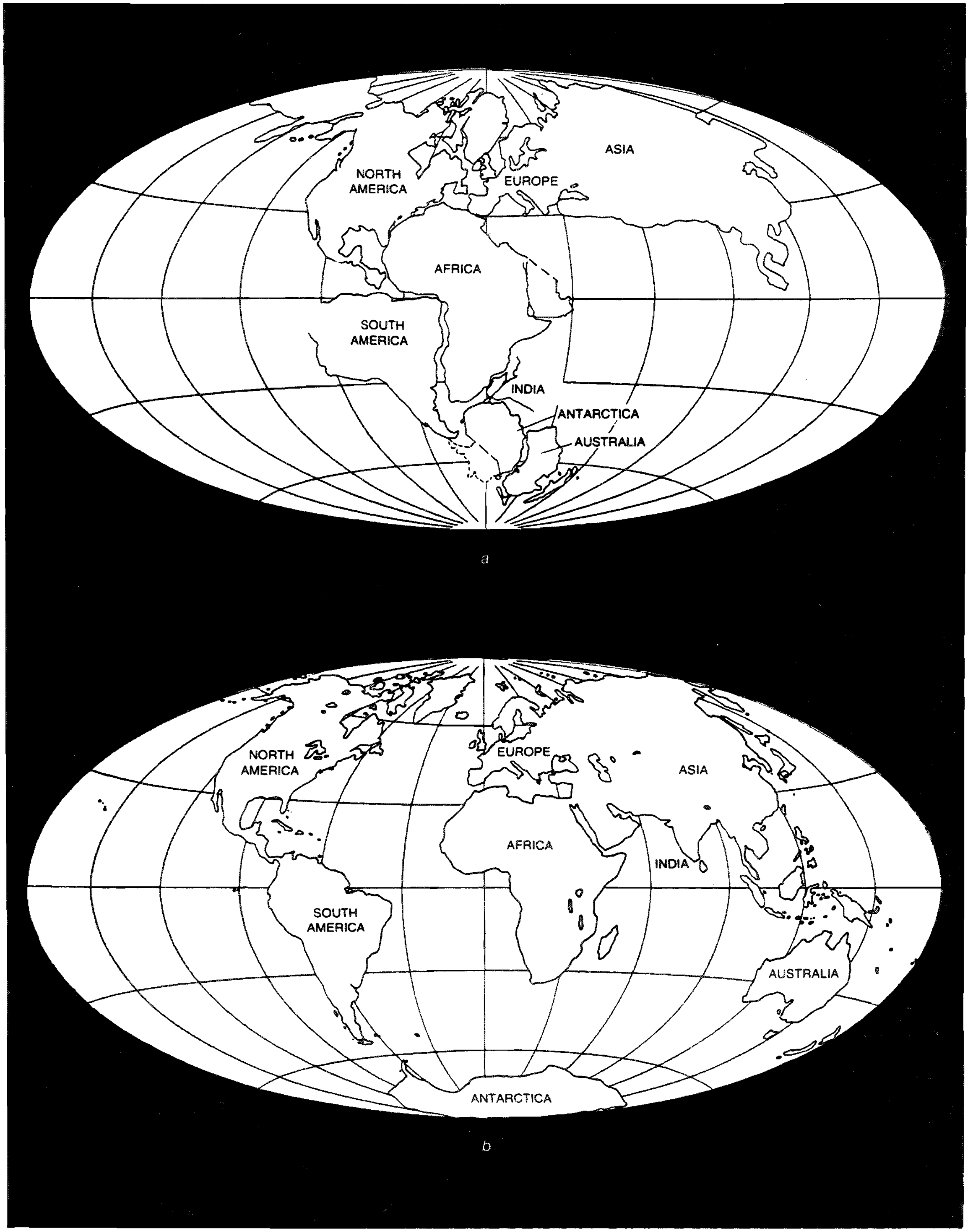

Figure 4. An ancient supercontinent called Pangea incorporated all the earth's large land masses at the end of the Paleozoic (a). For comparison, a map of the world as it appears today is shown (b). From Wilson, 1976, reprinted with permission of the publisher.

submerged and sediments accumulated. By late Tertiary (Miocene) time, about 20 million years ago, the Coastal Plain had been uplifted and was half its present width, and by the late Quaternary (Holocene) time the present shoreline had developed. Within the Coastal Plain many old beach ridges still can be recognized as long linear features with well-drained sandy soils, while the old offshore, finer-grained marine units tend to form valleys which control the present day drainage.

With the exception of the Coastal Plain, the northeast region was being eroded in late Mesozoic and Cenozoic times; yet the Appalachian region today stands moderately high and must have been differentially uplifted. Whether the region was ever completely worn down to a peneplain and then rejuventated (and, if so, how often) and whether the uplift was broadly continuous or spasmodic, are matters of debate.

The last major geologic event in the northeast region was the Pleistocene (early and middle Quaternary) continental glaciation. Much of the Pleistocene geology is summarized by Flint (1971). Continental ice covered a large part of the region and only central and southern New Jersey, the bulk of Pennsylvania, Virginia, and West Virginia were spared (Figure 5). The ice advanced and retreated many times, leaving terminal moraines which record two major Wisconsinan (late Pleistocene) advances and at least two older pre-Wisconsinan advances. The falling and rising of sea level, that accompanied the advance and retreat of the glaciers, alternately exposed the continental shelf and flooded the coastal areas from New Jersey to Virginia. The last ice retreat is well recorded by glacial till which was spread unevenly over the glaciated area, disrupting the drainage and producing the many lakes of New England and New York. Near the retreating ice, stratified drift was deposited in rivers, lakes, and the sea. Much of the constructional topography still is evident along the New England coastline where end moraines control the configuration of Long Island and Cape Cod (Figure 6). For a time during the Pleistocene, the sea covered much of coastal Maine and penetrated the St.

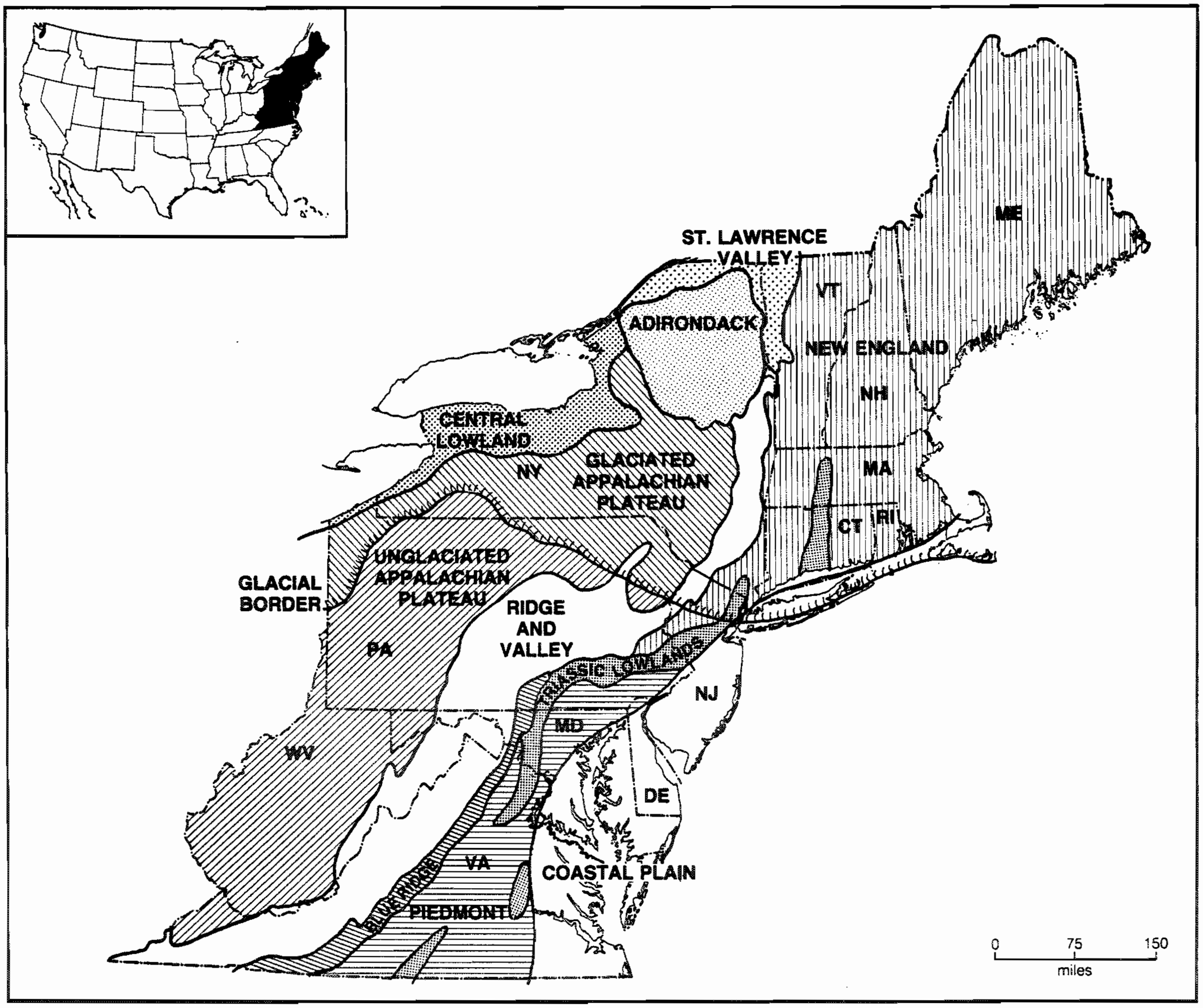

Figure 5. Physiographic provinces of the northeastern United States.

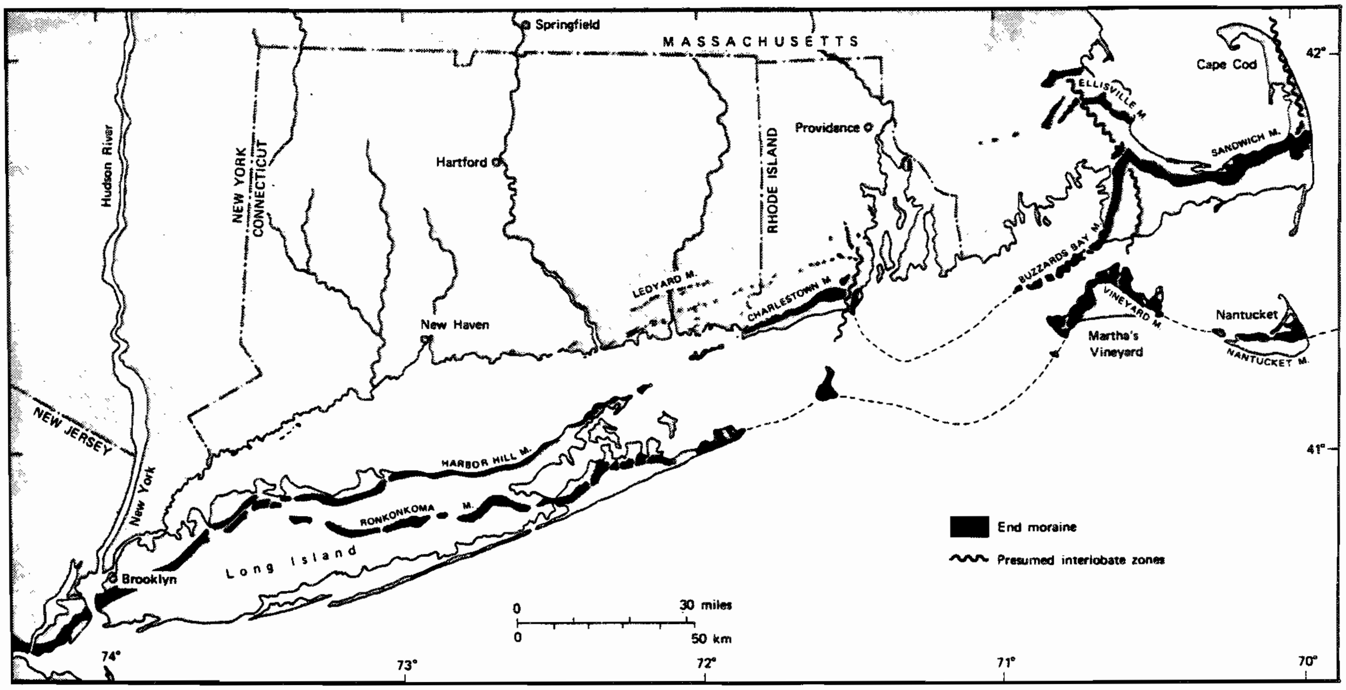

Figure 6. Map showing known end moraines between Hudson River and Cape Cod. From Flint, Fig. 22.2, 1971, reprinted with permission of the publisher.

Lawrence Valley to Lake Champlain and northern New York state, until glacial rebound caused its recession.

South of the glacial border, significant changes also were occurring during the Pleistocene. Periglacial conditions extended many hundreds of miles south of the glacial ice, particularly at higher elevations. Features produced by periglacial processes, such as patterned ground, involutions, ice wedge casts, grezes litees, and boulder fields, have been reported (Clark, 1968; Ciolkosz, 1978; Ciolkosz et al., 1979) to have formed during this time. Although these features are of interest, the occurrence of solifluction colluvial deposits over extensive areas (25 percent of an average county in the ridge and valley area of Pennsylvania) indicate that the surfaces of large areas were modified greatly during this time. In addition, these deposits indicate that a considerable amount of landscape modification occurred during the Pleistocene.

PHYSIOGRAPHY

When the North American, European, and African continents suffered the catastrophic collision about 250 million years ago, described in the previous section, sediments that had accumulated in the ancestral or proto-Atlantic were squeezed, folded, and thrust-faulted to create the Appalachian Mountain backbone of eastern North America. During the long intervening period up to the present, fluvial erosion and to a lesser extent glacial erosion, have etched out the zones of highly fractured rocks, faults, thrusts, and less resistant lithologies and have left the more resistant lithologies to stand as the accordant summits and monadnocks of the central and northern Appalachians.

The northeastern region has a wide variety of topographic forms. Some areas have similar characteristics or topography that tend to be elongated and subparallel to the Atlantic seaboard (Figure 5) and confined to zones of distinct structure and lithologic association (Figures 3g and 7). As a result of this unique relationship between topography, structure, and lithology, these regions have become known as physiographic provinces.

In the following sections a general discussion of the physiography (after Harrison, 1970; Hunt, 1967, 1974; Raisz, 1957; Thornbury, 1965; Denny, 1981) of the Northeast will be presented. The discussion will be limited to the features of these physiographic provinces found in the northeast region. Information on the character of these provinces in other regions should be obtained from the cited references.

Appalachian Plateau

The stable, interior region of the continent represented by the Appalachian Plateau Province extends from southern New York southwestward through West Virginia. The region is an elevated plateau that ranges from about 304 meters (1,000 feet) in elevation at its western edge to at places more than 912 meters (3,000 feet) at its eastern edge. The thickness of sedimentary rocks (predominantly interbedded sandstones, shales, conglomerates, and some coals) of Paleozoic age increases markedly from 3,950 meters (13,000 feet) near the western boundary to over 9,120 meters (30,000 feet) in central Pennsylvania. The plateau consists of a series of gently dipping rock units

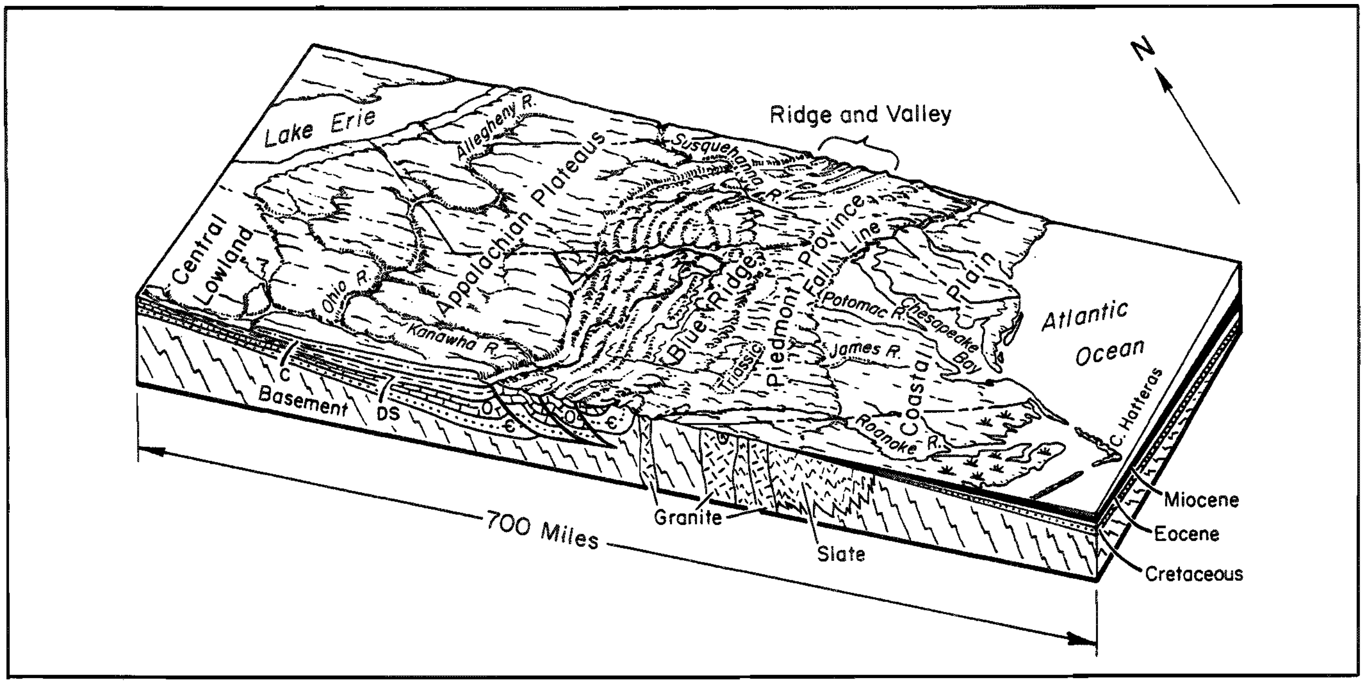

Figure 7. Structural and lithologic framework of the physiographic provinces in the central part of the northeastern United States. C, Cambrian; O, Ordovician; DS, Devonian Silurian; C, Carboniferous (Mississippian and Pennsylvanian). From Hunt, Fig. 11.3, 1974, reprinted with permission of the publisher.

arranged in the shape of a saucer. The deepest part of this structural basin, the Dunkard Basin, is centered in southwestern Pennsylvania and West Virginia. The rocks of the central basin mainly are thick red, montmorillonitic shales and interbedded sandstones that produce some of the most landslide-susceptible slopes in the Appalachians. In the eastern part of the plateau in southern New York, central Pennsylvania, and northern West Virginia, the formations are compressed into very gentle, northeast-trending folds which are the site of numerous oil and gas fields. These folds are a belt of transition from the plateau into the more intensely folded Ridge and Valley Province.

The plateau is bounded on all sides by outfacing escarpments which reflect its regional synclinal basin structure. The eastward facing escarpment, the Allegheny Front, forms the boundary between the Plateau and the Ridge and Valley area to the east. This feature forms one of the most persistent and striking topographic breaks in the United States (Hunt, 1967). The plateau has undergone considerable dissection, with the dissection generally being greater in the eastern area along faults, fractures, and zones of weakness. Although the greatest dissection is in the eastern area, along major rivers or streams in the central and western areas, the plateau also is deeply dissected.

Dendritic drainage patterns resembling the branches of a tree are typical of flat-lying sedimentary rocks and are well developed on the Appalachian Plateau. The drainage pattern is quite distinct from the trellis or rectangular pattern that has developed in the Ridge and Valley Province. In fact, maps showing only the drainage patterns of those two areas would provide sufficient information to distinguish the two provinces. The northern part of the Plateau in New York and Pennsylvania has been glaciated. The northern and western part of the glaciated plateau show the greatest amount of glacial erosion, deposition of till and valley filling, while the southcentral and southeastern areas show much less topographic modification.

Ridge and Valley Province

The folded and thrust-faulted Appalachian structural zone is the geomorphic Ridge and Valley Province. This zone is located just to the east of the Appalachian Plateau and extends from northern New York through Virginia. Its physiography is distinct from that of the plateau. It is the region of flat-topped, parallel or subparallel ridges and valleys that are carved out of anticlines, synclines, and thrust sheets. Local relief varies from 305 to 610 meters (1000 to 2000 feet). It often is termed the newer Appalachians, in contrast to the older Appalachians which include the Piedmont and Blue Ridge Provinces.

Most rocks of the Ridge and Valley Province are Paleozoic sediments ranging in age from Cambrian to Pennsylvanian. Their resistance to erosion varies greatly and has a major effect upon the topography. The Ridge and Valley area contains two levels of topography. The ridge crests appear to be accordant (at the same height), and are composed primarily of resistant sandstone and conglomerate bedrock. The valleys are underlain by less resistant shales and limestone. For example, the broad lowland of the Great Valley in Pennsylvania is composed of nonresistant Cambrian and Ordovician limestone and shales which characteristically produce fertile soils. The even-crested ridgetops and the uplands of the valley floors are thought to represent old erosional surfaces called peneplains (Johnson, 1931). It also has been proposed that different rock types produce surfaces of different elevations simply because of their differences in resistance, and therefore they do not record former peneplains (Hack, 1960). Regardless of their origin, the valley floors are not even plains. They are marked by significant stream dissection. Thus, most valley floors show a distinctive rolling topography with 30 to 90 meters (100 to 300 feet) of relief.

The topographic grain of the Ridge and Valley is predominantly northeast-southwest. Most streams and rivers follow this grain in the limestone and shale valleys. They break through the resistant sandstone ridges in spectacular water gaps, such as the Delaware Water Gap in eastern Pennsylvania. Many of the water gaps are located along zones of intense fracturing, termed lineaments. These short stretches through the ridges usually are oriented at right angles to the grain of the topography. This produces the trellis drainage pattern which is so characteristic of the Ridge and Valley.

The region is narrow in the north (22 kilometers, 14 miles, at the New York-New Jersey border) and it widens southward in Pennsylvania and Virginia (122 kilometers, 75 miles). One small area in Pennsylvania, New York, and New Jersey has been glaciated. In this area, in addition to the deposition of glacial till and outwash, lacustrine and marine materials were deposited, particularly in the section in northern New York.

Blue Ridge Province

The Blue Ridge Province extends from southern Pennsylvania through Virginia. The province gets its name from the Blue Ridge in Virginia which is a narrow mountainous ridge extending from the Potomac River to Roanoke. The famous Skyline Drive in Virginia is situated on the Blue Ridge. The Blue Ridge begins in southern Pennsylvania as the Carlisle prong, and ranges in width from 8 to 48 kilometers (5 to 30 miles). It is composed of primarily Precambrian granite, gneiss, and metamorphic volcanic rocks older than the folded Appalachians to the west, and often is referred to as the older Appalachians or Appalachian Highlands. The rocks are deformed intensely. Along its western edge the sedimentary rocks of the Ridge and Valley are turned up steeply, often along a fault contact which forms a sharp structural boundary between the two provinces. The rocks of the Blue Ridge are very resistant to erosion and have conspicuous relief, rising 305 to 610 meters (1,000 to 2,000 feet) above the Great Valley to the west and the Piedmont to the east.

Piedmont Province

The Piedmont Province extends from the Hudson River in the north through Virginia. It is composed primarily of Paleozoic metamorphic (schists and gneisses) and plutonic rocks (granites). Locally, belts of quartzite and pure marble are well developed. A belt of basic rocks containing talc and soapstone is present near the western boundary.

The Piedmont widens from about 16 kilometers (10 miles) in New York to about 190 kilometers (125 miles) at the Virginia-North Carolina border. Its topography is rolling, with relatively low relief varying from 15 to 90 meters (50 to 300 feet). Numerous hills or monadnocks rise 61 to 305 meters (200 to 1,000 feet) above the general surface. The rocks are intensely folded but beveled by an undulating erosional plain.

Within the Piedmont, various places in New Jersey, Pennsylvania, and Virginia have low lying areas of Triassic age sedimentary and intrusive, basic (high content of magnesium and calcium), igneous rocks. These Triassic basins are distinct from the other Piedmont rocks. The Triassic rocks exhibit gentle dipping in contradistinction to the gneisses and schists, and are cut by numerous faults. The sedimentary rocks are primarily red shales and sandstones which are intruded by diabase dikes and sills. The diabase is more resistant to erosion and tends to form ridges and higher areas surrounded by the sedimentary rocks. The outcome of the historic battle at Gettysburg was determined by the position of the Union Army on the high ground of one of the diabase sills. The diabase often is referred to as trap rock, from the German *treppen,* meaning step; and is used as an aggregate in road construction.

To the east, the Piedmont gives way to the Atlantic coastal plain where the crystalline rocks are overlapped by the flat-laying Cretaceous and Tertiary sediments. The boundary between the Piedmont and Coastal Plain is marked by the Fall Line, the location of falls along all major rivers crossing the Piedmont. The boundary also is a zone of mild earthquakes.

New England Province

The New England Province is essentially a northward extension of the Blue Ridge and the Piedmont. The geology of the Blue Ridge extends into the New England region through the Reading Prong (the Precambrian rocks extending from New England through New Jersey and into eastern Pennsylvania), the Berkshire Hills, and Green Mountains. The remainder of the New England area is composed of gneiss, schist, slates, and plutonic granites of an age similar to that of the Piedmont.

The major difference between the Piedmont and the New England area is that the New England area has been extensively glaciated. The effects of the glacial erosion are conspicuous. Much smoothing of the bedrock occurred during glaciation, but because of the hard crystalline nature of the bedrock, only thin patchy tills with many boulders were deposited. Also, as in the Northern Ridge and Valley area, there were extensive lacustrine materials deposited in the Connecticut River Valley as well as marine deposits along the coast of Maine. The topography of the New England area varies from a low relief of rolling glacial plain in its southern areas near the coast to more rugged areas of higher relief inland from the coast.

Like the Piedmont to the south, the New England Province has Triassic age basins. The largest basin is found in the Connecticut River Valley of Connecticut and Massachusetts (Figure 8). The glacial deposits derived from these deposits are much different than those derived from the gneisses, schists, and granites that are the bedrock in most of the New England area. The deposits tend to have more coloring and to have a different mineralogical composition.

Coastal Plain Province

The Coastal Plain, extending from Cape Cod in Massachusetts through Virginia, is a low-lying seaward-sloping plain found just east of the Piedmont. Relief generally is less than 150 meters (500 feet). North of Cape Cod the plain is submerged and is a part of the continental shelf. The geologic formations forming the surface of the Coastal Plain are Cretaceous, Tertiary, and Quaternary in age. The formations dip gently seaward (Figure 7) and outcrop as belts forming crests and valleys roughly parallel to the inner and outer edges of the Coastal Plain. The Cretaceous system forms an inland belt, the Tertiary an intermediate belt, and the Quaternary a coastal belt. These sediments are primarily sandy, but many silty and clayey sediments also occur. The Coastal Plain of the eastern United States has been subdivided into sections and within the Northeast region the Coastal Plain has been named the Embayed section. The most outstanding geomorphic feature of this section is the bays formed due to submergence of the Northern Coastal Plain area. The submergence reaches as far south as the James River in Virginia, where tidewater reaches the Fall Line. The submergence was the combined result of the depression of northeastern North America under the Pleistocene glacial ice, and the postglacial rise in sea level as the glaciers melted. Submergence was greater at the north than at the south, as evidenced by a decrease northward in the width of the Coastal Plain. This submergence created the Chesapeake and Delaware embayments by drowning the lower ends of the Susquehanna and Delaware rivers.

There are many distinctive levels, called terraces, on the Coastal Plain. The older levels adjacent to the Fall Line stand higher in elevation than the younger levels nearer the coast. The older levels also show a much greater degree of dissection and erosion. There tends to be a progression from the higher older levels to the lower younger levels of less dissection and flatter landscapes. The very lowest levels that are near sea level are characterized by many swamps and large areas of poorly drained soils.

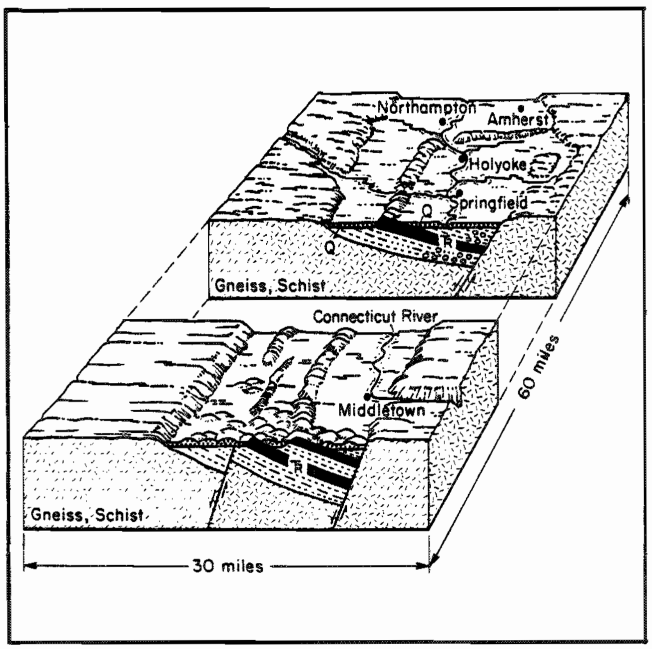

Figure 8. Block diagram illustrating the structure of the Triassic basin of the Connecticut River Valley. The basin is a half-graben because only one side is fault-bounded. Cross-hatched units are diabase, circles are conglomerates, dots are sandstones, and dashes are shales. From Hunt, Fig. 11.19, 1974, reprinted with permission of the publisher.

Adirondack Province

The Adirondack Province is found in northern New York and is one of the two extensions of Canadian Shield geology found in the United States; the other extension is the Superior Upland in Wisconsin. The Adirondack Province is a nearly circular structural dome more than 160 kilometers (100 miles) in diameter. The center of the dome has been uplifted more than 3.2 kilometers (2 miles). The bedrock of the Adirondacks is made up primarily of Precambrian igneous (granite) and metamorphic (quartzites, schists, and gneisses) rocks. Topographically, the Adirondacks are divided into central highlands and northwest lowlands. Many peaks in the highlands exceed 1,200 meters (4,000 feet) in elevation and local relief may be as much as 915 meters (3,000 feet). Like the remainder of New York, the Adirondacks were glaciated and glacial cirques are present on many of the higher peaks.

Central Lowland Province

The Central Lowlands encompass the relatively flat areas north of the Appalachian Plateau and west of the Adirondacks. From the lake level of Lake Erie, 175 meters (570 feet), and Ontario, 64 meters (244 feet), the land rises gently eastward and southward to an elevation of 305 to 450 meters (1,000 to 1,500 feet) at the Appalachian Plateau. The preglacial surface topography of this area has been modified greatly by the deposition of glacial till in the form of drumlins, moraines, and shoreline deposits.

St. Lawrence Valley Province

The St. Lawrence Valley is a small lowland area with some low hills (relief of about 30 meters, 100 feet) found north and east of the Adirondacks. It, like the central lowland area, has been greatly affected by lacustrine deposits as well as marine deposits caused by glaciation. The marine deposits accumulated in late glacial time and are a part of a much larger accumulation that was deposited from an arm of the ocean that flooded the St.

Lawrence Valley. Much of the earthquake activity of the northeastern United States is located in this province.

VEGETATION

The natural vegetation of the northeastern area is predominantly forest. Broadly speaking, there are five forest regions: spruce-fir, beech-birch-maple, white pine-hemlock-hardwood, oak-yellow poplar, and yellow pine-hardwood (Figure 9).

The spruce-fir region occurs in the coolest parts of the Northeast: central and northern Maine, the White and Green Mountains, the Adirondacks, and on mountain tops in West Virginia. This region is a dense evergreen forest with a few deciduous trees. Red, white, and black spruces and balsam fir predominate. Sugar maple, yellow birch, and beech prevail in some areas where changes in land use have modified the forest composition. White cedar may predominate in swamps, and white pine is common on many sandy sites.

The northern hardwood region (beech-birch-maple) has tall, broad-leaf, deciduous forest stands with a mixture of needleleaf evergreens. Predominant hardwoods are beech, yellow birch, and sugar maple. Also found in the region are stands of hemlock, sweet birch, red maple, basswood, white ash, northern red oak, and black cherry. Aspen, pin cherry, or paper birch may take over after a stand cutting and/or a fire. Spruce and fir intrude on poorer, colder sites and white pines and oaks may prevail on sandy soils. Black spruce and Northern white-cedar are common in swamps.

Northern red oak probably is the most widely distributed hardwood in the white pine-hemlock-hardwood region. Other oaks, hickory, and yellow poplar are found in the southern part of this zone. In the northern and higher areas, hardwoods typical of the beech-birch-maple region are common. White pines may occur in old fields or pastures, hemlock in ravines, and pitch pine on sandy areas. Swamp composition consists of red maple, black ash, elm, spruce, or Atlantic white-cedar in coastal sections.

Characteristic stands within the oak-yellow poplar region vary greatly because of differences in elevation, latitude, and soil properties. In the hills of West Virginia, Virginia, Pennsylvania, or Maryland, the characteristic oak species are black, white, chestnut, scarlet, and

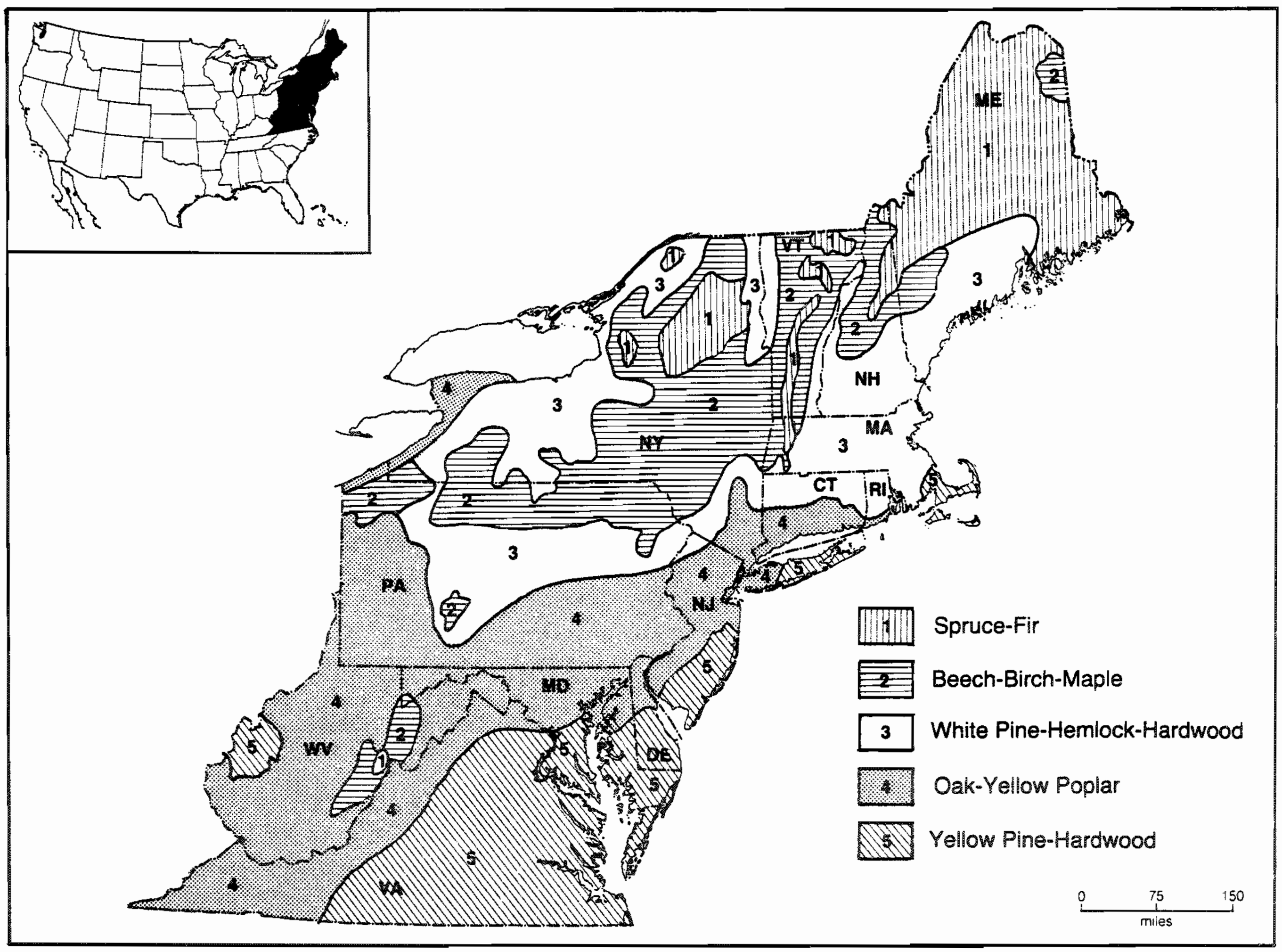

Figure 9. Major vegetation regions of the northeastern United States. From Kuchler, 1964; Lull, 1968.

northern red. In some sections of the lower Piedmont of Maryland or sections of northern New Jersey, nearly pure stands of yellow poplar are common. The pin, swamp, chestnut, and willow oaks, and sweetgum are characteristic on clayey soils of the upper Coastal Plain. Intrusions from other regions also can be found in localized areas and on the borders of this region.

The yellow pine-hardwood region has stands of medium tall to tall needleleaf evergreen as well as broadleaf deciduous forest stands. The characteristic pines of the region are pitch, loblolly, shortleaf, Virginia, and pond. Except for a small region in West Virginia, the yellow pine-hardwood region closely approximates the coastal plain areas and the Piedmont of Virginia. Yellow pines include pitch, loblolly, shortleaf, Virginia, and pond. Virginia and pitch pines predominate in West Virginia, as do pitch and shortleaf pines in New Jersey. In southern Maryland the major species is Virginia pine.

A wide range of hardwood species is found in association with the yellow pine. Species that commonly occur are hickory, white oak, post oak, red maple, and blackgum. Atlantic white cedar is common in the coastal swamps. Tupelo and bald cypress also are common on floodplains and in swamps in lower Delaware, eastern Maryland, and southeastern Virginia.

The forest species discussed in the previous paragraphs are the major species growing in the Northeast. However, variations can be noted throughout the region. Forest fires, clear cutting operations, and land-use histories have changed the composition of some forest stands.

CLIMATE

The climate of the northeastern states is extremely varied because of the differences in elevation, topography, and nearness to large bodies of water. The Appalachian Mountains have a major effect on the climate of the region. The Atlantic Ocean also has a modifying effect but it does not dominate the climate. Cold fronts moving in from the west are forced upward along the western slopes of the mountains. This orographic lifting is sufficient to cause a considerable amount of rainfall during the warm months and heavy snowfall during the winter (Baldwin, 1973). A similar effect occurs on the eastern slopes when fronts or storms move northward or westward from the Atlantic Ocean.

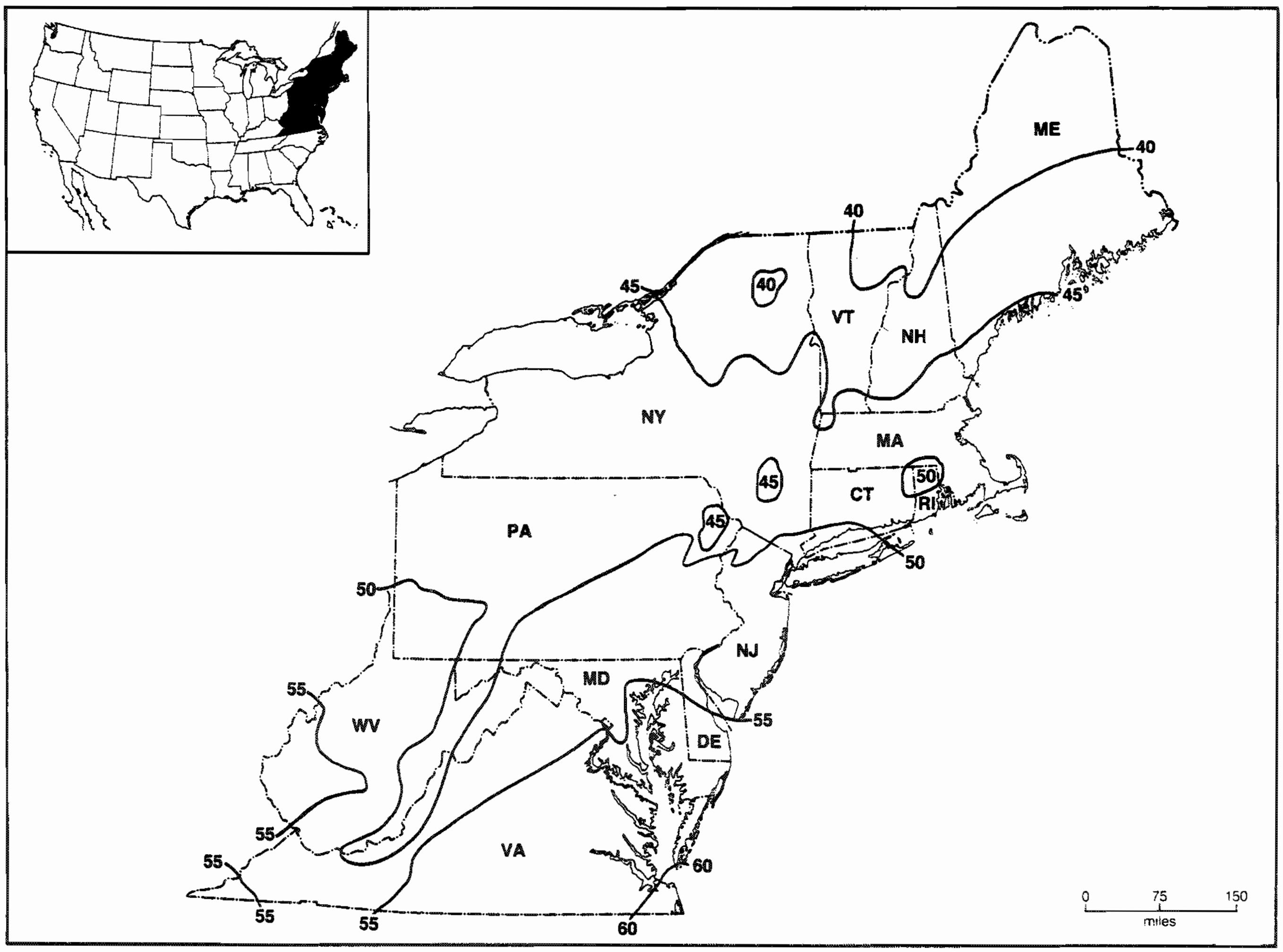

Figure 10. Average annual air temperatures of the northeastern United States. Presented in degrees Fahrenheit, convert to degrees centigrade by subtracting 32 then multiplying by 0.555. From USDA, 1941.

The climate of the region generally is characterized by the following statements (Ruffner, 1978):

1. An equal distribution of precipitation among the four seasons.
2. Large ranges in both daily and annual temperatures.
3. Great differences during the same season or same month of different years.
4. Considerable diversity of the weather over short periods of time.

The growing season, based on the number of freeze-free days (temperatures above 0°C, 32°F), ranges from less than 100 days in extreme northern Maine, New Hampshire, Vermont, and New York to over 220 days in southern Maryland and parts of Virginia (Ruffner, 1978). Because of the modifying effects of the ocean, coastal regions have longer growing seasons than the interiors of the states. The growing season for the coastal area of Maine ranges from 140 to 160 days. All other states along the coast have maximum growing seasons ranging from 180 to 200 days. Average annual air temperature of the region is shown in Figure 10.

Mean annual precipitation varies widely because of factors already mentioned (Figure 11). Snowfall also is quite variable within each of the states. In West Virginia, the average snowfall ranges from 50 centimeters (20 inches) at some of the lower elevations to over 254 centimeters (100 inches) at the highest. Snowfall in New Hampshire varies from 127 centimeters (50 inches) near the coast to 470 centimeters (185 inches) on Mt. Washington.

Cryic, frigid, and mesic soil temperature regimes are recognized in the Northeast (Smith, 1984). Recent data (Carter and Ciolkosz, 1980) show some central Appalachian regions probably have areas of frigid soils where none has been mapped.

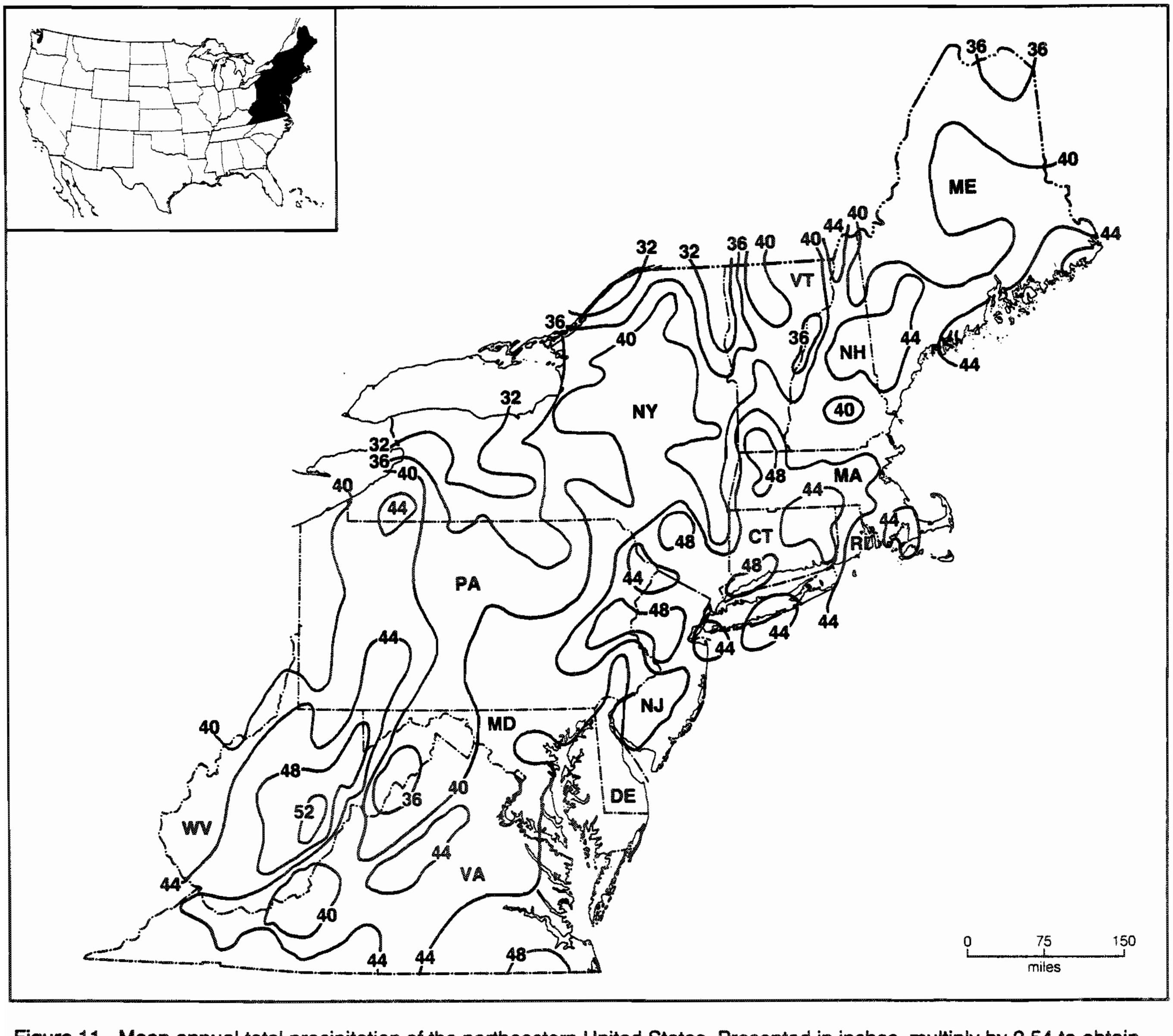

Figure 11. Mean annual total precipitation of the northeastern United States. Presented in inches, multiply by 2.54 to obtain centimeters. From Lull, 1968; Ruffner, 1978.

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CHAPTER 5

Geomorphology of Nittany Valley*

by

Thomas W. Gardner[1/]

Introduction

Nittany Valley lies within the Valley and Ridge physiographic province, very close to the geographic center of Pennsylvania (Fig. 5.1). Descending from the Appalachian Plateau to the northwest, it is the first primary agricultural valley of the Valley and Ridge (Fig. 5.2). The relatively flat valley bottom ranges from 800 to 1200 feet in elevation while nearby ridge crests display a nearly accordant elevation ranging from 2000 to 2400 feet.

Central Pennsylvania is characterized by a northern, moist, temperate climate (Cfa of Koeppen). In State College mean temperatures in January and July are -2°C and 22°C, respectively. Mean annual temperature is a mild 15°C. Mean annual precipitation is 980 mm/year with a slight summer maximum. No months show soil moisture deficits. The regolith is frozen to a depth of several feet during winter. Ridges around the valley experience slightly cooler temperatures and higher precipitation than the valleys. A greater percentage of precipitation falls on ridges as snowfall and it remains for a longer period during the winter season.

Spring Creek, Buffalo Run, and Little Fishing Creek are the main drainage lines in the northeastern half of Nittany Valley (Fig. 5.3). Bald Eagle Creek which is tributary to the West Branch of the Susquehanna River acts as local base level for those streams. Halfmoon Creek, Spruce Creek, and Warriors Mark Run drain the southern portion of the valley. The drainage divide occurs along a rough northwest line from Pine Grove Mills to Stormstown. The Juniata River, another large tributary of the Susquehanna, is local base level for the southern streams. The anomalously straight course of the Juniata through Nittany Valley and alignment of water gaps on opposing valley sides is controlled by the Tyrone lineament, a zone of intensely fractured bedrock. This structural feature was probably the result of cataclysmic mountain building forces during the Appalachian orogeny and has had a pronounced impact on drainage evolution.

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*Reprinted from E. J. Ciolkosz et. al. (editors). Soils and Geology of Nittany Valley. PA State Univ. Agronomy Series 64. 1980.

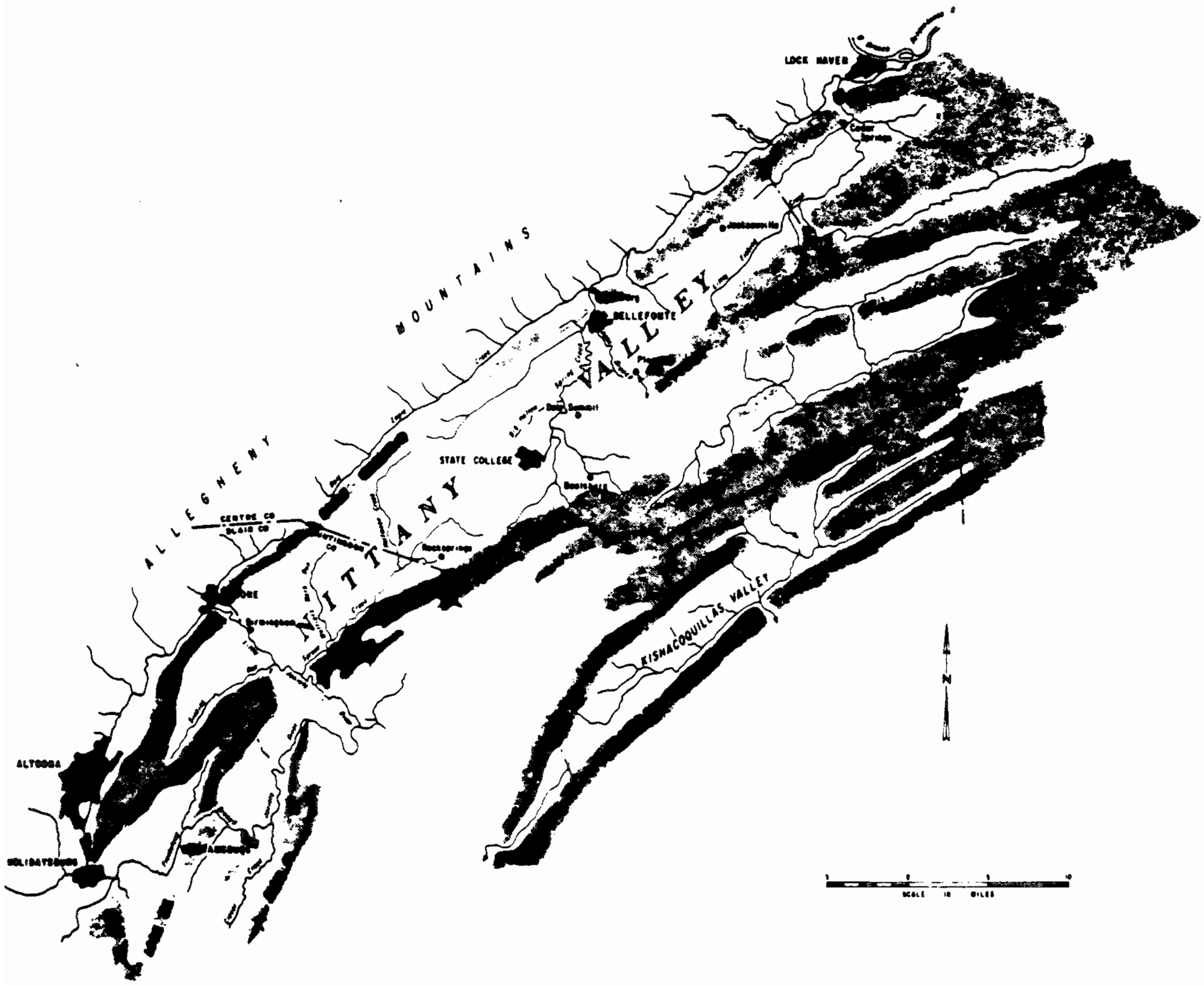


Fig. 5.3 Physiography and drainage in Nittany Valley (high relief areas are shaded).

General Geology and Geomorphology

The geology of Nittany Valley was mapped as part of the First and Second Pennsylvania Geological Survey (Rogers, 1858 and Lesley, 1885). More recently, Butts and Moore (1936) compiled a detailed geologic map of the Bellefonte quadrangle, encompassing a large portion of the State College area. In addition Clarke (1965) and Landon (1963) have mapped the local geology of specific areas in Nittany Valley. The general geology is summarized in Fig. 5.4.

Over 10,000 feet of lower Paleozoic sedimentary rocks are exposed in Nittany Valley and along adjacent rides. A geologic column of those rocks depicts the age, formational or member name, thickness and lithologic description and is given in Table 5.1. Approximately 7000 feet of Cambrian and Ordovician limestones, dolomites, and sandy dolomites underlie the valley floor. Over 3000 feet of upper Ordovician and lower Silurian sandstones, siltstones and shales outcrop along the flanks and crest of adjacent ridges. The oldest and youngest formations exposed in the area are the Mines Dolomite and Tuscarora Sandstone, respectively.

These sedimentary rocks were originally deposited as nearly horizontal beds, but attained their present structural position during the Appalachian orogeny in late Paleozoic time, approximately 250 million years ago. The most conspicuous structural feature, the Nittany Anticlinorium, is a large arch which extends from the boundary of the Appalachian Plateau for several miles to the southeast. This arch brings to the surface the oldest sedimentary rocks exposed in the area. Within the anticlinorium are numerous, smaller anticlines and synclines from which Nittany and adjacent valleys take their shape.

The topographic expression of Nittany Valley results from the complex interaction of structural setting, climatic and geologic history and weathering characteristics of different lithologic types. Fig. 5.5 shows a generalized relationship between lithology and topography. Based upon their weathering characteristics, various lithologies exposed in Nittany Valley can be divided into two fundamentally distinct groups; sandstones and shale that mechanically disintegrate and limestone and dolomite that chemically decompose. It is this basic dichotomy of weathering properties that creates the superb scenery of central Pennsylvania and gives it the name Valley and Ridge.

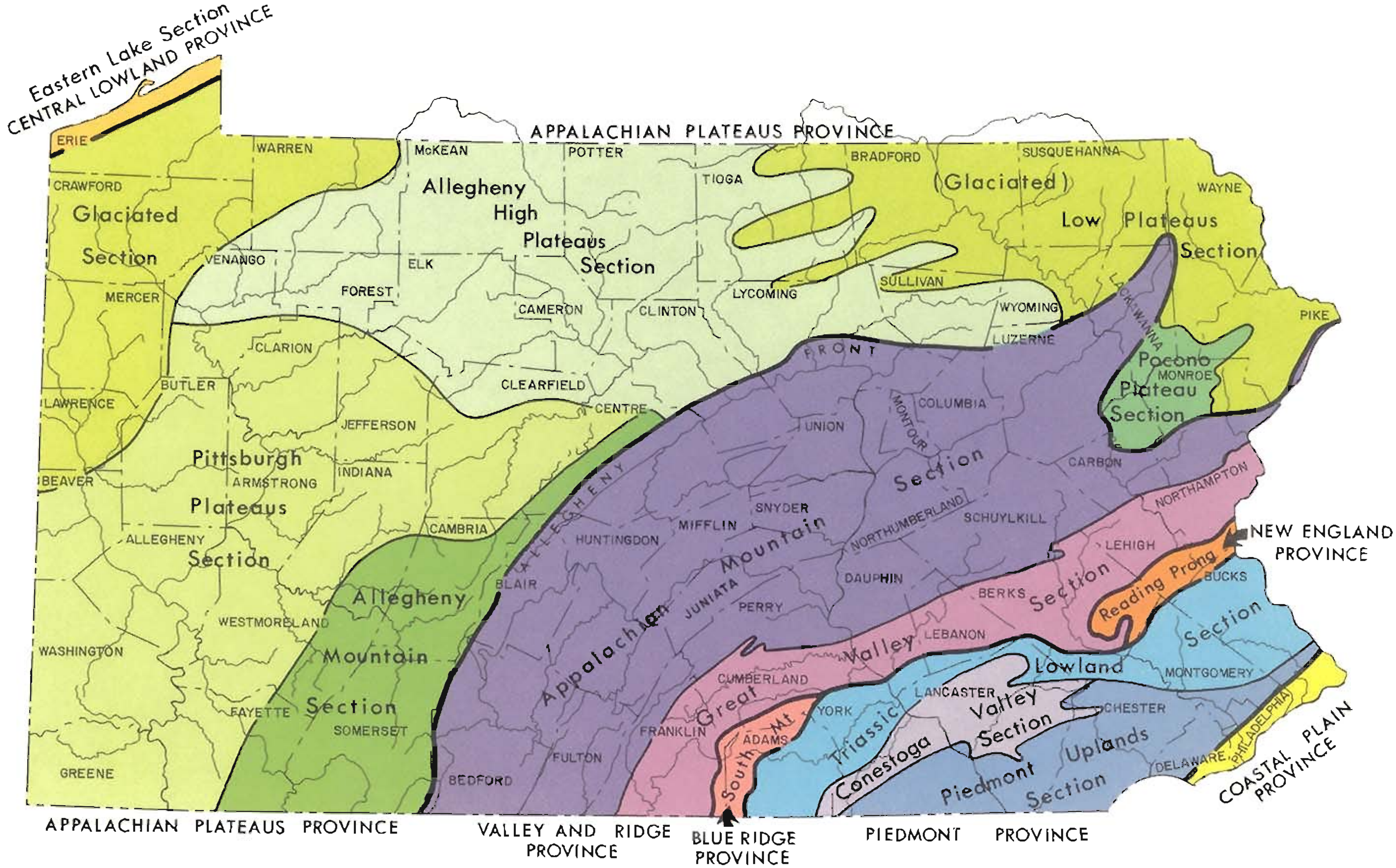
The Tuscarora, Juniata and Oswego Sandstones underlie ridges that bound Nittany Valley. The 500 foot thick, basal Silurian, Tuscarora Sandstone (St) is the most resistant formation in the lower Paleozoic section. It is a white to grey, fine to very coarse grained, siliceous cemented, quartzose sandstone. It forms the smooth unbroken skyline along most ridge crests. Ridges underlain by the Tuscarora maintain an elevation of approximately 2000 feet, but consistantly rise several hundred feet higher along the nose of plunging synclines and anticlines. Maximum elevations are usually found at structural noses where dip angles are small relative to flanks, thus exposing large areas of resistant rock at the surface. Slopes underlain by the Tuscarora range from 40° to 90°. A cliff or free face is commonly developed on the Tuscarora Formation.

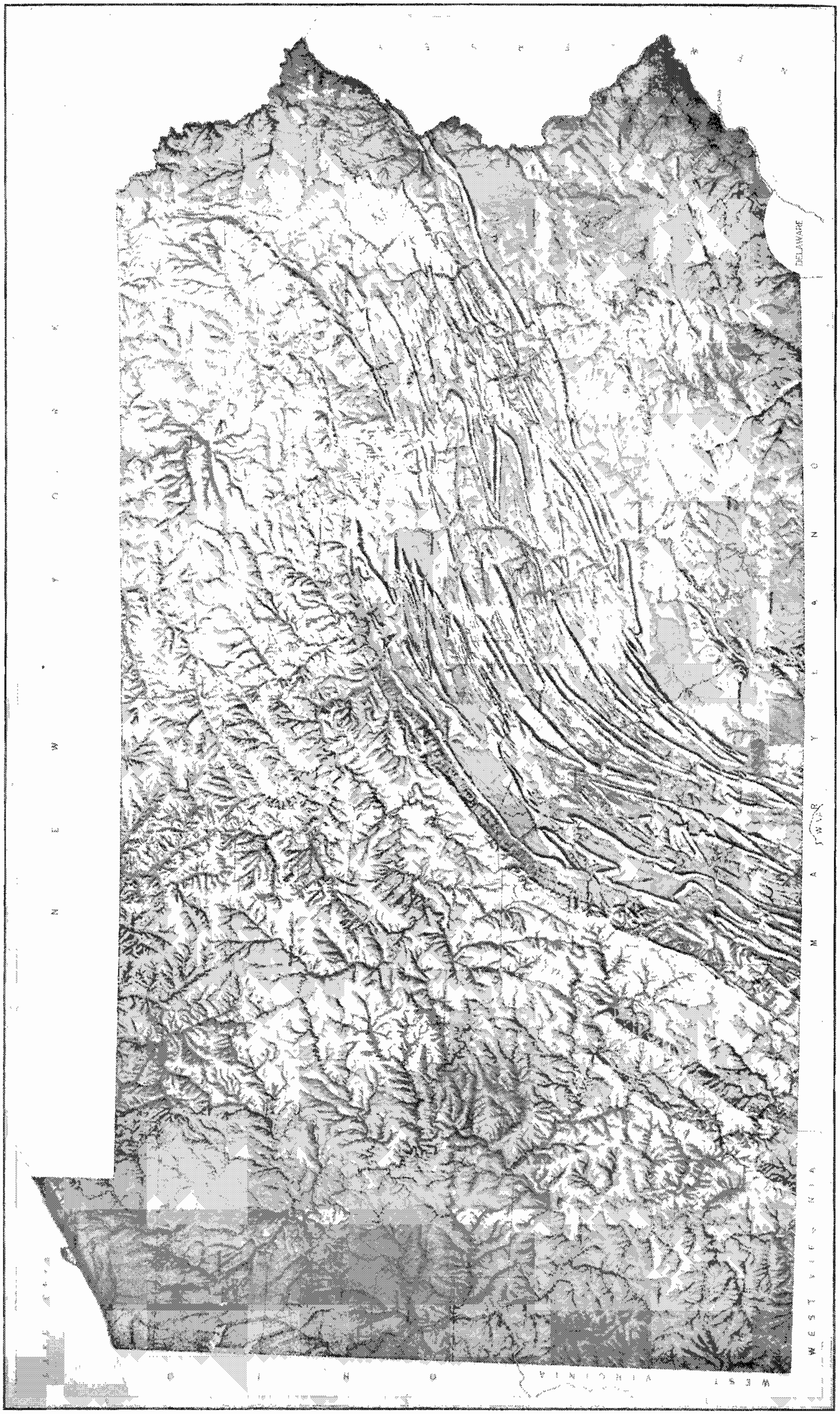
The upper Ordovician Juniata Sandstone (Oj) occurs immediately below the Tuscarora. It consists of 1000 feet of interbedded dark red to greyish red, fine to medium grained, sandstones, siltstones, and shales. The Juniata is less resistant than either the Tuscarora or subjacent Oswego and produces a high

MAP 13

COMMONWEALTH OF PENNSYLVANIA
DEPARTMENT OF ENVIRONMENTAL RESOURCES
TOPOGRAPHIC AND GEOLOGIC SURVEY
Arthur A. Socolow, *State Geologist*

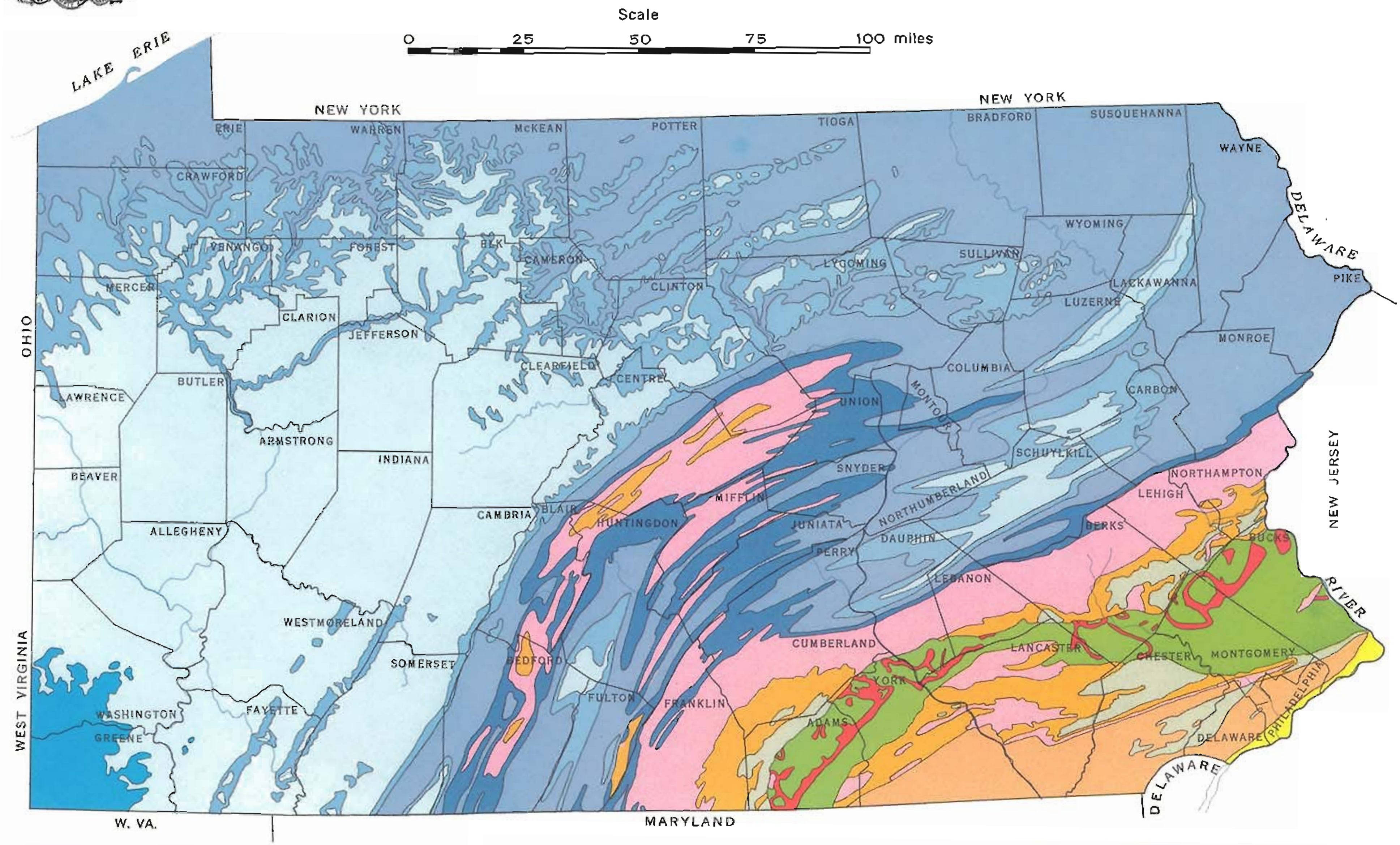
PHYSIOGRAPHIC PROVINCES OF PENNSYLVANIA




NEW YORK
DELAWARE
MARYLAND
WEST VIRGINIA
OHIO

GEOLOGIC MAP OF PENNSYLVANIA

COMMONWEALTH OF PENNSYLVANIA
DEPARTMENT OF ENVIRONMENTAL RESOURCES
TOPOGRAPHIC & GEOLOGIC SURVEY
Arthur A. Socolow, State Geologist



QUATERNARY (0-1 million yrs.) Sand and gravel. *sand and gravel.*

TRIASSIC (180-230 mil. yrs.) Shales and sandstones intruded by diabase. (red) *iron, building stone.*

PERMIAN (230-280 mil. yrs.) Cyclic sequences of sandstone, red beds, shale, limestone, and coal.

PENNSYLVANIAN (280-310 mil. yrs.) Cyclic sequences of sandstone, limestone, shale, clay, and coal. *coal, clay, lime.*

MISSISSIPPIAN (310-350 mil. yrs.) Red beds, shale, and sandstone.

DEVONIAN (350-400 mil. yrs.) Red beds, shale, sandstone, limestone and chert. *silica sand.*

SILURIAN (400-425 mil. yrs.) Sandstone, red beds, shale, and limestone. *gannister, lime.*

ORDOVICIAN (425-500 mil. yrs.) Shale, limestone, dolomite, sandstone. *slate, limestone, zinc.*

ORDOVICIAN and/or CAMBRIAN (425-600 mil. yrs.) Metamorphic rocks-schist, serpentine, gneiss and quartzite. *Building stone.*

CAMBRIAN (500-600 mil. yrs.) Limestone and dolomite; some sandstone and shale. *lime and gannister.*

PRECAMBRIAN (Older than 600 mil. yrs.) Gneiss, greenstone, serpentine and anorthosite. *zinc, building stone, graphite, sericite.*

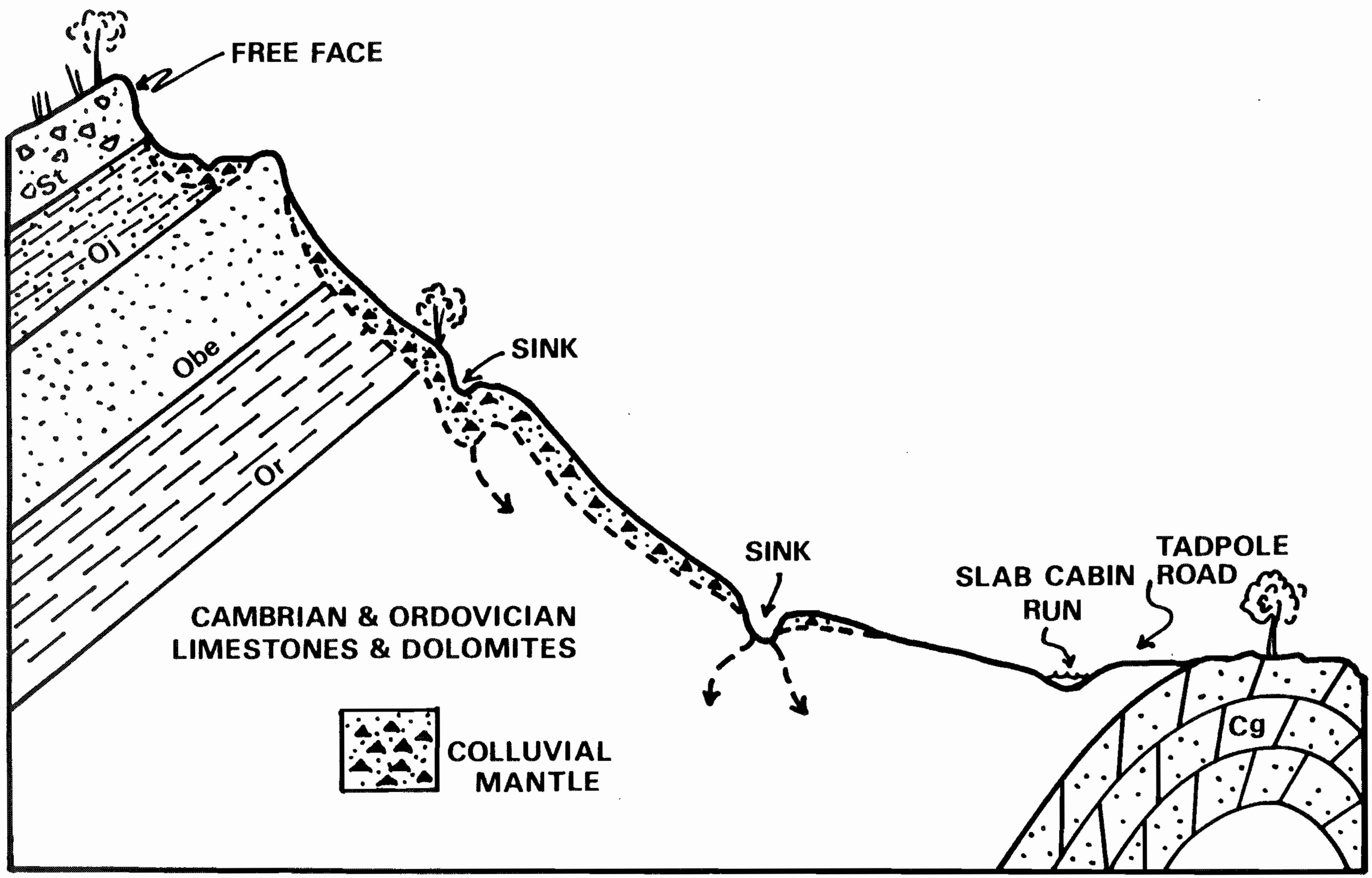


Fig. 5.4 Distribution of colluvial deposits along the ridges of Nittany Valley.

Table 5.1 Geologic column of rocks exposed in Nittany Valley.

SYSTEM	SERIES	FORMATION / MEMBER		THICKNESS (ft)	LITHOLOGIC DESCRIPTION	No. of Wells
ORDOVICIAN	MIDDLE	Snyder Formation		190	Limestone, 4 inch to 1-foot beds, fine to medium grained; interbedded dolomite, oölitic beds, mud-cracked beds, clay partings, and coarse bioclastic beds.	
ORDOVICIAN	MIDDLE	Hatter Formation		100	Limestone, 4-inch to 2-foot beds, fine to medium grained, with laminated argillaceous and arenaceous dolomite, fossiliferous, worm borings. (Unconformity)	
ORDOVICIAN	MIDDLE	Clover Limestone		80	Limestone, 2-inch to 2-foot beds, fine to very fine grained, laminated with fine to coarse-grained limestone.	
ORDOVICIAN	MIDDLE	Milroy Limestone		300±	Limestone, fine-grained, silty; laminated with wavy dolomitic bands.	2
ORDOVICIAN	LOWER	Bellefonte Dolomite	Tea Creek Member	200	Dolomite; fine-grained to sublithographic, thin shale partings, gashed weathered surfaces; 1 to 4 foot-beds.	22
ORDOVICIAN	LOWER	Bellefonte Dolomite	Dale Summit Member	0 to 14	Sandstone, fine to coarse-grained, conglomeratic.	
ORDOVICIAN	LOWER	Bellefonte Dolomite	Coffee Run Member	1000	Dolomite, interbedded fine-to medium-grained, cyclic successions.	
ORDOVICIAN	LOWER	Axemann Limestone		400	Limestone, fine to coarse-grained, oölitic, interbedded thin layersof impure dolomite, fine to medium-grained, partly conglomeratic, chert locally.	
ORDOVICIAN	LOWER	Nittany Dolomite		1200	Dolomite, fine-to coarse-grained alternating in cyclic manner, spherical chert nodules, oölitic chert, thin limestone and sandy beds.	15
ORDOVICIAN	LOWER	Stonehenge Limestone		600	Limestone aphanitic to fine-grained, argillaceous and dolomitic in part, flat pebble conglomerate abundant.	
CAMBRIAN	UPPER	Gatesburg Formation	Mines Member	150 to 230	Dolomite, interbedded coarse-to fine-grained, chert abundant, oolitic, thin sandy beds near base, vugular.	
CAMBRIAN	UPPER	Gatesburg Formation	Upper Sandy Member	500±	Dolomite; with interbedded orthoquartzites, and sandy dolomites, some shaly dolomites, fine to medium grained, vugular.	22
CAMBRIAN	UPPER	Gatesburg Formation	Ore Hill Member	260	Dolomite; fine to medium grained thick-bedded near top, fine-grained argillaceous dolomite near center; coarse-grained, massive bedded at base.	
CAMBRIAN	UPPER	Gatesburg Formation	Lower Sandy Member	700±	Dolomite, interbedded orthoquartzites, thin-bedded fine-to medium grained dolomite, medium to coarse-grained orthoquartzite; thin bedded shaly dolomite.	
CAMBRIAN	UPPER	Warrior Formation		600	Limestone, in part dolomitic, thick-bedded, with thin-bedded shale and sandy units.	

Table 5.1 (Continued)

SYSTEM	SERIES	FORMATION / MEMBER	SECTION	THICKNESS (ft)	LITHOLOGIC DESCRIPTION	NO. of Wells
DEVONIAN	LOWER	Helderberg Formation		150 to 350±	Shale, thin-bedded, calcareous. Limestone, thin-bedded, cherty. Sandstone, locally medium to coarse-grained.	1
SILURIAN		Keyser Formation		155	Limestone, thick-bedded to nodular.	
		Tonoloway Formation		400+	Limestone, thin-bedded to laminated, fine-grained, some calcareous shales.	1
		Wills Creek Formation Bloomsburg Formation McKenzie Formation		1500+	Shales, calcareous in part.(undifferentiated)	
		Rose Hill Formation Tuscarora Formation		400 to 550	Quartzitic Sandstone, fine to very coarse-grained, thin to thick bedded, mountain former.	
ORDOVICIAN	UPPER	Juniata Formation		1000+	Sandstone, fine-to coarse-grained, impure; interbedded siltstones and shales.	
		Oswego Sandstone		700 to 800	Sandstone, fine-to coarse-grained, interbedded shale near base.	1
		Reedsville Shale		1000	Shale, sandy in upper portion.	8
		Antes Shale		200	Shale, calcareous, soft.	
	MIDDLE	Coburn Limestone		275	Limestone, thin-bedded, fine to coarse grained, shale partings.	3
		Salona Limestone		175	Limestone, thin-bedded, fine-grained, shale partings.	
		Nealmont Formation		70	Limestone, impure bioclastic, fine to medium grained near top; thin to thick-bedded impure, fine-grained limestone near base.	5
					(Unconformity)	
		Linden Hall Formation: Valentine Member, Oak Hall Member, Valley View Member		180	Limestone, thick to thin-bedded, very fine to medium-grained. (Valentine Mb. laminated thick to thin bedded units.);(Valley View,Mb. 2-inch to 1 foot bedded well laminated limestone, thin clay laminae.); (Oak Hall,Mb. thick-bedded, fine to coarse-grained limestone.)	
		Linden Hall Formation: Stover Member			Limestone, thick bedded, bioclastic zones, thin bentonite beds, dolomite streaks.	

level valley between the two. Intercalation of the Juniata with the Tuscarora and Oswego produces the pronounced double ridge when viewed from inside anticlinal folds, ie. Nittany Valley. This high level valley serves as drainage collection areas for small streams that have carved and maintained gaps through the Oswego Sandstone.

The Oswego Formation (Os), locally termed the Bald Eagle (Obe), outcrops below the Juniata and is the first ridge as one proceeds out from Nittany Valley. It consists of 800 feet of white to grey, fine to coarse grained, well cemented sandstone with interbeds of shale and siltstone. The Oswego ridge usually remains several hundred feet below the dominant Tuscarora Ridge. The Tuscarora has been eroded from the synclinal Nittany Mountain, (Fig. 5.3), thus only a single ridge crest is preserved. Gaps in the Oswego are frequently localized along zones of structural weakness called fracture traces. They can frequently be traced into the carbonate valley where they are marked by a lineation of sinkholes. Oddly, they have very little effect on the topography of the Tuscarora Ridge.

Taken together these three formations form a very resistant sequence of sedimentary rocks. Mechanical disintegration and mass wasting are the primary modes of surface lowering. Disintegration results from expansion of water in joints, fractures, bedding planes, and pore spaces during freeze-thaw cycles. Hydrolysis of feldspars and other unstable mineral grains weakens the rock fabric, thus aiding disintegration. However, it is probably a volumetrically minor process. As a result, angular, blocky colluvium covers most of the slopes along the ridge flanks to variable depths. Origin of this colluvial mantle is discussed in a later section. Physical properties of the colluvium change along slope in both the horizontal and vertical direction. Blocks typically range from 10 inches to 5 feet in diameter and generally decrease in size downslope, while rounding increases. Admixtures of silt and clay tend to increase downslope, but locally the colluvium is devoid of fine grain sizes. The deposit is then referred to as a scree slope. Griffiths (1959) studied variation in size, size-sorting and axial proportions in scree blocks in an adjoining valley, but found no statistically significant geographic variation within a small sample site. Colluvial and scree deposits are very permeable. Few perennial streams maintain courses down the ridge flanks (Ciciarelli, 1971). Most flow is in the subsurface or in a few larger drainage lines that have cut gaps in the Oswego. However, during extreme storm events and snowmelt some local depressions do maintain surface flow which succeeds in removing fine grained material. This gives the ridges a ribbed appearance at low sun angles.

One thousand feet of Ordovician Reedsville Shale (Or) marks the transition between the relatively flat carbonate valley and the adjacent ridge. The contact between the Oswego and Reedsville is obscured by colluvium, but a sharp, discerning eye can usually see a break in slope several hundred feet below the Oswego ridge crest. This break in slope is usually close to the Oswego-Reedsville contact. A smooth concave upward profile ranging from 10° to 40° is characteristic of the Reedsville outcrop. Few perennial streams cross it for reasons already stated. Most flow is through the colluvial mantle. However, the colluvium frequently thins along the lower portion of the Reedsville Shale and groundwater discharge occurs along a line of seeps. Local residents use the seeps for irrigation and drinking water.

A basic change in the topography and weathering characteristics is reflected in the carbonate rocks of Nittany Valley. Over 7000 feet of interbedded limestones, dolomites and sandy dolomites outcrop on the valley floor. The surface of these rocks is lowered by chemical decomposition. Residual soils from 0 to 100 feet thick blanket most of the carbonate rocks (see Ciolkosz _et al_., 1980). Limestone and dolomite are relatively insoluble in pure water, but with addition of dissolved carbon dioxide from soil and atmospheric sources solubility increases rapidly with limestone being the more soluble mineral species. Dissolution is most rapid along zones of fracture concentrations, joints, bedding planes, faults, and other zones of increased permeability. Zones of fracture traces tend to localize drainage lines in the carbonate valleys (Fig. 5.6). Ciciarelli (1971) has shown significant alignment of joint orientation and straight stream segments in Sugar Valley, northeast of State College. First order tributaries occasionally follow formational boundaries in Nittany Valley and can be used with care to delineate bedrock geology. Soil series can also be used with similar caution to map geology.

Frequently, limestone formations are topographically above dolomite formations, yet limestone is more soluble. This interesting phenomenon has been explained by Parizek (1971). The model as depicted in Fig. 5.7 involves three stages. In the initial stage (Fig. 5.7a) surface topography is relatively flat. Thick residual soils mantle bedrock, attaining maximum thickness along fracture traces in more soluble limestone. Open sinks and large cavities that promote roof collapse develop on limestone, but not on less soluble dolomite. In stage two (Fig. 5.7b) internal drainage is soon developed on limestone and residual soils are transported into sinks and carried away by subsurface channels. As a result more numerous bedrock outcrops develop on limestone. When downcutting by streams is fast enough, limestone will project through the thin soil cover.

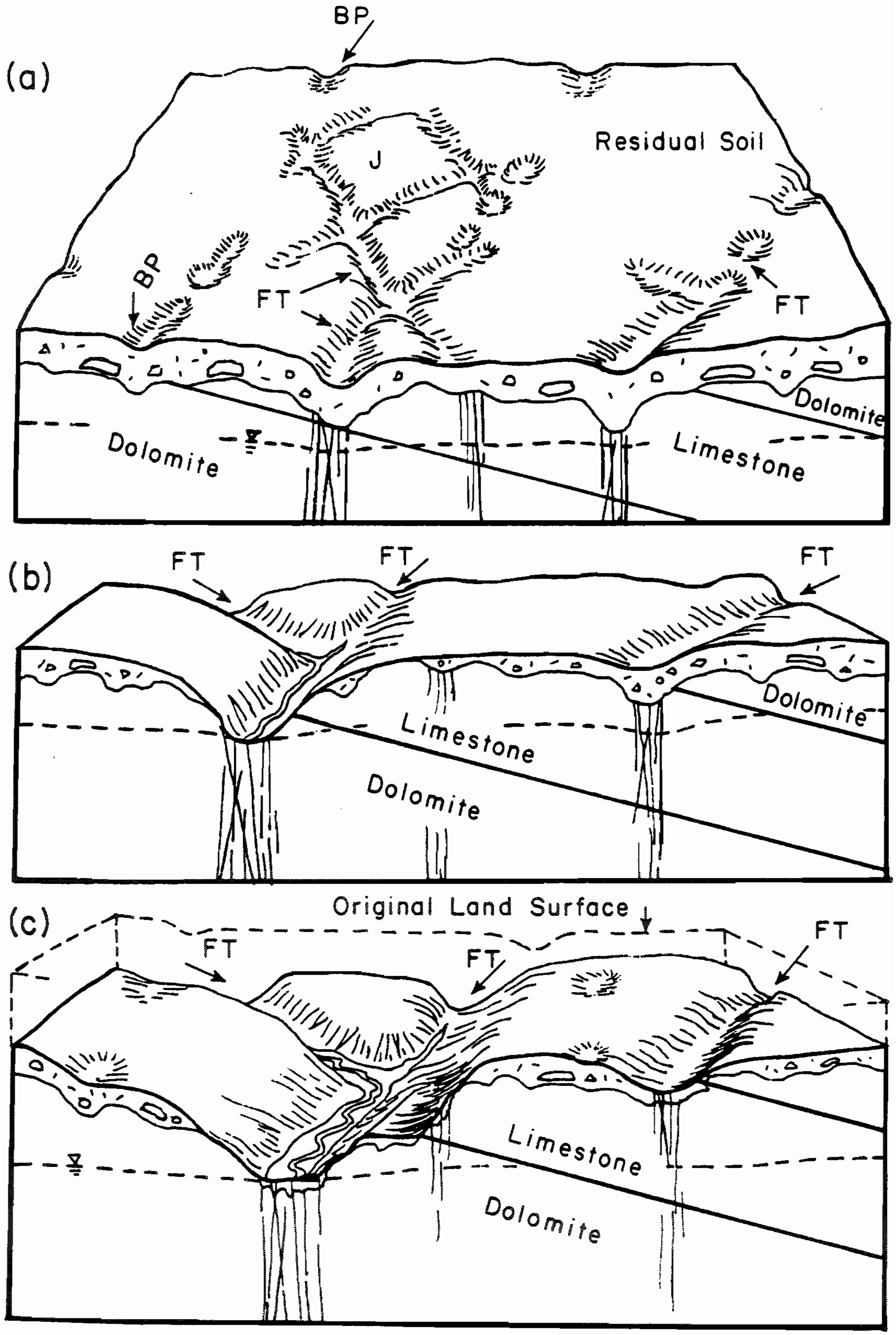


Fig. 5.6 Localization of valleys along fracture traces in limestone lithologies (after Parizek et al. 1971).

During stage three (Fig. 5.7c) the role of soil cover becomes important. Carbonation is an order of magnitude greater in soil water than in atmospheric water. Soil water is capable of dissolving more carbonate rock than an equal volume of rain water that has fallen on bare rock. Since soil cover is thicker on dolomite in stage three, dolomite will dissolve more rapidly and its surface will be lowered below the adjacent, exposed limestone. However, solution will continue in limestone along rapidly expanding passageways at depth. As a result of this differential weathering, limestone, dolomite and sandy dolomite lithologies display significant differences in mean and maximum slope angles as well as elevation.

Geomorphic Evolution of Nittany Valley

Since the late 19th century, the Appalachian Mountains and central Pennsylvania, in particular, have been the focus of an ongoing debate about the origin and evolution of topography and drainage. Many students of geomorphology have addressed this subject and publications are as numerous as ideas. Classic works on the subject of Appalachian topography include Ashley (1935), Davis (1889), Hack (1960), Meyerhoff (1972), Meyerhoff and Olmsted (1934, 1938), Oberlander (1965), Thompson (1931), and Ver Steeg (1930). The reader is referred to Melhorn, *et al*. (1972) and Craig (1978) for excellent reviews of the subject matter. The purpose here will be to present a detailed description of two aspects of this broad, complex subject that are pertinent to the geomorphic evolution of Nittany Valley.

Davis (1889) presented a model for the evolution of Appalachian topography that has prompted much controversy. A modified version illustrating Nittany Valley is presented in Fig. 5.8. In his study of Sugar Valley, Ciciarelli (1971) has elaborated on certain aspects of the initial stages of the model. The model can be segmented into four convenient stages. For simplicity, the model starts with an initial surface of Tuscarora Sandstone (Fig. 5.8A) that is folded into anticlines and synclines. At this stage anticlines are structurally and *topographically* the highest point on the land surface. First order consequent streams (streams without tributaries) develop along the flanks of anticlines and drain into larger, consequent streams in synclines. Synclines act as focal points for collection of drainage. Consequent synclinal streams drain to the northeast, down the structural plunge. The Tuscarora is breached by first order streams at the structural culmination along the crest of each anticline. Ciciarelli (1971, Fig. 29) has shown that breaching initially occurs at the structural culmination as a result of increased fracture density in that area.

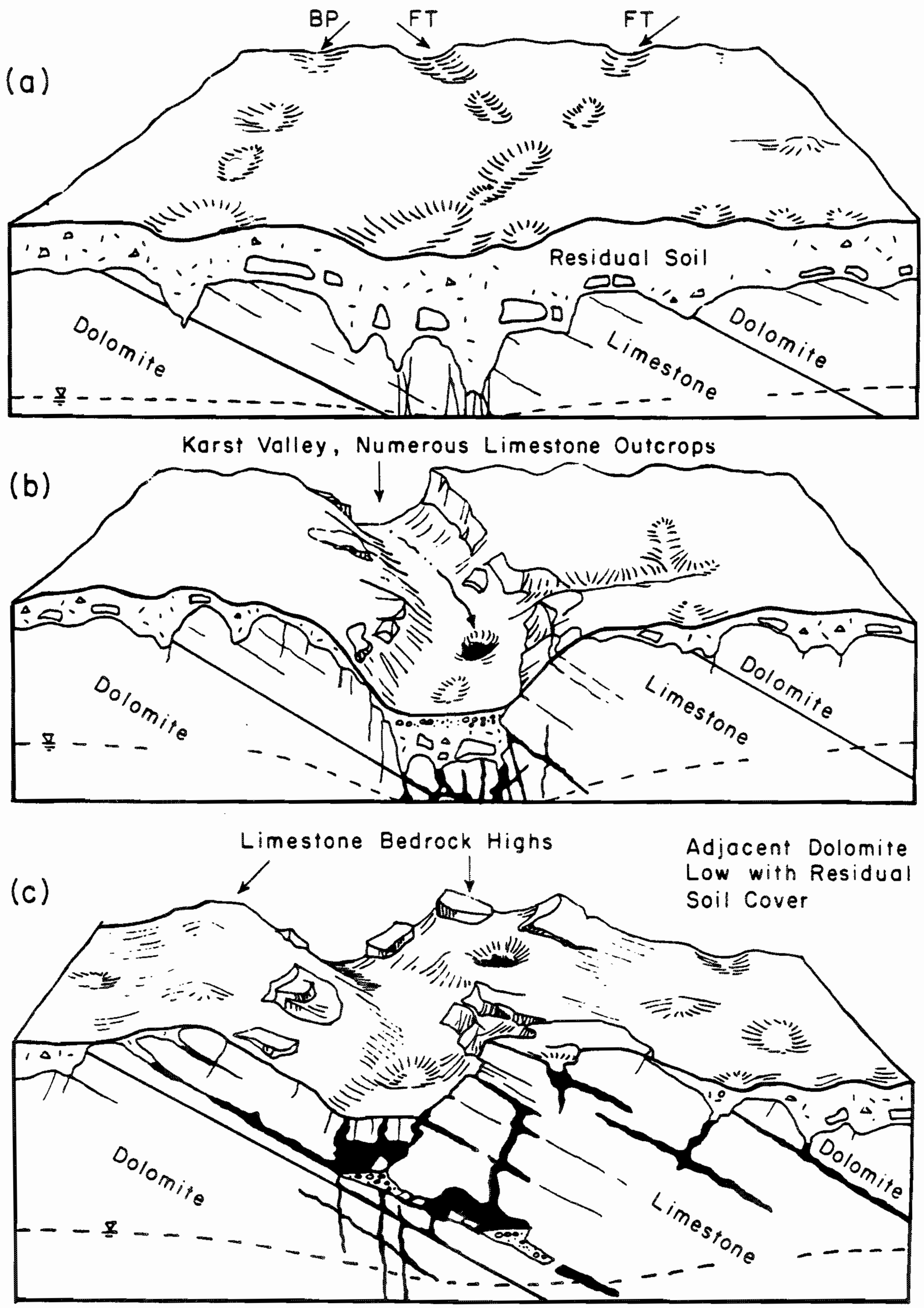


Fig. 5.7 Topographic highs developed in Stonehenge limestone in contrast to the Gatesburg dolomite.

The Tuscarora is mechanically weakened and more susceptible to erosional processes. Cliffs form in the Tuscarora and erode headward, exposing the less resistant Juniata. Small drainage basins with subsequent streams develop as the valleys expand in less resistant rock.

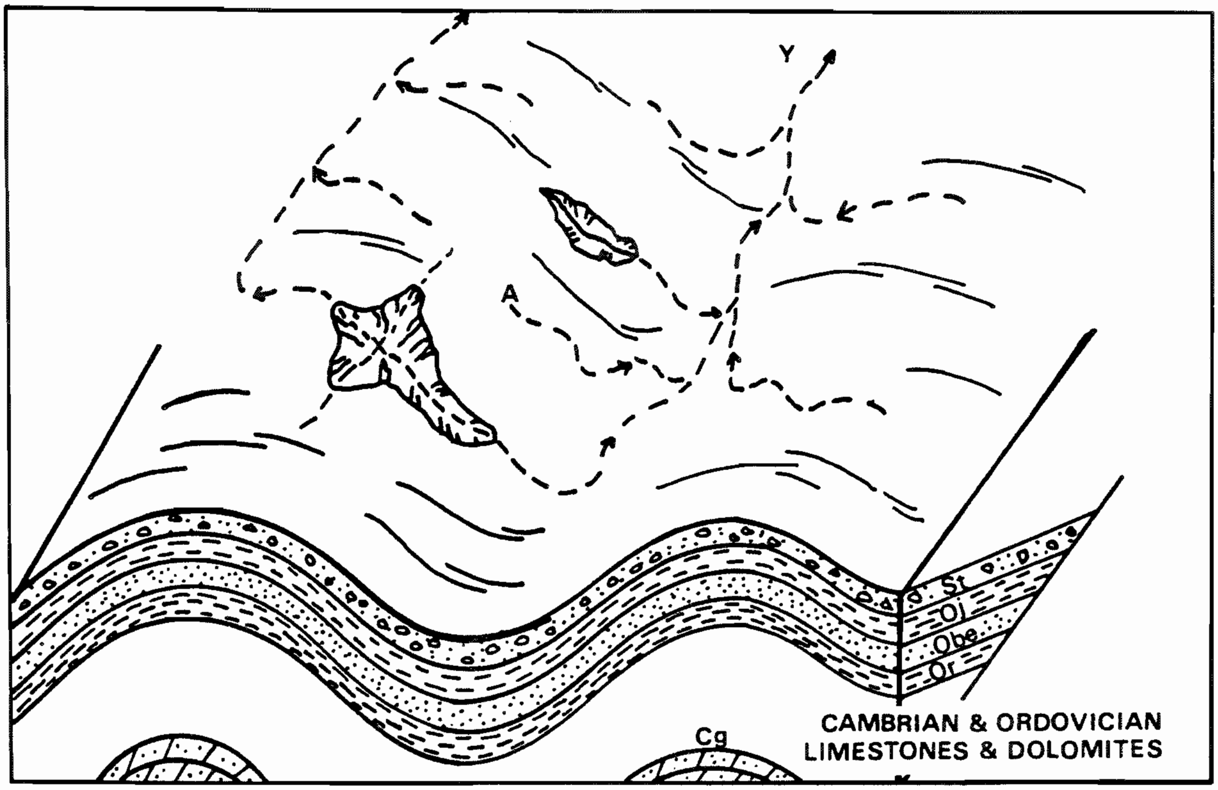
In the second stage (Fig. 5.8B) individual basins are separated by cols of Tuscarora. Drainage divides migrate headward and basins coalesce. As downcutting continues, the Oswego is exposed along the topographically high anticlinal crest and the characteristic double ridge takes shape. During the third stage (Fig. 5.8C) the drainage line which has an exit gap at the lowest base level (furthest down the plunge of each anticline) captures drainage from other subsequent basins. As the Tuscarora and Oswego cliffs erode headward, anticlinal valleys continually gain surface area at the expense of consistantly shrinking synclines. Cambro-Ordovician carbonates are exposed in the anticlinal core. Since carbonates are less resistant than the sandstones, a process of topographic inversion takes place. The limestone surface is lowered faster than the adjacent sandstones: the originally higher anticlines become topographic lows while the lower synclines become topographic highs. With continued topographic inversion drainage is diverted from the synclinal axes and an integrated drainage pattern is established in the newly formed anticlinal valley.

During stage four (Fig. 5.8D) major drainage lines in Nittany Valley assume their modern configuration. The present surface takes shape and topographic inversion reaches its culmination.

An interesting complication to this model arises from the fact that downcutting may not have been continuous during all stages of topographic development. Here, we must turn again to the ideas of Davis (1899a and b) for his concepts of the cycle of erosion and peneplanation. The concept of the cycle of erosion was a masterful synthesis of then current geomorphic thought on the evolution of regional landscapes. According to Davis, the ideal cycle begins with rapid uplift of a land mass. It is followed by a long period of tectonic quiesence. During the period of stability the landscape progresses through a sequence of stages; youth, maturity, and old age. As the region under consideration passes from one stage to another, its characteristic features gradually change until the once mountainous area has been worn down by erosion. What remains in old age is a flat, featureless plain, a peneplain, which develops at the regional base level.

It has been suggested that three major episodes of peneplanation are preserved in the Appalachian Mountains. From oldest to youngest, they are the

A



B

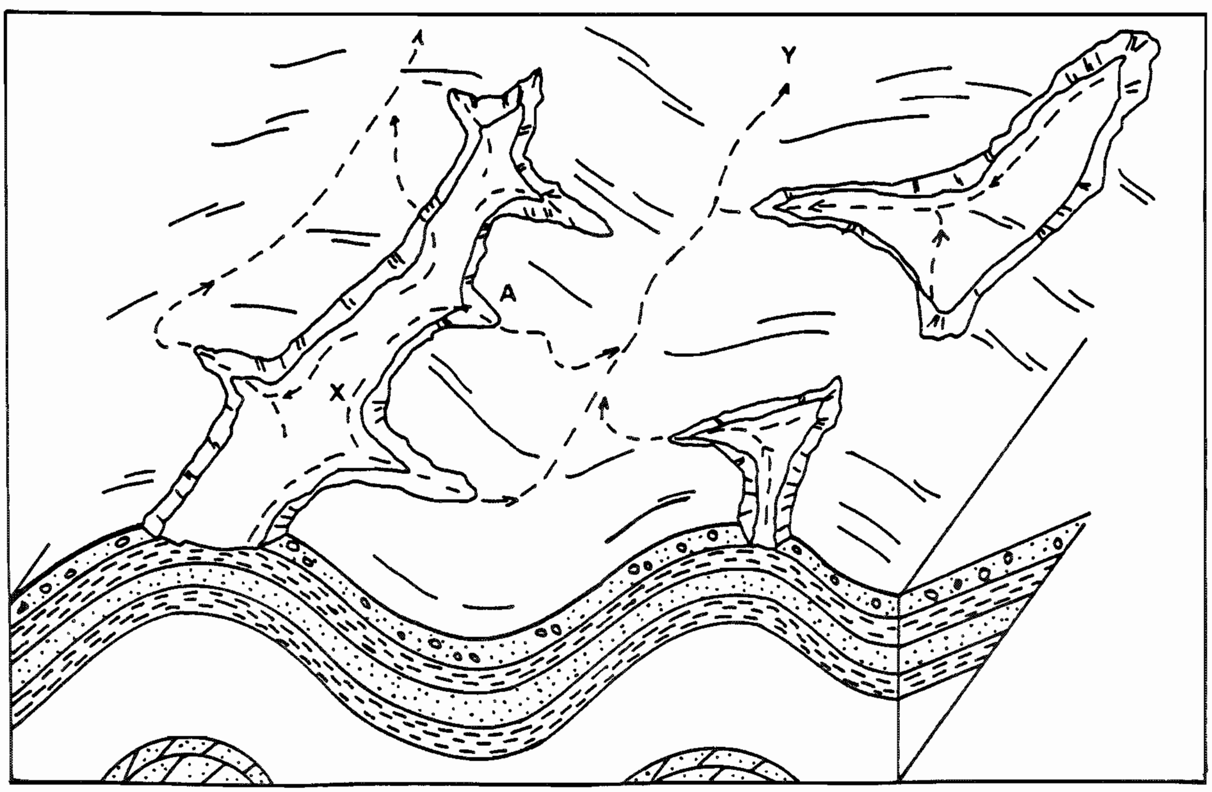
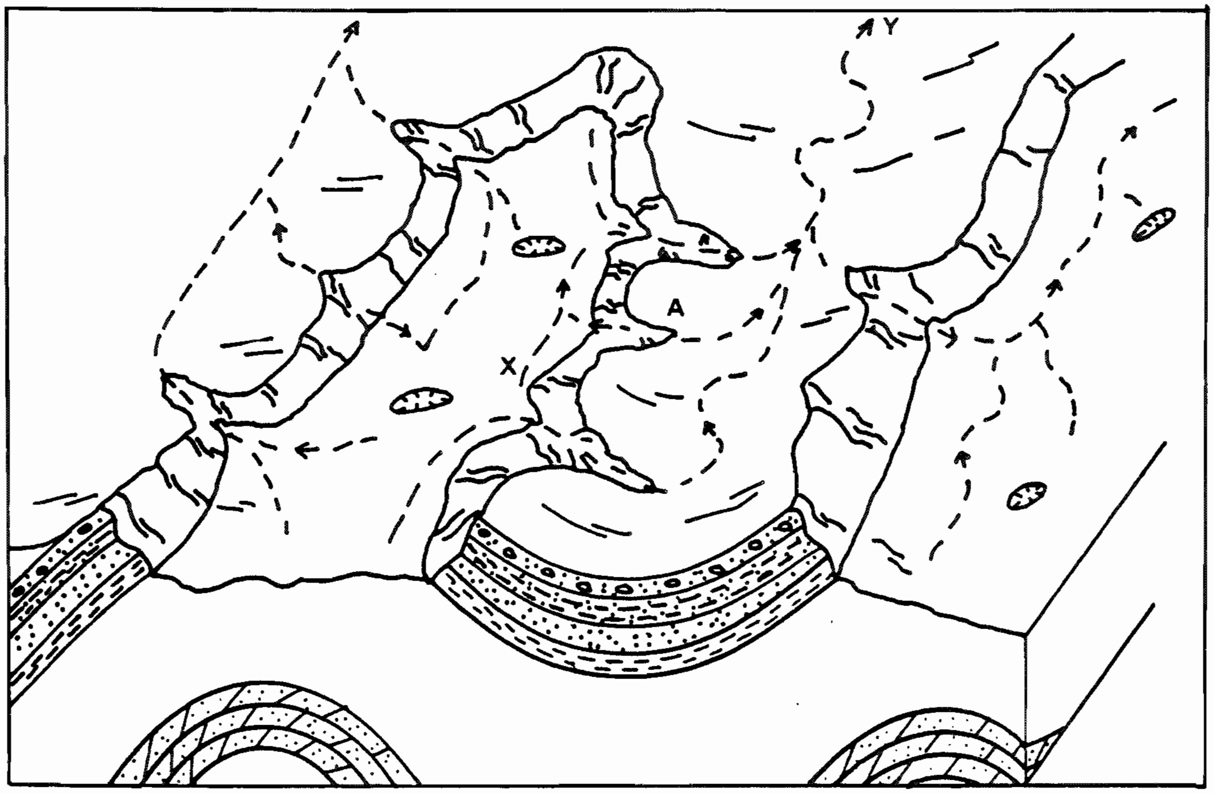


Fig. 5.8 Four stage (A, B, C and D) model for the geomorphic evolution of Nittany Valley. Y is a major synclinal, consequent stream. X is a developing anticlinal subsequent stream. A is a fixed reference point showing the capture of the syclinal stream in part C.

C



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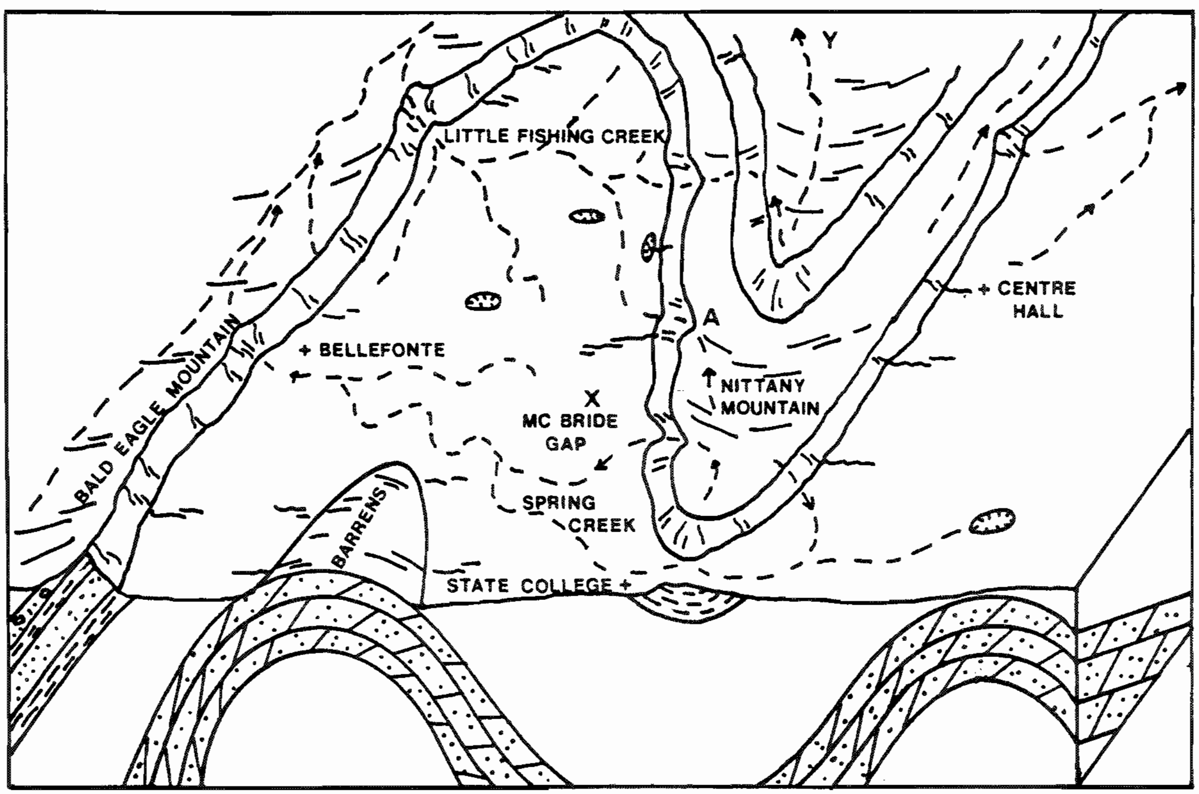

Fig. 5.8 (Continued)

Schooley, Harrisburg and Somerville Surfaces. The evolution of those peneplains from initial Appalachian folds is depicted in Fig. 5.9. The Schooley Surface is reported to be either Cretaceous (Davis, 1889) or early Tertiary (Johnson, 1931) in age, but data is equivocal. It is one of the most complete cycles, effectively beveling the folded sedimentary rocks in Pennsylvania and adjoining areas (Fig. 5.9B). One of the most striking attributes of the physiography of the Nittany Valley area, the accordant summits of the Tuscarora Ridges, is thought to have resulted from Schooley peneplanation. Superposition of major drainage lines from the Schooley Surface has been suggested as a cause for transverse Appalachian drainage, i.e. drainage that runs across the structural grain of the folded sedimentary rocks.

After Schooley peneplanation the area was subjected to renewed uplift and stream downcutting (Fig. 5.9C). Valleys were opened in less resistant rock types. Another poise in downcutting resulted in the formation of the Harrisburg peneplain in middle Tertiary time. However, the cycle was interrupted by renewed downcutting and the Harrisburg surface was only developed on less resistant lithologies. Thus, the Harrisburg Surface has been termed a partial peneplain. The wide expanse of Nittany Valley marks the Harrisburg surface. Development of the Somerville Surface (Fig. 5.9D) is marked by incised valleys of major drainage lines in Nittany Valley. Further southeast, it is more fully developed on carbonate rocks of the Great Valley. Incision is thought to have occurred in latest Tertiary or Quaternary time.

Many geomorphologists do not accept the idea of peneplanation. Hack (1960) proposed an alternative model for the evolution of Appalachian topography. His theory of dynamic equilibrium maintains that "....the landscape and the processes molding it are considered a part of an open system in a steady state of balance in which every slope and every form are adjusted to every other. Changes in topographic form take place as equilibrium conditions change, but it is not necessary to assume that the kind of evolutionary changes envisioned by Davis ever occur." Differences in topography are thus explainable in terms of differences in the erodibility of bedrock (Ashley, 1935). Using this model, the multi-level landscape of Nittany Valley is attributed to differences in bedrock erodibility rather than peneplanation. The Tuscarora and Oswego stand as ridges above the valley floor because they are more resistant. Furthermore, accordant summits of the Tuscarora ridges are a result of the nearly uniform resistance of that formation. Water gaps are located along zones of structural weakness.

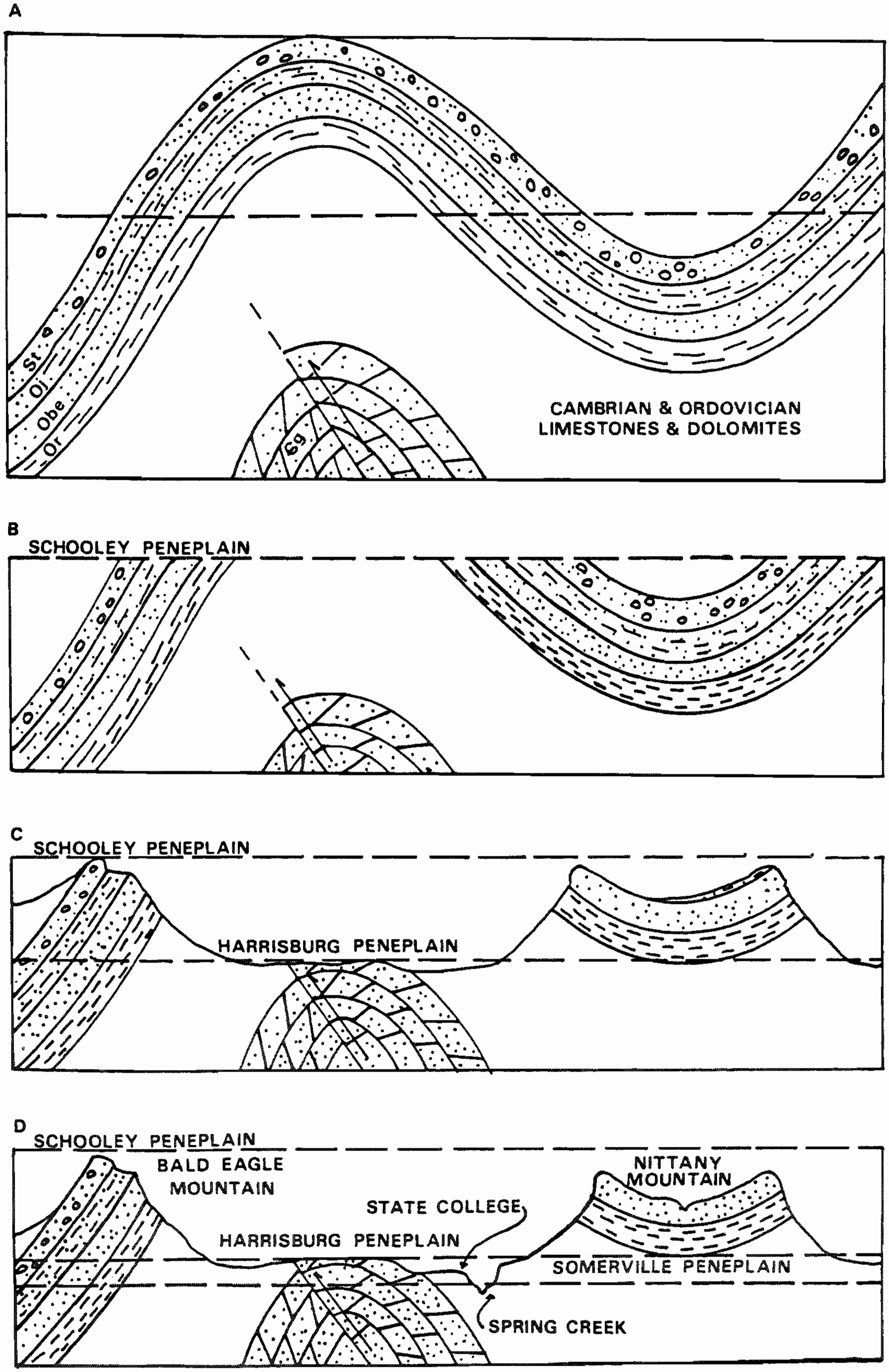

Fig. 5.9 Evolution of peneplain surfaces in the Nittany Valley area.

Periglacial Features

One final aspect of the geomorphology of Nittany Valley should be mentioned at this point. Hillslopes that bound major ridges of Nittany Valley are mantled with colluvial deposits of varying texture, thickness, and extent (Fig. 5.5). The extensive debris mantle, covering 27 percent of the surface area, consists of a poorly sorted matrix of clay, silt, and sand that support large angular to subangular blocks of tuscarora and Oswego Sandstone. It varies in thickness from less than one foot to more than 100 feet. At the mouths of gaps in the Oswego complex alluvial fans extend up to one mile from hillslopes. Since fan deposits are too stony to support agricultural activity, they produce a good stand of timber that extends as a wedge from the hillslopes into the carbonate valley. This makes identification from aerial photographs very easy.

Studies within the past thirty years hypothesized that the colluvial mantle is a relict or 'fossil' landform. It was proposed that colluviation occurred as a result of periglacial conditions during the Pleistocent (Peltier, 1949; Denny, 1956; Ciolkosz, 1978). Wilson (1969) defines a periglacial climate as one with a temperature range of -12 to +3°C and a precipitation range of 50 to 1250 mm/year (ET, EM and Dc climates of Koppen). In the periglacial climate running water, frost action and gelifluction are the dominant physical processes of erosion.

Several lines of evidence support a periglacial origin of the colluvial mantle. The colluvium typically exhibits features that are forming in modern periglacial climates (Embleton and King, 1975). These features include various types of patterned ground (frost polygons (Trout, 1971) and stone stripes), grezes litees (Sevon and Berg, 1979), boulder fields (Rapp, 1967) and gelifluction deposits (Ciolkosz, 1978). The term gelifluction is preferred to solifluction because it is restricted to periglacial environments. Intensified freeze-thaw activity during cooler glacial maxima was thought to have accelerated mechanical disentegration of ridge forming sandstones, supplying material for gelifluction deposition.

In modern periglacial environments gelifluction is most dramatic in permanently or seasonally frozen sub-surface layers. Water released from the active layer during a thaw cannot penetrate below the subjacent frozen layer. The effect of excess water in the active layer is to reduce its internal friction and cohesion. Significant gelifluction probably occurs at moisture values corresponding to or exceeding the Atterberg liquid limit (Washburn, 1973) when soils have little effective strength. Such material can flow on slopes as gentle as 1°. Rapid and significant displacement can occur on slopes in excess of 5°.

Obviously other factors such as soil texture, slope angle and vegetation play a local modifying role. Vegetation, especially deep rooted vegetation, acts as an important restraining factor. Displacement of the boreal forest by tundra encroachment during glacial maxima as suggested by Martin (1958) would certainly favor gelifluction deposition by removing the stabilizing forest cover. Studies of oxygen isotope ratios from speleothems (Harmon, et al. 1978) and refinement of Wisconsinan pollen stratigraphy (Stingelin, 1965) confirm significant temperature depression and vegetation zone shifts during glaciation. However, tundra conditions during the Pleistocene have not been conclusively established beyond the glacial boundary in central Pennsylvania.

A second line of evidence that supports a Pleistocene age rests on the fact that colluvial slopes appear to be stable today. Trees on colluvium are undeformed and few blocks show evidence of recent movement. Weathering rinds and lichen growth show marked asymmetric distributions on boulders. Fragipans and argillic horizons in the colluvial soils indicate landscape stability. Rapp (1967), Troutt (1971), and Ciolkosz (1978) concluded that the colluvial slopes have udnergone insignificant modification since the Pleistocene.

There may be, however, hidden instability in the colluvial slopes. Field data together with simulation models (Carson and Kirkby, Chapt. 10, 1972) suggest that hillslopes tend to become convex when modified by solifluction or gelifluction deposition. On the other hand, hillslopes modified by channelized flow tend to be concave in longitudinal profile. This suggests that modern stream processes will tend to reshape the former convex, gelifluction profile. Vegetation may act as a stabilizing agent, keeping the hillslopes in a metastable state. Deforestation could be the trigger for significant modification of the colluvial slopes.

It should be noted that Lougee (1950) and Hack (1960) among others contend that colluviation is presently active; the result of modern, humid, temperate processes. It appears to the author that the slope is still active today, but at a greatly diminished rate than was experienced during the Pleistocene.

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Distribution and Genesis of Soils of the Northeastern United States*

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Abstract

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The soils of the northeastern United States (central and northern Appalachians) north of the glacial border are Spodosols, Inceptisols, Alfisols, and some Entisols and Histosols. These soils are relatively young (12-18 ka). Climate (frigid soil temperatures), parent material (acidic, sandy), and vegetation (conifers) are the major soil forming factors affecting the genesis of the Spodosols, while parent material (calcareous material) is a determining factor affecting the genesis of the Alfisols. The Inceptisols of southern New England are slightly different (better developed cambic horizon) than those of the Appalachian Plateau. The genesis of the Alfisols of the unglaciated Appalachian Plateau and Triassic Lowlands are also significantly influenced by carbonates and basic minerals in their parent material. Ultisols are found only south of the glacial border. Ultisols on the Appalachian Plateau and in the northern Ridge and Valley are young soils and should be identified as parent material Ultisols, while those of the southern Ridge and Valley, Piedmont and Coastal Plain show significant soil development and should be identified as genetic Ultisols. The soils south of the glacial border, on steep slopes, have been affected significantly by erosion during the cold cycles of the Pleistocene, while soils on gentle slopes have been affected to a lesser degree. Much of the material eroded during the Late Pleistocene has accumulated on lower sideslopes as colluvium. The erosional and depositional processes have resulted in a complex mosaic of moderate to well developed soils on the more stable landscape surfaces, and moderate to weakly developed soils on the erosional and depositional surfaces.

Introduction

The distribution and genesis of the soils of northeastern United States (central and northern Appalachians) is a complex subject. To facilitate the presentation of this material, a section on soil classification and soil genesis precedes the discussion of the distribution and genesis of the soils of the region. In addition, only the major soils of the region will be discussed.

Soil classification

The soil classification system used in this presentation is Soil Taxonomy (Soil Survey Staff, 1975). This system has been the official soil classification system of the United States

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TABLE 1

Soil orders, formative elements (order, suborder, and great group), and a brief description of the main characteristics of the formative elements of the major soils of the Northeast region

Formative element (order)	Description
Order	
Ent (Entisol)	Weakly developed (young) soil without a B horizon, A/C profile or if very sandy a weakly developed color B horizon
Ept (Inceptisol)	Weak to moderately well developed (relatively young) soil with a color and/or structural B horizon (cambic) or a fragipan B horizon
Alf (Alfisol)	Moderately well developed soil with a B horizon of illuvial clay accumulation (argillic horizon) and relatively high subsoil base (Ca+Mg+Na+K) saturation status (>35% of the cation exchange capacity), can also have a fragipan B horizon
Ult (Ultisol)	Moderately well to very well developed soil with a B horizon of illuvial clay accumulation (argillic horizon) and low subsoil base saturation status (<35% of the cation exchange capacity), can also have a fragipan B horizon
Od (Spodosol)	Weak to moderately well developed, sandy soil with a spodic B horizon (subsurface horizon of humus, Fe and Al accumulation) can also have a fragipan B horizon
Ist (Histosol)	Soil composed of organic materials (peats and mucks)
Suborder and great group	
Aqu	Wet-aquic moisture regime (somewhat poorly and poorly drained)
Dystr	Low base saturation (acid) soil
Eutr	High base saturation soil
Fibr	Composed of mostly undecomposed plant material (peats)
Fluv	Composed of recent alluvium (on floodplain)
Fol	Freely drained organic soil
Frag	Fragipan-dense impermeable subsurface B horizon
Hem	Composed of partially decomposed plant material (peaty muck)
Hum	High organic matter content
Ochr	Thin, light colored, A horizon
Orth	Most common type of profile
Psamm	Sand or loamy sand texture
Sapr	Composed of mainly decomposed plant material (muck)
Ud	Humid climate
Umbr	Thick, acid, dark colored A horizon

since 1965. Soil Taxonomy has six hierarchal levels (order, suborder, great group, subgroup, family, and series) and a soil can be discussed or mapped at any of the levels. At the order level the soil name is made up of a specific formative element joined to the suffix-sol (Latin-solum, soil), e.g., Incepti-sol, Alfi-sol, and Ulti-sol (Table 1). At the suborder through subgroup levels, other formative elements are added to the order formative element to give these categories distinctive names, and each of the formative elements provides a significant amount of information about the soil. The criteria used for classification in the upper categories of Soil Taxonomy were chosen to indicate the genetic pathway of soil development (Smith, 1983). Thus, Soil Taxonomy is an appropriate tool to use in discussing the distribution and genesis of soils of a region. Additional information on the structure of Soil Taxonomy is presented by Soil Survey Staff (1975; 1987), Buol et al. (1989), Birkeland (1984), and Witty and Arnold (1987).

Soil genesis

Definition of soil

The soil is an unconsolidated three-dimensional natural body that mantles the landscape. The lateral boundaries of the soil are bedrock or water too deep to support the growth of rooted plants. Its upper boundary is the atmosphere and its lower boundary is the rooting depth of the natural vegetation of the area or the depth to which the soil forming reactions have significantly altered the underlying material. The characteristics and properties of soils are the result of a complex interaction of physical, chemical and biological reactions (soil forming reactions). The rate and extent to which these reactions proceed is controlled by the soil forming factors (climate, organisms, parent material, topography, and time).

Soil forming factors

Introduction

The effect of the various soil forming factors is not equal on the development of all soils. Table 2 gives an estimate of the relative influence of the soil forming factors on soil formation in the Northeast. This relative rating is only applicable for comparisons between soils within the Northeast and should not be applied to soil of other regions.

Time

As indicated in Table 2, time is one of the major factors influencing soil genesis in the Northeast. The northern half of the region was glaciated (Fig. 1), and the deposits in this area vary in age from about 18 ka in Pennsylvania to about 12 ka in northern New York and northern New England (Mickelson et al., 1983). Since deposition of the glacial materials, only minor changes in the landscape of this area have occurred (Schafer and Hartshorn, 1965; Sevon, 1985). Thus, the age of these deposits can be taken as the age of the soils of this area.

The age of the soils south of the glacial border is a more complex issue, due to a complicated landscape evolutionary history. Although no attempt will be made to discuss the long-term geomorphic history of this area, a brief discussion is needed of the most recent landscape history. Because of its location and the reported occurrence of periglacial features in the area (Ciolkosz et al., 1986b; Clark and Ciolkosz, 1988), the most appropriate model to describe these landscape relationships appears to be a combination of those presented by Budel (1982) and Osterkamp et al. (1987). Budel's model developed in Europe in an area analogous to the Northeast indicates that with the exception of slopes of greater than 50 to 60%, today's Northeast landscapes have changed very little in the last 10 ka (Holocene), Budel's model also indicates that Northeast landscapes are the result of periglacial erosion and deposition during the last glacial period and that 95% of current landforms existed prior to the Holocene. The multi-erosion-stability cycle model of Osterkamp et al. (1987) indicates many cycles in the Pleistocene with most soil formation occurring during stable interglacial periods. The combination of these two models into a general geomorphic model provides and overall framework for viewing the soils of the unglaciated area.

The application of this general model to soils of the region requires consideration of the processes than encompass periglacial erosion, in particular the degree of landscape truncation with steepness of slope. According to Budel (1982) and Goldthwait (1976), little down-slope mass movement occurs on slopes of less than about 3% in a periglacial environment. They also indicate that some down-slope movement occurs on slopes of 3–10% and significant movement occurs where slopes exceed 10%. This indicates that soils on gentle slopes would persist, although possibly cryoturbated during a glacial period (Goodman, 1953), while soils on steeper slopes would be truncated on upper slopes with a significant amount of the truncated material accumulating on the lower slopes as colluvium. The larger quantity of colluvium on the footslopes of the region attests to this process. For example, in the Ridge and Valley area of Pennsylvania a typical county has about 27% of its area covered with footslope colluvium (Ciolkosz et al., 1986b). The magnitude of this process is also illustrated by Pennsyl-

TABLE 2

Relative influence of the various soil forming factors on the development of soils in the Northeastern region

Soil order	Parent material	Time	Topography	Climate	Organisms
Entisols	• • •	•	• •	•	•
Inceptisols	• • •	• •	• •	•	•
Alfisols	• • •	• •	•	•	•
Ultisols (PM) +	• • •	• •	•	•	•
Ultisols (GEN) + +	•	• • •	•	• •	•
Spodosols	• • •	• •	•	• •	• • •
Histosols	• • •	•	• • •	• •	•

• = weak influence • • = moderate influence; • • • = strong influence; PM = parent material; GEN = genetic.

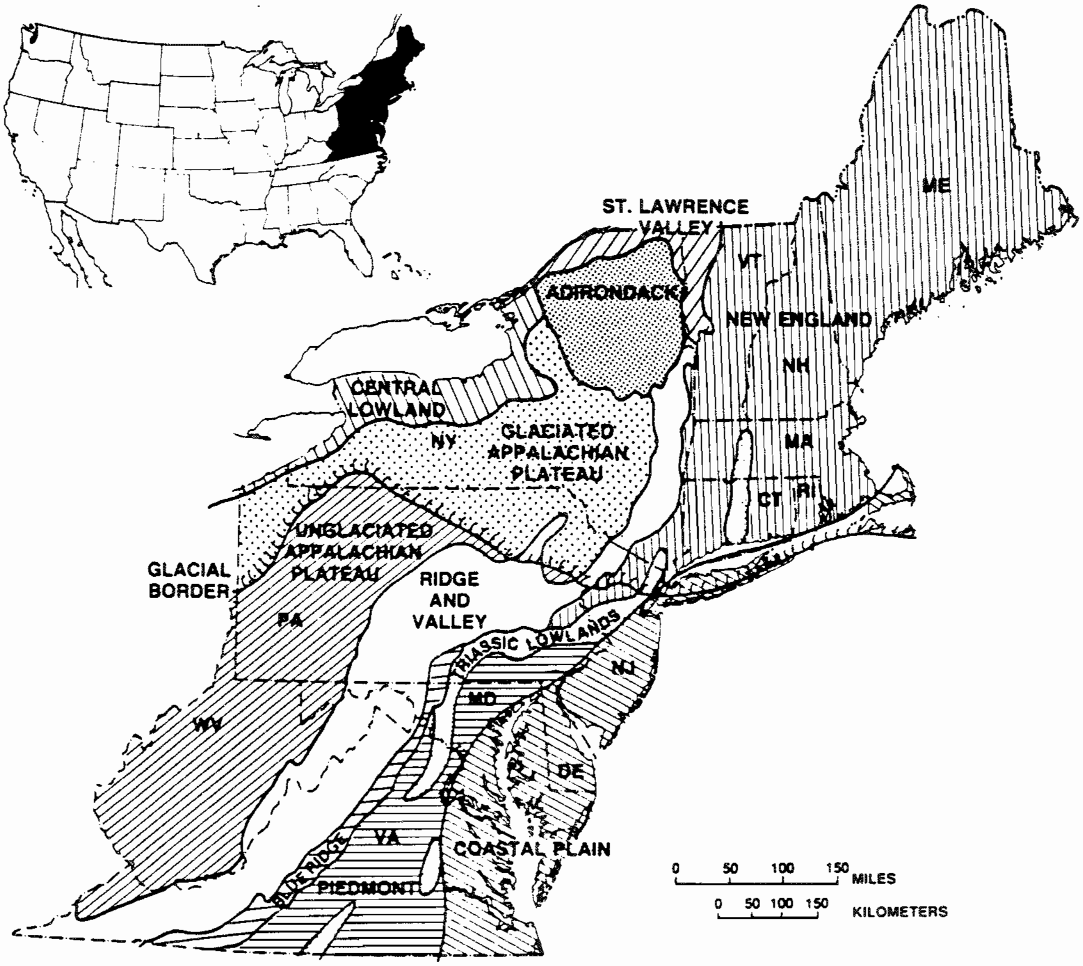

Fig. 1. Physiographic provinces of the northeastern United States. Modified from Cunningham and Ciolkosz (1984).

vania slope data which indicates that only 11% of the state has slopes of 3% or less and about 53% of the state has slopes of 10% or greater (Ciolkosz et al., 1989). In addition to mass movements, fluvial erosion (slope wash) also contributed to denudation during the cold periods (Lewkowicz, 1988), particularly on steep slopes (Budel, 1982). The results of this fluvial process have been observed in areas adjacent to footslopes in central Pennsylvania. In these areas, silty alluvium with a few pea sized sandstone gravels (from adjacent ridges) has filled drainage ways and low areas adjacent to the colluvial mantled footslopes with up to 3 m of material (Ciolkosz et al., 1986a).

Thus, a landscape evolution model involving may glacial–interglacial cycles appears to best explain most of the landscapes of the unglaciated area. An exception may be the Piedmont and Coastal Plain areas of Virginia. In these areas the cycle may still have operated (Whittecar, 1985; Costa and Cleaves, 1984); but, because of less relief and a warmer climate, the landscape may have undergone less severe landscape truncation than in other parts of the region. The Piedmont and Coastal Plain areas of Virginia today have a thermic soil temperature regime (Fig. 2), and the present-day mesic-thermic soil temperature boundary is thought to mark an approximate zone that separates an

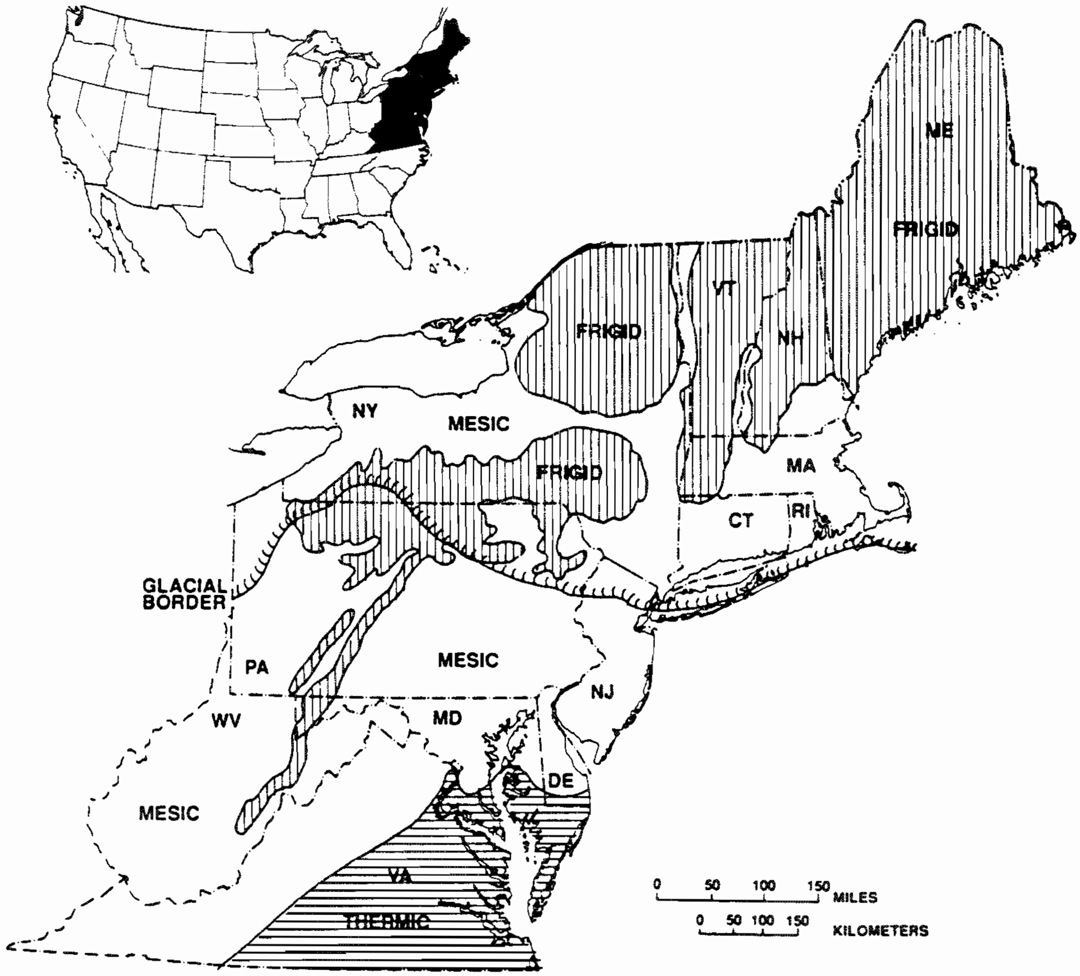

Fig. 2. Soil temperature regimes of the northeastern United States. Frigid = 0–8°C mean annual soil temperature (MAST). Mesic = 8–15°C MAST and Thermic = 15–22°C MAST. Modified from the map of Smith (1984), and Carter and Ciolkosz (1980).

area to the north that has undergone a higher degree of soil truncation than the area to the south (Ciolkosz, 1986b). In the Ridge and Valley area, Clark et al. (1989) reported that a typical county in southern Virginia has about half as much footslope colluvium as a typical county in central Pennsylvania. These data also indicate less landscape truncation in the southern part of the area than in the northern part.

Thus, the age of the soils in the unglaciated area may vary from old on stable parts of the Coastal Plain and Piedmont (Pavich, this volume) to an age similar to that of the glacial soils on truncated bedrock surfaces, in footslope colluvium and on low alluvial terraces. Although some of the soils may be old, many have had the upper part of their profiles partially truncated or altered by frost processes (Cronce, 1988). In addition, some of the partially truncated soils have had colluvium and/or aeolian deposits added to their surfaces creating a welded profile in which the present day soil forming processes have reached through the added material into the underlying truncated paleosol (Hoover, 1983; Waltman, 1985; Ciolkosz et al., 1988; and Darmody and Foss, 1982).

Parent material

Parent material is also a very important soil forming factor in the Northeast region. The glacial deposits in New England and northern New York are derived primarily from granites and schists and are acid materials with sandy loam textures. In central New York, the deposits are derived from sandstone, shales, and limestones. In southern New York and Pennsylvania the glacial deposits are also derived from sandstone and shale but little, if any, limestone. The texture of these materials are loam to silt loam, and their reaction is calcareous in central New York and northeastern Pennsylvania. The parent materials in the area south of the glacial border are more diverse than north of the border. The bedrock of the Unglaciated Plateau is sandstone and shale with exception of southwestern Pennsylvania and northwestern West Virginia, which has in addition to sandstone and shale, interbedded limestone and calcareous shale. The bedrock of the Ridge and Valley area is primarily sandstone in the ridges and shale and limestone in the valleys. The bedrock of the Blue Ridge and Piedmont is primarily granites and schists. Inset into the Piedmont are areas of Triassic lowlands with red shales and sandstones which have been intruded by diabase dikes and sills. The Coastal Plain is composed of multilevels of weakly consolidated to unconsolidated acid sands and clays.

Topography and climate

The topographic factor of soil formation encompasses mainly erosional and soil drainage relationships. With the exception of very steep slopes (greater then 50–60%), the present landscape is relatively stable and has not been significantly affected in the Holocene by surface erosion other than during some catastrophic events such as debris flows triggered by unusual precipitation events (Mills et al., 1987; Jacobson, this volume) or by accelerated erosion caused by man's activities. Thus, present day surface erosion is a minor factor while episodic erosion in the past is the major erosion factor that has affected regional soil formation.

The water relations within soils (well vs. poorly drained) is partially determined by the topographic factors of slope steepness and landscape position. Soils on gentle slopes as well as on lower or footslope areas which receive surface run-on as well as subsurface discharge, tend to be more poorly drained than other soils. The type and rate of physical and chemical reactions that occur in poorly drained conditions produces a much different soil morphologically and chemically than develops under well drained conditions. In particular, poorly drained soils tend to be less leached and less weathered than associated well drained soils (Ranney et al., 1974). They also tend to have a thinner solum and a secondary clay mineral suite with a greater abundance of expandable silicates than their well drained counterparts (Ciolkosz, 1967).

Present day climate (temperature and precipitation), particularly in relation to its impact on vegetation, is significant in soil formation in the Northeast. One of the major impacts is the effect of cool temperatures on the development of coniferous vegetation in the northern and higher elevation parts of the region. The coniferous vegetation in conjunction with the cool temperatures has a major effect on the development of Spodosols. In addition, some slight climatic variations in the recent past in the Northeast have been important. It has been reported that the warmer and drier climate in the period 4–8 ka was responsible for the occurrence of tall grass prairies and their associated soils (Mollisols) in parts of Pennsylvania (Waltman, 1988). This climate trigger may also be responsible for the occurrence of prairie soils in other parts of the region (Table 3). Earlier climate changes and their impact on soil formation are more difficult to evaluate. Any soil that is mid or early Pleistocene in age has undergone many episodes of temperature and precipitation change, landscape stability change, as well as vegetative change (Delcourt

TABLE 3

Percentage of soil orders and suborders in the northeastern U.S. by region and by state (data from Ciolkosz and Dobos, 1989)

Soil order and suborder	Northeast region	States												
		ME	NH	MA	RI	CT	VT	NY	PA	NJ	DE	MD	WV	VA
Entisols	5.79	0.19	0.19	0.75	0.07	0.24	0.22	1.21	0.80	0.62	0.15	0.27	0.38	0.70
Aquents	2.00	0.11	0.02	0.04	**	0.02	0.05	0.40	0.46	0.16	0.01	0.15	0.12	0.43
Fluvents	0.32	-	**	0.03	-	0.01	0.03	0.05	0.05	**	-	0.02	0.04	0.10
Orthents	1.56	0.08	0.08	0.24	0.06	0.17	0.08	0.27	0.27	0.02	-	**	0.19	0.10
Psamments	1.92	**	0.09	0.44	0.01	0.04	0.06	0.49	0.02	0.44	0.14	0.10	0.03	0.06
Inceptisols	36.90	3.44	1.03	1.87	0.36	1.76	1.10	11.67	7.94	0.47	0.03	0.68	3.72	2.83
Aquepts	8.35	2.82	0.28	0.34	0.06	0.21	0.33	3.08	1.02	0.12	0.02	0.02	0.01	0.05
Ochrepts	28.51	0.62	0.75	1.53	0.30	1.55	0.77	8.58	6.92	0.35	0.01	0.66	3.68	2.78
Umbrepts	0.04	-	-	**	-	-	-	0.01	-	-	-	-	0.03	**
Alfisols	14.50	-	-	-	-	-	0.27	5.60	3.71	0.39	0.01	0.38	2.18	1.96
Aqualfs	3.28	-	-	-	-	-	0.10	1.90	1.00	0.05	**	0.02	0.04	0.17
Udalfs	11.22	-	-	-	-	-	0.17	3.70	2.71	0.34	0.01	0.36	2.14	1.79
Ultisols	25.73	-	-	-	-	-	-	0.01	6.28	1.37	0.62	2.59	3.79	11.07
Aquults	2.51	-	-	-	-	-	-	-	0.61	0.12	0.24	0.58	0.05	0.91
Udults	23.22	-	-	-	-	-	-	0.01	5.67	1.25	0.38	2.01	3.74	10.16
Spodosols	15.28	8.92	2.41	0.49	-	**	2.23	0.97	0.07	0.16	**	0.02	0.01	**
Aquods	0.53	0.14	0.04	0.01	-	**	0.06	0.12	-	0.16	**	0.01	-	**
Orthods	14.74	8.78	2.37	0.48	-	-	2.17	0.85	0.07	**	-	0.01	0.01	**
Histosols	1.63	0.47	0.14	0.18	0.02	0.05	0.07	0.32	0.01	0.13	**	0.16	**	0.08
Fibrist	0.15	0.14	-	-	-	-	**	**	-	-	-	-	-	-
Folist	0.06	0.01	0.05	-	-	-	**	**	-	-	-	-	-	-
Hemist	0.51	0.16	0.09	0.04	**	0.01	**	0.05	-	**	-	0.13	-	0.02
Saprist	0.91	0.15	-	0.14	0.02	0.04	0.07	0.27	0.01	0.13	**	0.03	**	0.06
Mollisols	0.17	-	-	-	-	**	**	0.09	0.03	**	-	0.01	0.03	0.01
Aquolls	0.08	-	-	-	-	**	-	0.07	0.01	**	-	**	**	**
Udolls	0.09	-	-	-	-	-	**	0.02	0.02	-	-	0.01	0.03	0.01

** <0.005%; - = none.

and Delcourt, 1983, 1986). Precipitation changes may have in some cases altered the drainage condition of the soil while temperature changes have affected the rate of various soil forming reactions. In addition, some reactions may have been triggered which have temperature dependent thresholds. For example, the presence of red (rubified) soils in Pennsylvania has been attributed to a hotter climate during the Sangamonian interglacial (Waltman, 1985) than at present. The rubification process has been associated with a mean annual temperature of >15°C (Bullock, 1985). The present average temperature in State College, Pennsylvania (Central Pa), is 10°C. If Bullock (1985) is correct, the mean annual temperature was about 15°C in central Pennsylvania during the Sagamonian, and this region has experienced a temperature variation of from 15°C (Sagamonian) to −5 to −10°C (Woodfordian glacial) (Ciolkosz et al., 1986b) to 10°C today. These type of climatic variations have significantly affected the soil forming reactions, but their effect is too complex and too poorly understood to be discussed here, although the subject holds many interesting opportunities for future research.

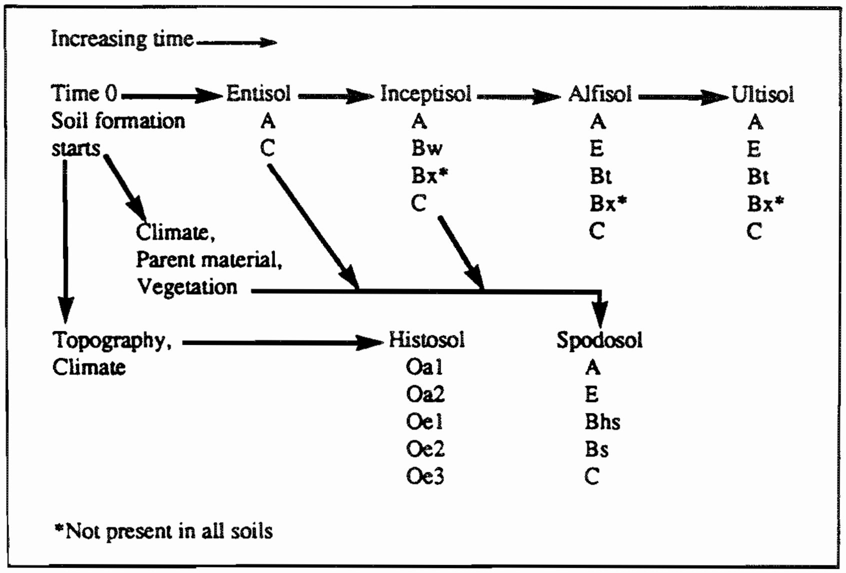

Fig. 3. Idealized soil development sequence for soils of the northeastern United States.

Developmental model

The effect of most soil forming reactions decreases from the surface downward. Thus soil characteristics vary from the surface downward from one zone or horizon to another with the major soil horizons being the A, E, B, and C. Figure 3 gives an idealized developmental sequence for Northeastern soils. This sequence will be used as a conceptual model of soil formation in the following discussion. For additional discussion of the principles of soil genesis see the texts of Birkeland (1984), Buol et al. (1989), Jenny (1980), Wilding et al., (1983a, 1983b), Fanning and Fanning (1989), Tedrow (1977), Rieger, (1983), and the Soil Survey Staff (1975).

Distribution and genesis

Entisols

Entisols are pedologically very young soils (Grossman, 1983) and are found mainly on the Coastal Plain (Fig. 4) in the glaciated area, and on floodplains throughout the region. They are developed in well drained, very sandy glacial or Coastal Plain deposits (Psamments: 60% are in the glacial area and 40% are in the Coastal Plain area), in poorly drained Coastal Plain or flood plain deposits (Aquents: 80% are on floodplains and 20% are on the Coastal Plain) and in well drained glacial fluvial or man-disturbed areas such as strip mines (Orthents). Only the Psamments occur as extensive soil areas. Although the Aquents and Orthents cover about the same total area as the Psamments (Table 3) they occupy smaller individual areas intermixed with other soils and can not be shown in Fig. 4. The Entisols of the region show very weak development because: (1) their parent material is of recent origin (glacial material or floodplain deposits); and/or (2) they are saturated with water most of the time; and/or (3) they are very resistant to change because of sandy texture and quartzose mineralogy. In Pennsylvania, the floodplain soils which are mainly silt loam texture have been dated by Bilzi and Ciolkosz (1977) and range between a few hundred to a few thousand years old. In this span of time, the well drained soils have developed into Dystrochrepts and Eutrochrepts while the poorly drained soils are classified as Fluvaquents. This relationship generally holds for the entire re-

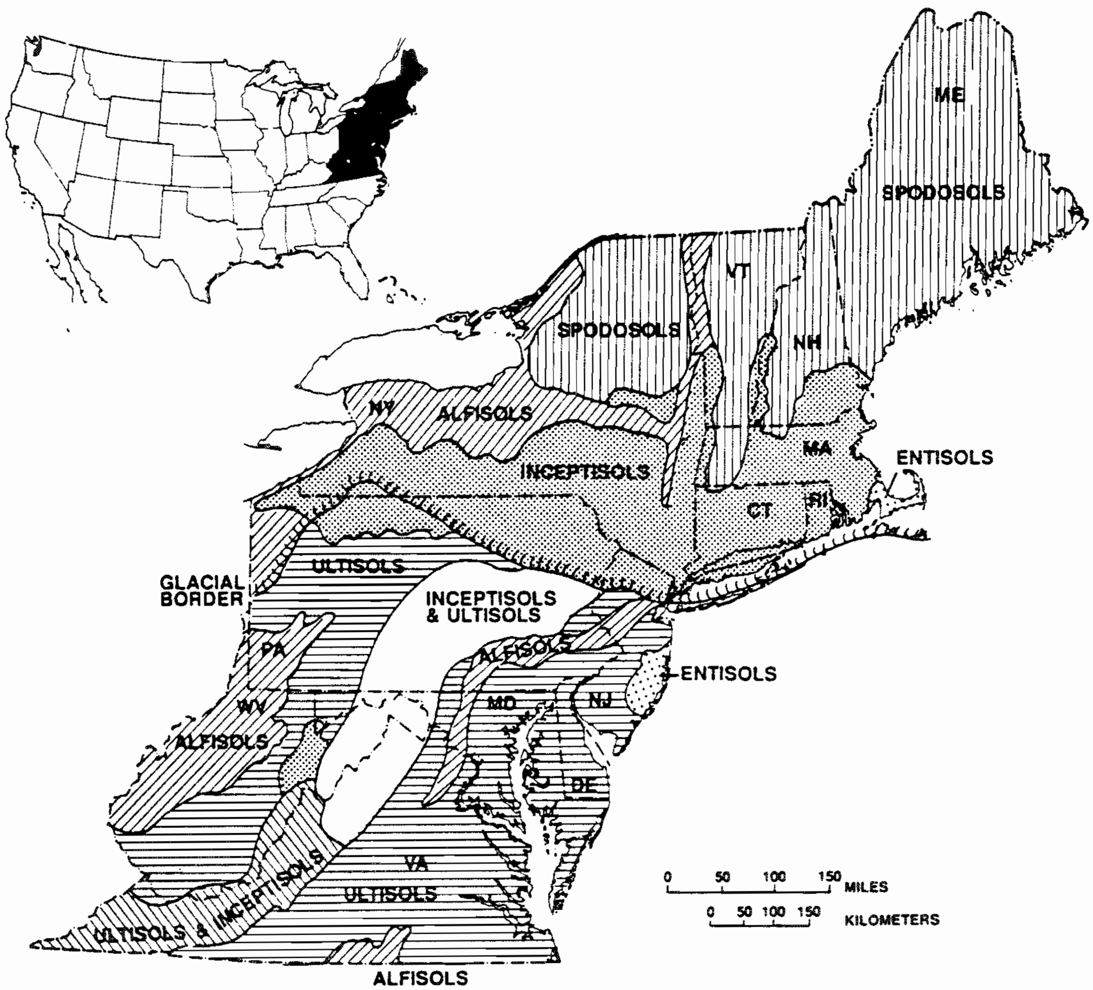

Fig. 4. Soils of the northeastern United States. Modified from the map of Smith et al. (1984).

gion. With additional time and landscape stability many of the Entisols of the region will develop into Inceptisols, Alfisols or possibly Ultisols (Fig. 3).

Inceptisols

Inceptisols are pedologically more developed than Entisols, and they have cambic and/or fragipan B horizons. They are found in glacial deposits in southern New England, southern New York, and northeastern Pennsylvania and intermixed with Alfisols and Ultisols in the unglaciated Ridge and Valley area from Pennsylvania through Virginia (Fig. 4).

The Inceptisols of southern New England, with the exception of those in the Connecticut river valley, have developed in glacial till derived primarily from igneous and metamorphic rocks; are acid; have fine sandy loam to sandy loam textures; and have dense subsoil horizons or layers (Soil Survey Staff, 1968). These Inceptisols (Dystrochrepts), which were previously classified as Brown Podzolics (Lyford, 1952), have a well-expressed color Bw cambic horizon. These color B's are the result of deposition in the B horizon of Fe and some humus which has been eluviated from upper horizons (Flach, 1960; Lyford, 1952; Baur and Lyford, 1957). Weathering of the crystalline rock minerals in the glacial materials of these young soils (about 18 Ka; Mickelson et al., 1983) under the climatic conditions of this area apparently has not caused the release of very much iron within the B horizon to form the more typical cambic color Bw horizon. The lack of weathering of the parent material is also indicated by the lack of illuvial clay in these soils. In order to accumulate clay in the B horizon, the clay, particularly the fine clay, must be present in the parent material or it must be produced by weathering. The material in the upper part of the soils in many areas of southern New England underwent sorting and redistribution by wind shortly after deposition (Schafer and Hartshorn, 1965; Colby et al., 1953). This sorting has given many of these soils very similar textures (fine sandy loam) particularly in the upper part of the soil profile. The dense subsoil horizons in the northern and southern New England soils and in the soils of Adirondacks of New York have in the past been identified as fragipans. Recent studies have questioned whether these dense horizons are true genetic fragipans or just dense basal till (Lindbo and Veneman, 1989), and the soil survey of the USDA-Soil Conservation Service has reclassified the majority of the fragipan soils in New England and the Adirondacks as non-fragipan soils. This reclassification may have been premature, particularly for the Inceptisols of southern New England (Lindbo and Veneman, 1989) and a more comprehensive study is needed to resolve this issue.

The Inceptisols (Fragiochrepts and Fragiaquepts) of the glaciated Appalachian Plateau have developed from glacial till derived from acid sandstones and shales. These materials have imparted a silt loam to loam texture to these soils as compared to the coarser sandy loam textures of the Inceptisols of New England. These soils were previously classified as Sols Bruns Acides (Baur and Lyford, 1957; Flach, 1960) and do not have as strong a degree of cambic color B horizon development as do the the New England Inceptisols. The reason for this difference is the Appalachian Plateau Inceptisols have little illuvial Fe and humus, and their color is obtained primarily from limited in situ mineral weathering in the B horizon. Although the cambic horizon is not well developed, the fragipan in these till-derived soils shows very strong development and is recognized as a true genetic fragipan (Lindbo and Veneman, 1989). Historically, these fragipans form the core of the fragipan concept developed by Dr. Marlin Cline and his students at Cornell University. In addition to having formed in glacial till, fragipans also have formed in other transported materials that are not too sandy, too clayey, or calcareous. These materials include loess, colluvium, and old alluvium, but not recent alluvium (Ciolkosz et al., 1979). Movement of the material apparently enhances packing of the soil particles giving the fragipan its characteristic high bulk density. Fragipans are not found in recent alluvium on floodplains but they are found on older fluvial and outwash terraces above the floodplains, and it has been estimated that it takes more than 2–5 ka for them to form (Hall et al., 1982). They also are not permanent fixtures in soils. Fragipans in well drained glacial till soils form and are then degraded by oxidation, leaching, and argillic horizon formation. This conclusion is based on their presence in Woodfordian age till soils in northeastern Pennsylvania and their absence in Pre-Wisconsinan Age tills soils just outside the Woodfordian border (Ciolkosz et al., 1985). Supporting evidence comes from Louisiana where fragipans formed in loess under more rigorous weathering conditions are now degrading (Bartelli, 1973; Miller, pers. commun., 1985). Fragipans persist in poorly drained soils of the Pre-Wisconsinan till catena (Waltman, 1981). Apparently leaching and argillic horizon formation have not proceeded to the point of destroying the fragipan in the more poorly drained soils, although it is found deeper in the profile and is less dense than in Woodfordian age soils

(Waltman, 1981). With adequate landscape stability these fragipans may degrade following the pathway noted by Steele et al. (1969) for fragipans on the North Carolina Coastal Plain. Smeck and Ciolkosz (1989) give an up-to-date presentation on recent work done on fragipans.

The Inceptisols (Dystrochrepts) of the unglaciated Ridge and Valley area are developed on the sandstone ridges and in the shale valleys. These soils have cambic B horizons without fragipans. Both ridge tops and shale valleys underwent significant erosion during the late Pleistocene. The erosion and the resistance of the parent material to pedological development has resulted in weakly developed soils on these landscapes (Ciolkosz et al., 1986a; Carter and Ciolkosz, 1986; Carter, 1983). In particular, soils that developed on acid brown and gray shales show very weak development as indicated by profiles which are shallow or moderately deep to bedrock and contain a high content of rock fragments. In contrast to the grayish brown shales, red shales on the same landscape weather more rapidly giving deeper soils with fewer rock fragments and weakly expressed argillic horizons (Clark et al., 1989). Soils formed from calcareous shales tend to be deeper and better developed than soils formed from acid shales. This apparently results from rapid leaching of carbonate cement which then releases a relative large mass of rock material to soil development. An example of this relationship occurs near Burnham, Pennsylvania (central Ridge and Valley area) where the vertical bedrock comprises and alternating sequence of thin (2–5 m thick) acid brown shale and weakly calcareous shale beds. The depth to the hard bedrock (R horizon) is about 1 m in the acid shale and 3–4 m in the calcareous shale.

Alfisols

In an idealized sequence of development, Alfisols are a step more developed than Inceptisols (Fig. 3). They have argillic horizons and a relatively high subsoil base saturation. They are found in the glaciated areas of central New York and northwestern Pennsylvania, as well as in the unglaciated areas of southwestern Pennsylvania and northwestern West Virginia. Alfisols are also found in the Triassic lowland area of southeastern Pennsylvania and the limestone valleys of the Ridge and Valley area in Pennsylvania, West Virginia, and Maryland.

The occurrence of Alfisols in central New York forms an apparent anomaly because they are younger in age (about 12 ka) than the Inceptisols found in southern New York and northeastern Pennsylvania (about 18 ka). However, these Alfisols have developed from calcareous glacial deposits, and the carbonates have significantly enhanced the development of these young soils. As the carbonate is dissolved and leached from the upper horizons during the development of these soils, the clay, which is an insoluble residue, is released and eluviated from the A and accumulates in the B horizon forming an argillic horizon. The depth of the clay accumulation represents the depth of movement of water carrying the clay or a zone of clay flocculation. The work of Cronce (1988) indicates that a base (Ca+Mg) saturation of 40% or greater forms a threshold for the flocculation of clays in argillic horizon formation. Thus calcareous parent materials, which are very susceptible to dissolution, foster argillic horizon formation, even in young glacial deposits (Rostad et al., 1976; Smeck and Wilding, 1980). The calcareous drifts of central New York vary in their carbonate contents from high ($>25\%$) to medium (5–14%) to low ($<5\%$). Argillic horizon development parallels this sequence with the most developed argillic horizons having the highest carbonate content (Merritt, 1958). Much of the clay that is dispersed in the upper horizons of an acid (low base saturation) soil may be moved completely out of the soil by the percolating water. The reason argillic horizons have not formed in the Inceptisols of southern New York is probably twofold. Firstly, the acid shale and sandstone till parent material is not releasing much clay which can be dispersed and moved; and secondly, some of the clay that is being moved may be completely removed from the soil by percolating water. The first reason is probably dominant because these soils have fragipan B horizons which do not allow movement of much material. Although the soil developed in the calcareous till are Alfisols they are not strongly developed soils, and have thin solums (A + B horizon) that are only 60–75 cm thick, with weakly expressed argillic horizons.

The Alfisols of northwestern Pennsylvania are also developed in glacial till. These Alfisols have marginal argillic horizons and are close to being Inceptisols. The till of northwestern Pennsylvania has a lower carbonate content (2–3%; White et al., 1969) than that in central New York, and this is probably the reason for the weakly developed argillic development in these soils. In addition to the weakly developed argillic horizons, these soils also have well expressed fragipans (Ciolkosz et al., 1989).

The Alfisols in southwestern Pennsylvania and northwestern West Virginia have developed from interbedded shale, limestone, and calcareous shale bedrock. These parent materials are one of the major factors controlling the development of argillic horizons in these soils as well as giving them a high base status. These soils show somewhat thicker argillic horizons that the calcareous till soils, but they are still only weak to moderately well developed, which is apparently a reflection of their development on recently truncated landscapes (Ciolkosz et al., 1989). The amount of truncation on these landscapes, as well as others in the unglaciated area varies directly with slope. Also, either a significant amount of rock disaggregation predated the last truncation or frost churning processes have contributed significantly to the disaggregation of the bedrock. The importance of these two processes needs further study to determine their relative importance on the development if these residual soils.

The Alfisols of the Triassic lowland area are developed from red shale, diabase, hornfels, and some limestone. The presence of deep, well drained soil on the diabase of southeastern Pennsylvania indicates that landscape truncation over time has been a complex process because these soils are too thick to have developed since the last glacial period. Most of the Alfisols in the northern Ridge and Valley area are formed on limestone. These Alfisols, which can be many meters thick, occur on gently rolling landscapes and may represent some of the oldest soils in the Northeast. An estimate based on the accumulation of insoluble residues for one meter of soil material indicates a 300 ka age (Ciolkosz et al., 1986a). Similar results were obtained by Cronce (1988) using additional indicators such as quartz and Fe_2O_3 content. Cronce also indicates that aeolian material added to these soils complicates the interpretation of their genesis, but helps to account for the very common silt loam texture of their surface horizons. Cronce's work also indicates that the upper part of the limestone soils has been significantly affected by frost processes giving them a very uniform upper profile horizon sequence.

Ultisols

All the Ultisols of the region are found in the unglaciated area (Fig. 4). They are found on the Appalachian Plateau, the Piedmont , the Coastal Plain, and intermixed with Inceptisols in the Ridge and Valley area. These Ultisols can be grouped into two categories: parent material Ultisols and genetic Ultisols (Clark et al, 1989). Parent material Ultisols show weak development in residual acid shale materials on the Appalachian Plateau and in the Ridge and Valley and weak to moderate development in colluvial deposits in the Ridge and Valley area (Ciolkosz et al., 1979). The low base status of these soils results from a very low base status inherited from the parent material and not from extreme leaching. These soils contrast with the well developed genetic Ultisols developed on the igneous and metamorphic rocks of the Piedmont, on the limestones as well as some older alluvium and colluvium in the southern Ridge and Valley, and on the sands and clays of the Coastal Plain. The argillic horizon of genetic Ultisols differs from that found in Alfisols and in parent material Ultisols. In uneroded genetic Ultisols the argillic horizon occurs deeper in the profile, has a more gradual transition from eluvial to illuvial zone, has a thicker zone of accumulation, and has more clay than can be accounted for by eluviation from overlying horizons than is found in Alfisols and parent material Ultisols (Ciolkosz et al., 1985). These differences are a result of a greater period of time for clay to move down the soil profile as well as adequate time for clay to be stripped from the upper part of the argillic and translocation downward. In addition, the longer period of time has allowed clay to form within the argillic horizon by weathering. The age of the soils and pedogenetic trends on the Coastal Plain have recently been studied by Markewich et al. (1987). These studies indicate progressive development from the youngest age material (about 30 ka) to the oldest age material (> 1 Ma). The most highly developed soils in the Northeast region occur on the high Coastal Plain and on nearby high terrace deposits. Solum thicknesses in excess of 5 m and the presence of plinthite horizons have been observed in these soils. An interesting future comparison between the youngest soils of the Coastal Plain and the slightly younger glacial till soils should be made, particularly in relation to the effect of climate on soil development.

The Ultisols formed on the granites and schists of the Piedmont do not show the thick solum (A + B horizon) development that is observed in soils on the middle and upper Coastal Plain on similar landscapes. Although these soils do not have thick solums, they do have very thick C horizons that frequently are called saprolite. This material has been estimated to be about 15 m thick in northern Virginia (Froelich and Heironimus, 1977; Pavich, 1986). The Piedmont is unique in this respect, because no other residual soils of significant extent in the region have thick saprolite C horizons. An example of this relationship occurs in southeastern Pennsylvania where the Piedmont is in close proximity to a long belt of acid shale (Martinsburg) in the great valley north and south of Harrisburg. The shale area and the Piedmont have similar landscapes, but the shale soils lack the thick saprolite C horizons of the Piedmont soils. Thus, parent material appears to be a significant factor in the development of the thick saprolite zones. The thick saprolites and thin solums of the Piedmont soils has been discussed by Pavich (1986). He suggests that the soils on the Piedmont are in a dynamic equilibrium and will always have about the same development that we see today. If this is true, the Piedmont is a very unique pedological system that is not operating in the same manner as the Coastal Plain or other areas of the region. A more plausible explanation is given by Costa and Cleaves (1984), Cleaves (this volume), and Poag and Sevon (this volume) who evoke episodic erosion to explain these relationships.

The Ultisols in the Ridge and Valley of southern Virginia are developed dominantly from limestones. Although developed from limestone, they have undergone a greater degree of leaching of bases and weathering than the limestone soils found in the northern Ridge and Valley which are Alfisols.

Spodosols

Spodosols are found primarily in the northern part of the glaciated area associated with acidic, sandy textured parent materials and frigid soil temperature regimes. Some Spodosols occur in the unglaciated area and are associated with acidic, sandy (quartzose sandstone) parent materials and frigid soil temperatures (Fig. 2). This association also extends southward into the southern Appalachians at higher elevations (Lietzke and McGuire, 1987). The only Spodosols found outside of the frigid areas are poorly drained (Aquods). They occur primarily on the lower and middle Coastal

Plain from New Jersey through Virginia and in glacial outwash plains in Central New York. The temperature and landscape relationships noted for the Northeast also hold for the United States as a whole (Rourke et al., 1988). Some Aquods also occur in the frigid areas but they account for only about 2–3% of the total Spodosol area (Table 3). Although very minor in occurrence, Aquods tend to have the best expressed ortsteins (cemented spodic horizons) (McKeague et al., 1983). In addition to cold temperatures, and acidic sandy texture parent material, coniferous vegetation is also a major factor influencing Spodosol development (McKeague et al., 1983).

Spodosols, previously called Podzols, were originally defined in Russia on the basis of their E horizons, but later the illuvial B horizon (spodic) of humus plus aluminium with or without iron became the definitive horizon. This change is of major significance because some of the soils classified today as Spodosols have little or no E horizon development. Although the reactions involved in spodic horizon formation have been studied for years, a clear understanding of the exact mechanism of their development has not been established. The most feasible mechanism encompasses the complexation of iron and aluminium with organic acids to form soluble complexes. These complexes move downward in percolating water until they are precipitated by increases in the concentration of metal ions, by desiccation or by oxidation of the organic complex (Mokma and Buurman, 1982). This mechanism seem appropriate for Orthods but not for Aquods. Although less is known of spodic horizon genesis in Aquods, some progress is being made in this area of research (Lee et al., 1988).

Climate, particularly temperature, in concert with vegetation appears to be very important in the development of Spodosols in the Northeast. Stanley and Ciolkosz (1981) reported a trend of increasing spodic horizon development with decreasing soil temperature for Spodosols developed from sandstone on the unglaciated Appalachian Plateau. This trend continues into New England and helps explain the genesis of the very well developed spodic horizons which have been called "super spodics" in some Spodosols of northern New England and northern New York. A general trend of decreasing E horizon (Albic horizon) thickness in Spodosols from the unglaciated Appalachian Plateau to New England has also been observed. This trend is believed to be associated with a much higher quartz content in the sand size material of the sandstone of the plateau than in the till of the glaciated area, and a much higher content of weatherable minerals in the till. The sand size quartz provides the whitish mineral surfaces that give E horizons their light color and the weatherable minerals release bases which precipitates the soluble organic complexes near the soil surface (DeConinck, 1980). It also has been speculated that windthrow of trees, which tends to mix the upper horizons, also may contribute to this trend.

Most well drained Spodosols in the northeastern U.S. have a solum thickness of a meter or less, and according to Franzmeier and Whiteside (1963), this amount of soil formation could have taken place in 10,000 years. Thus, the Spodosols in the unglaciated area may be only about 10 ka or they may be older and in a state of equilibrium. If the mechanism of Spodosol genesis proposed by McKeague et al. (1983) is correct, the E horizon, and probably the spodic horizon also should become thicker with age. This mechanism seems to apply in Aquods genesis in tropical and subtropical regions (McKeague et al., 1983). Thus, it appears that the Spodosols of the unglaciated area are only 10–15 ka. This conclusion supports the contention that these areas were eroded during the last glacial period, otherwise thicker, better developed Spodosols should be found in the unglaciated area.

Histosols

Histosols are organic soils and are found throughout the region, but they are more abundant in the glaciated area and on the lower Coastal Plain. The organic matter of Histosols accumulated because its decomposition is retarded by water saturation and/or cold temperatures (Everett, 1983).

Histosols can be divided into two groups: topographic and climatic. Topographic Histosols are found in depressional areas that are saturated with water most of the year. The glaciated area, with its disrupted drainage and many undrained depressions, and the lower Coastal Plain, with its high water table and tidal inundation are major areas of topographic Histosol formation. Climatic Histosols (Folists) are freely drained organic soils. They have formed in climates that are cold and humid. In the northeast region they are of minor extent and are found on moderate to steep slopes at high elevations in northern New England and the Adirondacks. In this region they also tend to form directly on bedrock and are relatively thin (usually <0.5 m thick). Many of the salt water Histosols, when drained, become very acidic due to the oxidation of sulfides that have accumulated with the organic material. Fresh water Histosols do not tend to accumulate sulfides and many have been drained and are used for agricultural purposes.

With drainage, Histosols oxidize rapidly and are transformed from Fibrists to Hemists to Saprists. Under natural conditions this conversion is greatly reduced, although, there is a general trend of more rapid conversion with increasing temperature from northern New York to the southern part of the region.

Conclusions

The soils of the northeastern United States have followed a varied and complex pathway of development. The pathway would not be nearly as complex if the soil forming factors had been constant for all soils of the region. Of particular importance in the region has been the impact of glaciation and associated variations in climate. The impact of glaciation in the northern part of the region has been to create new parent material and set the pedological clock of soil formation at a known time. Since the deposition of the glacial materials, the variation of climate, vegetation, and other soil forming factors has been minor. Thus, the glaciated area provides a less complex area than the unglaciated region and probably explains, why more is known of the genesis of glacial soils than of nonglacial soils.

In the unglaciated area, the pathway of soil formation for the area as a whole is more complex because there are many different ages of surfaces. Consequently, the pedological clock of soil formation has started at different times in different places. One of the major factors causing the complexity in the age of the surfaces is the changes in climate associated with glacial–interglacial cycles and its impact on landscape stability. A particular complication associated with instability and erosion is the complete removal of previous soils on one landscape and partial to nonremoval on other landscapes or slopes. Consequently, soil formation can start in a new soil material, in a partially truncated soil or in a soil that has not been truncated.

In addition to the challenge of learning more about soil genesis in the glacial area is the even larger challenge of sorting out the genetic pathways of soils in the unglaciated area. In particular, what are the relationships of slope under present conditions to past conditions with respect to possible thresholds of erosion, and what impact has varying climatic conditions over time had on the rate, extent, and direction of soil formation reactions. In addition, what can the soil properties observed today tell us about the processes that have shaped the soils and landscapes in the northeastern United States.

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Genesis of soils and landscapes in the Ridge and Valley province of central Pennsylvania

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ABSTRACT

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The characteristics and properties of the soils on the ridge tops, footslopes, and adjacent limestone valley areas in the Ridge and Valley of central Pennsylvania have been strongly influenced by their parent material and geomorphic history. The ridge top soils have developed in sandstone colluvium which mantles sandstone residuum. The upper part of the original residual ridge top soil was truncated during late Wisconsinan time and then covered with local colluvium or it was cryoturbated. These sandstone parent materials have been stable since the late Wisconsinan and have sandy skeletal Dystrochrepts and Haplorthods developed in them. The Haplorthods are minor soils and are associated with local concentrations of coniferous vegetation. During the late Wisconsinan, the sandstone colluvium also moved downslope and was mixed with bedrock and residual material from shale and limestone and deposited on the footslope over a pre-Wisconsinan soil developed in older colluvium or limestone residuum. The footslope surface colluvial soils vary in texture and drainage because of their parent material, their location in discharge areas, and fragipan development. The age of the brown surface colluvium is considered late Wisconsinan and the age of the pre-Wisconsinan buried soils is not known. The buried soil's bright red (rubified) color and argillic horizon indicate a much greater degree of soil development than noted in the brown surface colluvium, and its age may be correlated with isotope stage 6. The soils developed at the surface in the colluvium are mainly Ultisols although some poorly drained soils, particularly in limestone material, are Alfisols. The Ultisols are parent material Ultisols and the poorly drained Alfisols have a high base status in their parent material or were recharged with bases from the groundwater. The soils of the limestone valleys are developed in residuum. The residuum accumulated from the insoluble residues after the $CaCO_3$ was leached from the bedrock. If the residuum is thick (2 to 3 m), the soils date back many hundreds of thousands of years, but are still classified as Hapludalfs. In addition to being old, the surface of the Hapludalfs has been modified by periglacial truncation and the addition of eolian materials. These processes have given the upper 50 cm of the Hapludalfs a late Wisconsinan age but the bulk of the B horizon is many thousands of years older. Eolian material has also been added to the ridge top soils as well as the colluvial soils. The mineralogy of all the soils in this landscape, including the buried paleosols, does not show significant clay mineral weathering. This appears anomalous considering the degree of oxidation of the buried paleosols and the limestone soils and their estimated age.

Introduction

In a recent review on the genesis of soils of the northeastern United States, Ciolkosz et al. (1989) indicate that the soils of the unglaciated part of the region have had a very complex genetic pathway. The general reason given for the complexity is that climatic change during the late Pleistocene and Holocene affected the soil-forming reactions and landscape pro-

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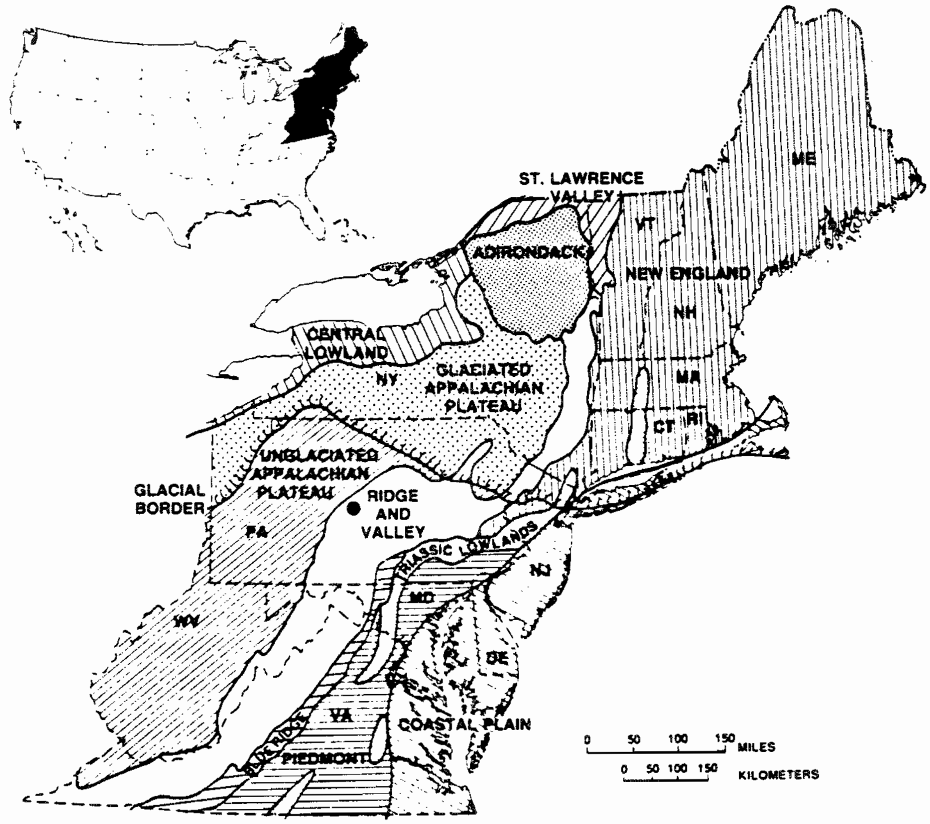

Fig. 1. Physiographic provinces of the northeastern United States (modified from Cunningham and Ciolkosz, 1984.) The large black dot in the northern Ridge and Valley indicates the general location of the study areas of Carter (1983), Hoover (1983), and Cronce (1988).

cesses differentially over time. In their review, Ciolkosz et al. (1989) did not discuss in detail the properties of the soils or their genetic complexity. To partially overcome this shortcoming, an examination of the genesis of the soils and landscape of a typical sandstone ridge, its sideslope, and an adjacent limestone valley in central Pennsylvania will be presented. This presentation is a synthesis of detailed studies which were done on different slopes in the same area (Fig. 1) and represent a recurring soil and landscape condition. Although the detailed studies were done in central Pennsylvania, the results are applicable to the entire unglaciated Northern Ridge and Valley area (*PA* to central *VA*; Fig. 1).

General landscape relations

The major or primary ridges in the northern Ridge and Valley area are underlain by sandstone while the adjacent valleys are underlain by limestone or shale (Fig. 2). The dip of the bedrock in the ridges is usually steep while in the valleys it is frequently not as steep and in many places it is horizontal. The major ridge sideslopes are mantled with colluvium. The ridge tops have a relatively constant elevation of about 650 m, and the valley floors have relief of 30 to 100 m, while the relief between the valley floor uplands and ridge tops is about 300 m. In terms of classical geomorphology, the ridge tops have been described as remnants of

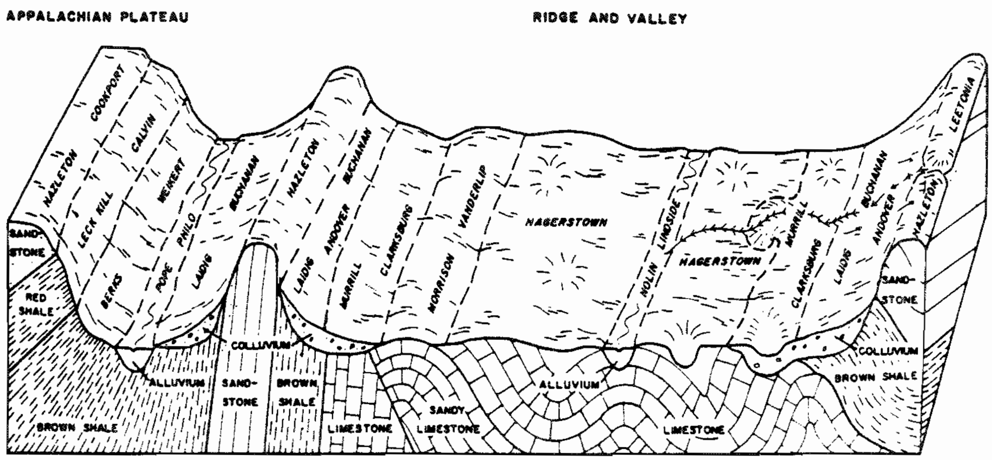

Fig. 2. Generalized diagramatic soil–landscape relations of the Ridge and Valley and adjacent plateau area in central Pennsylvania. (Modified from Ciolkosz et al., 1986a.)

the Schooley peneplain and the uplands of the valley floor as remnants of the Harrisburg peneplain (Sevon et al., 1983). However, recent work (Poag and Sevon, 1989; Sevon, 1989) has cast some doubt on the reality of these surfaces. The limestone valley floors have less relief than the shale valley floors and frequently show no surface drainage patterns. Where there is a major stream on the limestone valley floor, the relief near the stream can equal that of the shale areas. Thus, the limestone areas generally show a more gently rolling landscape with less relief than the shale areas except adjacent to major streams. Complicating these generalizations is a complex bedrock structure which frequently is manifested on the valley floors as long, weakly cemented sandstone, hard shale, or very cherty limestone secondary ridges of low relief parallel to the primary ridges (Ciolkosz et al., 1974).

Ridge top soils

The major soils on the ridge top and shoulder slope areas in central Pennsylvania are deep (> 1 m to bedrock), well drained, acid, sandy, and have a high rock fragment content (Table 1; Fig. 2). They are classified as Dystrochrepts (Hazleton) and Haplorthods (Leetonia) (Carter, 1983; Braker 1981; Ciolkosz and Dobos, 1990).

Parent material and time

The ridge top soils are considered residual (developed from the underlying sandstone or conglomerate). Although considered residual, these soils do not show intensive development. In central Pennsylvania, intensive development is indicated by red high-chroma colors, argillic horizons, and thick solums such as those found in the Allenwood pre-Wisconsinan glacial till soil of the central Susquehanna River valley (Marchand et al., 1978; Levine and Ciolkosz, 1983; Harden and Taylor, 1983). The lack of development in the ridge top soils is most likely related to parent material influences (high quartz content) and a relatively short period of time for soil formation.

The parent material of these soils are rocks of Paleozoic age, and the soils are much younger than the age of the rocks. On these ridge tops the soil extends from the surface to consolidated bedrock and the disaggregation of the rock is a part of the genetic process. The depth to bedrock on the ridge tops according to early soil surveys was primarily 50 to 100 cm, and the Dekalb soil series (Dystrochrept)

TABLE 1

Physical and chemical laboratory data for selected Pennsylvania Ridge and Valley soils. (Data from Carter, 1983; Hoover, 1983; Cronce, 1988; Ciolkosz and Dobos, 1990.)

Horizon	Depth (cm)	Munsell color	Rock fragments > 2 mm (wt%)	Sand 2.0–0.05 mm (%)	Silt 0.05–0.002 mm (%)	Clay < 0.002 mm (%)	Organic carbon (%)	Free iron oxides Fe_2O_3 (%)	Base saturation[a] (%)	pH[b]
Ridge top Spodosol (Leetonia; 14–54; sandstone material)										
A	0–5	7.5YR 3/1	11.1	76.0	22.0	2.0	4.47	0.2	9	3.5
E	5–13	10YR 5/2	15.2	74.1	24.2	1.7	1.02	0.1	22	3.6
Bhs	13–15	7.5YR 4/4	19.9	61.4	29.8	8.8	2.51	0.9	4	3.7
Bs	15–20	7.5YR 5/8	21.3	61.4	29.1	9.5	2.52	1.0	4	4.4
Bw1	20–28	10YR 5/6	68.2	62.0	28.5	9.5	0.87	0.6	5	4.4
Bw2	28–58	7.5YR 5/6	37.8	63.3	26.6	10.1	0.25	0.7	3	4.3
Bw3	58–117	7.5YR 5/4	54.5	57.6	34.1	8.3	0.08	0.7	13	4.4
BC	117–150	7.4YR 7/2	36.0	80.7	17.1	2.2	0.02	0.2	59	4.5
C	150–290	7.5YR 7/4	55.7	80.9	17.0	2.1	0.01	0.1	50	4.5
Ridge top Inceptisol (Hazleton; 31–33; sandstone material)										
A	0–5	10YR 3/1	41.6	68.6	25.8	5.6	4.46		10	4.6
BA	5–36	7.5YR 5/6	54.9	65.8	25.0	9.2	1.44		10	5.0
Bw1	36–74	5YR 5/4	75.8	64.6	23.2	12.2	0.24		9	5.0
Bw2	74–142	7.5YR 5/4	91.6	52.8	35.0	12.2	0.08		7	4.9
C	142–163	7.5YR 5/4	82.1	55.7	32.7	11.6	0.24		9	5.0
Mid-footslope Ultisol (Laidig; 14–63; sandstone–shale material; colluvial buried paleosol)										
A	0–3	10YR 5/3	> 50.0	33.2	48.6	18.2		1.4	5	4.1
E	3–36	10YR 7/4	> 50.0	42.6	43.4	14.0		1.3	3	4.4
BE	36–46	10YR 6/4	62.4	44.9	38.4	16.7		1.6	4	4.5
Bt1	46–61	7.5YR 5/6	53.3	42.7	34.9	22.4		1.7	5	4.6
Bt2	61–79	7.5YR 5/6	42.3	40.9	36.5	22.6		2.4	8	4.8
Bxt1	79–107	10YR 5/6	39.4	49.4	33.0	17.6		2.3	8	4.6
Bxt2	107–150	10YR 4/6	48.8	48.9	31.2	19.9		2.4	10	4.8
Bxt3	150–175	10YR 4/6	45.7	46.5	30.7	22.8		2.1	12	4.7
Bxt4	175–203	10YR 4/6	53.1	44.9	33.0	22.1		2.2	13	4.8
Bxt5	203–218	7.5YR 4/6	50.7	58.7	27.8	13.5		1.7	18	4.7
2Bxtb	218–241	5YR 4/6	42.9	52.1	27.3	20.6		2.7	15	4.7
2Btb1	241–292	5YR 5/8	62.8	58.7	23.4	17.9		2.4	12	4.7
2Btb2	292–368	5YR 5/8	42.5	57.1	22.9	20.0		2.5	12	4.8
Lower footslope Ultisol (Murrill, 14–30; limestone material; limestone residuum buried paleosol)										
Ap	0–28	10YR 3/3	22.1	33.0	53.4	13.6	1.17	1.7	27	5.6
Bt1	28–56	7.5YR 5/4	51.4	43.5	33.9	22.6	0.21	2.9	36	5.6
Bt2	56–86	7.5YR 5/6	40.3	34.3	36.8	28.9	0.11	2.9	16	5.0
Bt3	86–104	7.5YR 5/6	29.4	28.4	36.2	35.4	0.11		11	5.0
2Btb	104–132	5YR 5/8	22.3	19.5	36.9	43.6	0.08	3.8	18	4.9
2BCb	132–170	5YR 5/8	2.7	17.9	31.8	50.3	0.09		11	4.9
2Cb	170–190	10YR 6/8	4.5	5.9	56.4	37.7	0.06	5.3	18	4.8
Valley floor Alfisol (Hagerstown; 14–75; limestone residuum)										
A	0–10	10YR 3/2	17.4	10.3	73.2	16.5	5.32	1.2	65	6.6
E	10–25	10YR 4/6	19.4	10.6	73.8	15.6	0.78	1.2	38	5.7
BE	25–41	7.5YR 6/6	15.8	9.2	69.3	21.5	0.28	2.0	19	4.9
Bt1	41–61	5YR 5/6	12.6	5.6	45.4	49.0	0.19	3.9	32	4.8
Bt2	61–86	5YR 5/6	5.9	3.8	38.4	57.8	0.18	4.1	57	5.4
Bt3	86–117	5YR 5/6	5.7	6.5	43.5	50.0	0.20	3.6	61	5.8
Bt4	117–150	5YR 5/6	2.5	1.9	43.5	54.6	0.26	4.3	76	6.9
Bt5	150–160	7.5YR 5/6	3.1	4.3	42.6	53.1	0.30	3.8	87	7.5
R	160–170	10YR 4/1	hard, thick bedded dolomite							

[a]Sum of cations method.
[b]1 : 1 soil : water.

was mapped extensively. In the last 10 to 15 years, with the use of power equipment, it has been found that these soils are much deeper to bedrock. The reason they were thought to be only moderately deep (50 to 100 cm) in the past was that when they were dug with hand tools, many large rock fragments were encountered and it was assumed that the large rock fragments were bedrock. Many recent observations indicate that bedrock is greater than a meter deep, but the maximum depth is not well documented. To the authors' knowledge, the work of Carter and Ciolkosz (1986) is the only systematic study of depth to bedrock on a ridge top in the central Ridge and Valley area. Their study indicates that on the ridge top upland the bedrock is 3 to 4 m deep while on the shoulder slopes and upper back slope areas the depth to bedrock is 5 to 7 m. Their data also indicate that on the ridge top upland and shoulder slopes (slopes < 15 to 20%) the upper 1 to 2 m of the material is not residual. They also indicate that on the steeper upper backslope areas the upper 2 to 4 m of material is also not residual. They concluded that the upper material has moved down slope and is colluvium while the lower material between the colluvium and bedrock is residuum. This conclusion is based on seismic and rock fragment orientation data and the presence of sorted and stratified material in the colluvium. The upper material in some broad level areas may not have moved a long distance and may be primarily cryoturbated. These upper and lower materials are very similar in most respects (color and texture), and distinguishing one from the other without detailed studies is difficult. This is the main reason why it has been assumed that the soils developed on these landscape areas were residual soils. Carter and Ciolkosz (1986) also concluded that the colluvium had moved down slope via periglacial solifluction processes during Wisconsinan time. At this time glacial ice was just 70 km north (Crowl and Sevon, 1980) of their research site. Although Denny (1956) states that periglacial activity decreased rapidly away from the Wisconsinan glacial border, the work of Clark and Ciolkosz (1988) and Ciolkosz et al. (1986b) indicates extensive periglacial activity throughout Pennsylvania and southward in the Appalachians. Thus, the model developed by Carter and Ciolkosz (1986) appears to be applicable to all the ridge tops in the northern Ridge and Valley area. A slight modification of the model may be necessary on some of the ridges which have a more resistant sandstone such as the Tuscarora Formation. But even in these cases the modification probably would be only a shallower depth to bedrock, and therefore slightly less residual material.

According to Carter and Ciolkosz (1986), these landscape areas are stable today and have been stable since the retreat of the Wisconsinan glacial ice from Pennsylvania. Although logical, these authors do not provide support for this conclusion, nor do they provide any information as to the timing of the colluviation. Radiocarbon dates of 21 to 28 ka for colluvium in northern West Virginia (Jordan et al., 1987) and thermoluminescence dates of 12 to 18 ka from basal loess on colluvium in the unglaciated area of southwestern New York (Snyder, 1988) provide support for late Wisconsinan age colluviation. An additional date of 12 ka from basal bog material on top of colluvium in central Pennsylvania at Panther Run (Watts, 1979) also indicates a termination of the colluviation at about 13 to 15 ka. The age of the initiation of the colluviation is uncertain although the West Virginia data all come from the lower part of the colluvial deposit and may indicate that most of the colluviation is of late Wisconsinan age (14 to 28 ka). The amount of early Wisconsinan (28–75 ka) colluviation is also uncertain, although oxygen isotope data (Braun, 1989; CLIMAP, 1984) indicate climatic conditions less conducive to colluviation in early Wisconsinan than late Wisconsinan time. In addition, the work of Eyles and Westgate (1987) in Canada, and of Ridge et al. (1990) in Pennsylvania and New Jersey which indicates no glacial ice in the northeastern United States in early Wisconsinan time also suggests that little colluviation occurred prior to the late Wisconsinan. These various studies indicate that the last colluviation of the ridge tops in central Pennsylvania most likely took place in late Wisconsinan time.

The production of the residuum by disaggregation (weathering) of the bedrock on the ridge tops is a continuous process. Data on the weathering rates of sandstones as well as other consolidated rocks are sparse (Alexander, 1988). Leneuf and Aubert (1960) suggest that coherent rocks weather at the rate of 1.3 to 3.3 cm ka^{-1}. Alexander (1985, 1988) presents data on the quantity of soil material formed from consolidated rocks. These data give a general value of 3.3 cm ka^{-1} of soil formed when converted using a soil bulk density value of 1.5 g cm^{-1}. Alexander's (1985) data also provide a value of about 1 cm ka^{-1} of soil formed for sandstones. If this value is applicable to the ridge top sandstones, then about 15 cm of rock has weathered to soil on the ridge tops since the late Wisconsinan (last 15 ka). The only known weathering rate study in Pennsylvania is that of Sevon (1984), who estimated that a conglomeratic sandstone in Pike county has weathered at the rate of 0.026 cm ka^{-1}. Using this value only 0.4 cm of sandstone soil residuum would have been produced on the ridge tops since the Wisconsinan. Although the difference between these two estimates is about 40 times, they both point out that the 2 to 3 m of residuum on the ridge tops could not have formed in the last 15 ka. During the Wisconsinan, these areas were a tundra with permafrost (Watts, 1979; Ciolkosz et al., 1986b; Delcourt and Delcourt, 1983, 1986). Under these conditions the amount of rock disintegration that took place by frost processes is unknown. However, with the exception of near the surface it probably was not extensive and produced primarily rock fragments and only small amounts of fine earth (<2 mm) material. Thus the majority of the rock weathering that has taken place on the ridge tops must have occurred in the pre-Wisconsinan. Although weathered in the pre-Wisconsinan, the solum (A and B horizon) of the pre-Wisconsinan soil has been removed from most areas and only the lower C horizon remains as the residuum. Although upland interfluves are usually least affected by truncation, the sandstone ridge tops were significantly affected because they are narrow and consist of two shoulder slopes that meet to form the interfluve. The study of Carter and Ciolkosz (1986) did not extend to the backslope area, and according to Ciolkosz et al. (1989) steeper slopes are truncated more than gentle slopes by periglacial colluviation. Thus the back slope areas which are the steepest part of the slope profile should have been truncated more than the interfluves or shoulder slopes. This conclusion has yet to be substantiated by a detailed study, although some observations of depth to bedrock and the occurrence of unweathered shale masses from the lower back slope area in the footslope colluvium indicate that the backslope was severely eroded. Thus, it appears that the ridge top interfluve, shoulder slope and backslope areas were all truncated during the Wisconsinan, although the back-slope probably was truncated more than the other areas.

Genetic pathway

The genetic pathway of the ridge top soils has been complex. Their A, B, and in most cases, upper C horizon, have developed in sandstone colluvium and appear to have an age of about 15 to 20 ka. Their lower C horizon is developed in residuum and is much older than the upper horizons. It represents the lower part of a truncated pre-Wisconsinan soil. On the ridge tops, skeletal Haplorthods (Leetonia) and skeletal Dystrochrepts (Hazleton) occur. The Haplorthods, although very interesting soils, account for only 3 to 4% of the area (Lipscomb and Farley, 1981; Merkel, 1978).

The Spodosol forming process is maximized in areas with acid, sandy parent material, frigid soil temperatures, and coniferous vegetation (Stanley and Ciolkosz, 1981; Mokma and Burrman, 1982; DeConinck, 1980). The ridge tops are close to the frigid–mesic soil temperature boundary (Carter and Ciolkosz, 1980), with some being slightly below and some slightly above the boundary. In this temperature boundary zone, the intermittent occurrence of the Spodosols suggests that the vegetation varied with some areas having coniferous vegetation, probably hemlocks, which have a strong influence on Spodosol formation (Johnson and Siccama, 1979) while other areas had hardwood vegetation. Although these soils show typical Spodosol morphology many, according to the Soil Taxonomy (Soil Survey Staff, 1975) classification criteria, do not qualify as Spodosols. The reason is that during their formation clay as well as Fe, Al, and humus has accumulated in the B horizon (Table 1), and the accumulation of clay makes the Fe+Al to clay ratio too low to allow many of the soils to qualify as Spodosols (Stanley and Ciolkosz, 1981). The Spodosol classification criteria are presently under review and most likely will be amended to eliminate this problem. The solum thickness of the ridge top Spodosols is about a meter, and according to Franzmeier and Whiteside (1963) this amount of Spodosol formation can take place in about 10,000 years. These data tend to confirm that the last soil-forming cycle started on these ridge tops about 10–15 ka ago.

The major soil on the ridge tops is a skeletal Dystrochrept (Hazleton). In addition to being the dominant soil on the ridge tops, Hazleton is the most extensive soil in Pennsylvania. It occupies 10.4% of the state (Ciolkosz et al., 1986a, 1990) and is found both on the ridge tops and on large areas of the unglaciated Appalachian Plateau. This soil owes its sandy texture and cambic subsurface horizon to its sandstone parent material which contains little clay that can be eluviated to form an argillic horizon. There are also few minerals present which can be weathered to form clay. Although these parent materials resist rapid changes the soils developed in them would show much greater development including an argillic horizon and red oxidized colors had they not been truncated in the Wisconsinan. This conclusion is based on the work of Waltman (1985) who found residual, red paleosols with argillic horizons developed from sandstone residuum buried by Wisconsinan colluvium on the unglaciated Appalachian Plateau in north-central Pennsylvania.

The work of Carter (1983) also indicates that the ridge top soils have lamellae which have about 4% more silt and 5% more clay and a redder color (7.5YR 5/6 vs 10YR 7/3) than the matrix fine earth. The lamellae are 3 to 20 mm thick, occur in the BC and C horizons as a layer parallel to the soil surface every 15 to 30 cm, and decrease in development with depth. When the lamellae intersect rock fragments, they encircle and stain the rock fragments indicating they post-date the colluviation and are not inherited from the bedrock. Silt caps were also found on top of many of the rock fragments in the lower part of these soils. The lamellae and silt caps occur in the colluvial as well as the underlying residuum. Silt caps have been noted in other soils (Fitzpatrick, 1987; Lyford and Troedsson, 1973; Ugolini, 1986) and by the senior author in glacial till soils in New England. The silt caps are 1 to 25 mm in thickness, yellowish in color (10YR 5/3 vs 7.5YR 7/4-matrix), and their particle size distribution show a greater amount of clay and silt and less sand than the matrix (Table 2). Of the sand, silt, and clay, the silt content shows the greatest difference between the matrix and the silt cap. The accumulation of the silt cap material on the rock fragment surfaces has been attributed to sorting in the soil. This type of sorting has been explained by both frost and water processes, although frost sorting seems to have the most support (Ugolini, 1986). Carter observed that the lamellae cut across the silt caps. Thus, the lamellae are younger. This

TABLE 2

Particle size distribution data[a] for silt caps and adjacent matrix material from four different sandstone soil profiles on the same ridge top. (Data from Carter, 1983.)

Soil No.	Depth (cm)	Sand (mm)						Silt (mm)			Sand	Silt	Clay
		2.0–1.0	1.0–0.5	0.5–0.25	0.25–0.1	0.1–0.07	0.07–0.05	0.05–0.02	0.02–0.005	0.005–0.002	2.0–0.05	0.05–0.002	<0.002
Silt cap													
14-55-08	165–206	0.2	0.4	3.6	22.7	6.7	4.8	5.0	43.5	9.2	38.5	57.7	3.8
14-56-09	183–287	0.9	1.5	8.8	34.5	7.3	3.5	6.6	21.4	10.2	56.6	38.2	5.2
14-58-11	274–284	0.8	0.5	3.5	22.6	8.8	7.7	18.3	25.8	7.7	43.8	51.8	4.4
14-60-10	170–211	1.1	0.8	2.6	17.2	7.1	6.1	16.2	30.8	12.1	34.9	59.1	6.0
Matrix													
14-55-08	165–206	1.2	2.5	10.3	51.3	10.5	5.5	5.8	7.0	4.1	81.3	16.9	1.8
14-56-09	183–287	0.8	2.0	14.3	39.9	8.8	5.7	9.6	12.4	3.8	71.5	25.8	2.7
14-58-11	274–284	1.6	2.6	10.5	52.7	13.6	7.6	5.0	5.6	0.4	88.6	11.0	0.4
14-60-10	170–211	1.3	2.9	16.8	53.0	7.6	4.5	4.6	5.4	1.7	86.1	11.7	2.2

[a]Values (except depth) in percent.

relationship indicates that the silt caps are probably frost sorted fossil features that are the result of deep freezing after the permafrost melted out of the soils. The sorting would also have to have taken place in a relatively short period of time because under present natural forest conditions these soils do not freeze to a depth of more than 25 cm (Carter and Ciolkosz, 1980). These data would indicate that the silt caps are late Wisconsinan in age.

The A and B horizons of the ridge top soils also have been enriched in clay and silt (Table 1). Table 4 gives an estimate of the amount of silt and clay added to the soils studied by Carter (1983). This material was not produced in the soil by weathering and is probably an eolian addition. Eolian additions are common in soils (Smith et al., 1970; Syers et al., 1969; Cronce, 1988). The ridge top soils are natural dust traps because the many rock fragments and coarse texture of the surface allows rapid infiltration into the soil. Using the present day accumulation rate of 235 kg ha^{-1} yr^{-1} given by Smith et al. (1970), it would take less than 6000 years to account for the added material in the A and B horizons of the ridge top soils. Thus, eolian additions are a very likely source for the bulk of the silt and clay in the ridge top soils. Eolian additions may also account for some of the silt and clay in the buried sandstone paleosols studied by Waltman (1985).

TABLE 3

Estimated amount of clay and silt added to ridge top soils. (Data from Carter, 1983.)

Site[a]	kg/ha[b] ($\times 10^6$)	Clay[c] (%)	Silt[c] (%)
TS	1.41	37	63
NW-8	0.45	35	65
NW-31	0.36	26	74
NW-57	0.43	26	74
SW-10	0.88	33	67
SW-21	0.98	25	75
SW-23	0.74	28	72
SW-52	0.41	8	92
SW-55	0.63	0	100

[a]TS = top site, NW-8 = northwest aspect 8% slope, SW-10 = southwest aspect 10% slope.
[b]Clay and silt.
[c]Relative amount of added material.

Carter's (1983) work also indicates that ridge top soils with slopes of less than about 15% have very similar properties. When the slope exceeds about 15%, the soils show a decrease in solum (A+B horizon) thickness with increasing slope gradient. In addition, the soils on steeper slopes show a thinner solum on southern aspects than on comparable northern aspects. These trends have been attributed by Carter (1983) to variations in runoff on gentle versus steep slopes and sun exposure with varying aspect and its effect on the moisture relations of the soils.

The clay mineralogy (<2 μm) of the ridge top soils is dominantly kaolinite. The kaolinite does not indicate an advanced stage of soil weathering. The kaolinite shows a high degree of crystallinity as indicated by X-ray analysis and its content is constant with depth or increases from the surface horizons to the B and C horizons. These data indicate that the kaolinite is not being synthesized in the soil, but is largely inherited from the bedrock. The kaolinite in the bedrock is probably at least in part an in situ weathering product. The rate of kaolinite synthesis with depth in the bedrock is not known although Singh et al. (1982) indicate a 6 m weathering zone in a sandstone in northern West Virginia.

Footslope soils

The major soils on the footslopes of the ridges are deep, acid, well drained to poorly drained, medium to fine textured, and most have fragipans (Table 1; Fig. 2). They are developed in colluvium (Ciolkosz et al., 1979) and are classified as Fragiudults (Laidig, Buchanan), Fragiaquults (Andover), Fragiudalfs (Clarksburg), and Hapludults (Murrill).

Parent material and time

The footslope colluvium is much different than the colluvium on the ridge top interfluves and shoulder slopes. It was derived from sandstone from the ridge tops, sandstone and shale from the backslopes, and shale and limestone material from the footslope area. In general there is a trend of increasing clay and silt content from the top to the bottom of the footslope. This is a reflection of the colluvium having picked up finer materials as it moved down over the shale and limestone areas. Typically the colluvium is a heterogeneous mixture of rock fragments (primarily sandstone) with a fine earth (<2 mm) matrix that varies in texture from sandy loam (upper footslope) to silty clay loam (mid footslope) to silty clay (lower footslope). In many respects, the colluvium resembles glacial till found in northeastern Pennsylvania, although it tends to be more heterogeneous with respect to texture, rock fragment content, and balls or masses of material, particularly shaly or paleosol material. In an up and down slope cross-section, the colluvium forms a wedge with the maximum thickness in the center and thinner up and down slope limbs (Fig. 2). In areas adjacent to gaps in the ridges large fans occur. Many of these gaps presently have streams that head in the ridge top areas, pass through the gap, and over the fan. Limited observations of these fans indicate they are very complex and are composed of both alluvium and colluvium.

Recent work of Hoover and Ciolkosz (1988) indicates that the stratigraphy of the colluvium is more complex than previously perceived. Their work indicates that older colluvium is buried beneath the colluvium at the surface. The buried colluvium is characterized by bright red (rubified) colors (5YR 5/6 to 2.5YR 5/8) while the upper colluvium is primarily yellowish brown (10YR 5/4 to 10YR 5/6). The relative thickness of the red and brown colluviums is not known although from the work of Hoover and Ciolkosz (1988) and other limited observations, it appears that the red colluvium is thicker than the brown colluvium. This appears to be the case for both simple side slope and fan deposits. An example of this relationship was observed in a very large fan on the Susquehanna River at McElhattan, Pennsylvania (6.5 km up river from Jersey Shore). In this fan there is about a meter of brown colluvium overlying > 3 m of red colluvium. The lateral extent (parallel to the ridges) of the red colluvium is not known although the work of Hoover and Ciolkosz (1988) and other observations indicate that it is very extensive. It is also not known whether the red colluvium is continuous or discontinuous beneath the brown colluvium.

The deposition of the brown colluvium is believed to be contemporaneous with the deposition of the colluvium on the sandstone ridge tops. Thus, it is a part of a colluvial continuum. The age of the red colluvium is unknown. Although not dated, the red colluvium's position below the Wisconsinan colluvium and an examination of the isotope record (Braun, 1989; CLIMAP, 1984) indicate that it is probably associated with the stage 6 (128 to 195 ka) glaciation, and therefore would be correlated as pre-Wisconsinan 1. The bright red color, plus clay films and soil structure indicate that the red colluvium represents a buried paleosol (Hoover, 1983). In addition to these buried paleosols, other paleosols have been identified in Pennsylvania (Ciolkosz et al., 1988). It is also not known if there are buried colluviums older than pre-Wisconsinan 1. In the future, deep excavations in the footslope areas may reveal additional pre-Wisconsinan buried colluviums. Although it appears that the bulk of the colluvium is pre-Wisconsinan, very little pre-Wisconsinan material is found at the surface. It is almost always buried by the brown Wisconsinan material. Although it is logical that there should be a latitudinal trend in the distribution and thickness of the colluvium because of moderating climatic conditions southward, no studies have been done to identify and quantify these trends. The only distribution data known to the authors are those given by Clark et al. (1989), which indicate that counties in the Pennsylvania Ridge and Valley have about twice the coverage (25% vs 12% of the total county) of colluvial soils as counties in the Ridge and Valley of southern Virginia.

Genetic pathway

The texture of the footslope soils is variable and primarily a reflection of the parent material. Superimposed on these inherited textures is a weak to moderately well developed argillic horizon. The weakly developed argillic horizon is usually associated with the sandy to loamy soils while the better expressed argillic horizons are associated with the finer textured materials. The colluvial soils are acid throughout and are classified as Ultisols except for some poorly drained soils, particularly those developed from limestone materials, which are Alfisols (Ciolkosz et al., 1979). The Ultisols are parent material Ultisols which means that during their development they never had a base status high enough to be an Alfisol (Ciolkosz et al., 1989). In the soils developed from limestone materials, the base saturation frequently increases in the lower part of the profile (Ciolkosz et al., 1979). In the poorly drained soils, particularly those developed from limestone colluvium, the higher subsoil base saturation appears to be a result of recharging with bases from the groundwater (Ranney et al., 1974). The moisture regime of the poorly drained soils is associated with landscape position (discharge area) or impermeable zones in the soils (fragipans). The soils developed in sandstone–shale colluvium are more abundant and tend to be more poorly drained than those developed from limestone colluvium (Table 4).

The argillic B horizon formed in the brown Wisconsinan colluvium appears anomalous when compared to the cambic B horizon developed in Wisconsinan glacial till soils derived from similar rock lithologies. No radiocarbon dates are available in Pennsylvania to date the colluvium, and the only dates available for similar colluvial material are between 21 and 28 ka (Jordan et al., 1987). Thus, it appears that it takes less time to develop an argillic horizon in the colluvium than in lithologically similar glacial till. A possible explanation may be that the colluvium was initially

TABLE 4

Relative amount (percent) of colluvial soils in various drainage classes in the central Pennsylvania Ridge and Valley. (Data from Braker, 1981; Merkel, 1978; Lipscomb and Farley, 1981.)

County	Soil drainage classes		
	Well	Moderately well and somewhat poorly	Poorly and very poorly
Sandstone–shale colluvium			
Centre	36	27	24
Huntingdon	36	40	15
Juniata and Mifflin	51	26	15
Limestone colluvium			
Centre	11	2	–
Huntingdon	6	3	–
Juniata and Mifflin	7	1	–

more weathered than the till, and that the clay was more easily dispersed in the upper horizons and available for eluviation into the B horizon.

Fragipans occur in the soils developed in the brown colluvium with the exception of the well drained soils developed in limestone colluvium. Fragipans develop in transported parent material of moderate clay content and their genesis is associated with both close packing of their mineral grains and some grain to grain cementation (Ciolkosz et al., 1989; Lindbo and Veneman, 1989). The fragipans found in the brown colluvial soils start at 50 to 100 cm from the surface and continue downward and sometimes penetrate into the red paleosol material, but usually only a short distance. It appears that where the fragipan penetrates into the red material, the red material has been moved down slope and redeposited and is not an in situ paleosol. The lack of penetration of the fragipan into the in situ paleosol may be related to the fact that the paleosol is a more highly weathered material. The in situ paleosol probably had a fragipan in it during its early stage of development, but it was destroyed by pedogenesis (Ciolkosz et al., 1985, 1989), and now the soil material is not amenable to fragipan formation. It is not known why a fragipan has not formed in the in situ paleosol, but it may be related to a reorganization of the packing arrangement of the soil particles and the leaching of fragipan cements during pedogenesis.

In lower footslope areas the red paleosol buried below the brown colluvium has formed both in older colluvium and in limestone residuum (Table 1). Regardless of the parent material, the paleosol appears highly oxidized, although its iron oxide content is not excessively high. In addition, its clay mineralogy does not indicate extreme weathering (Hoover, 1983; Ciolkosz and Dobos, 1990). These data agree with the data of Levine and Ciolkosz (1983) which indicate that bright red pre-Wisconsinan till soils do not show extensive alteration of silicate clays to kaolinite and gibbsite when compared to late Wisconsinan till soils. The high chroma red color which characterizes the paleosol is a reflection of its iron oxide mineralogy. The iron oxide mineralogy of the Ridge and Valley soils has not been studied, but other soil studies indicate that red colors are due to a relatively high proportion of the soil's free iron oxides being hematite (Schwertmann and Taylor, 1977). This leads to the question, why are the soils in the upper colluvium brown and in the lower colluvium red. The common explanation is that in the pre-Wisconsinan interglacial, the climate was warmer and drier than today; therefore, hematite synthesis was favored over goethite as iron was released from mineral lattices during weathering (Waltman, 1985). Schwertmann (1988) adds high pH and a rapid release of iron to the weathering environment as additional factors favoring hematite (red) synthesis over goethite (yellowish brown). It would appear that with the possible exception of residual limestone soils, the present day climate interacting with the parent materials in the

Ridge and Valley area favor a goethite pathway of iron oxide synthesis resulting in yellowish brown soil development in Wisconsinan age parent materials. The oxygen isotope record (Braun, 1989; CLIMAP, 1984) indicates since stage 6 (pre-Wisconsinan colluviation 1) that only stage 5e compares climatically with today. In addition, 5e was only about equal in length to the Holocene. If the climate during 5e was similar to today then it is difficult to explain the genesis of the buried colluvial paleosol. Therefore, the climate must have been different, but how different? The data of King and Saunders (1986) and King (pers. commun., 1986) indicate that the climate in central Illinois during the 128 to 118 ka (5e) period was much warmer with no annual frost (possibly equivalent to south Florida or southern Texas today) and that there was less precipitation (possibly equivalent to western Iowa today). They also believe that the climate during the period 118 to 75 ka was similar to that of today. If Saunders and King are correct, then the mean annual soil temperature (MAST) in central Pennsylvania during 5e was about 22°C in contrast to the 8 to 10°C of today (Smith, 1984; Carter and Ciolkosz, 1980), and significantly exceeded the 15°C that has been used as an indicator of a soil rubification (reddening) threshold (Bullock, 1985). If the climatic interpretations of King and Saunders approximate true conditions, then the red colluvial soils were exposed to temperatures equal to or greater than today's temperatures for about 50,000 years (75 to 128 ka) with only a minor part of this time (10,000 years) having conditions which would cause rubification of the soil.

The soil material above the fragipan in these and many other fragipan soils tends to have more silt and clay than the fragipan horizons. This has led some to the conclusion that fragipans form at lithologic discontinuities in the soil parent material (Hoover, 1983; Smeck et al., 1989). This may be the case in some situations but does not seem to apply to the fragipans of the Ridge and Valley. A better explanation is that the increase in silt and clay is related to weathering above the fragipan and the addition of silt and clay from eolian sources.

Valley soils

Adjacent to the footslopes in limestone valleys the major soils are deep, well drained, acid and clayey (Table 1; Fig. 2). They are classified as Hapludalfs (Hagerstown) and are developed in residuum. In the shale valleys the major soils are moderately deep, well drained, acid and have silty textures. The shale soils are classified as Dystrochrepts (Berks, Weikert) and are developed in residuum. Detailed studies of the shale soils are not available; thus, the following discussion will concentrate on the limestone soils.

Parent material and time

The limestone valley soils have developed in residuum that accumulated as the $CaCO_3$ of the limestone dissolved and was leached away. The amount of insoluble residue in the limestone is usually 3 to 10%. The amount of the residue and the solubilization rate of the limestone have been used to estimate the age of the limestone residuum. Using this approach, Cronce (1988) estimates the age of the 1.5 to 2 m of residuum at sites in central Pennsylvania to be 50 to 350 ka. Using a similar technique, Ciolkosz et al. (1986a) estimate that limestone residuum accumulates at the rate of 30 cm $(100\ ka)^{-1}$ and 300 cm Ma^{-1}. Accumulations of 2 to 3 m of residuum are common on level landscapes, although because of the very uneven weathering of limestones the depth to bedrock can vary by a few meters in a lateral distance of 3 to 4 m. These data indicate that the residuum on the level, more stable parts of the landscape approaches mid-Pleistocene in age. Thinner residuum is noted on sloping areas particularly near drainage ways. The difference in thickness of the residuum with landscape position is not well documented. The main reason for the lack of documentation is the great difficulty in determining the depth to bedrock laterally on the landscape due to the uneven pinnacle weathering pattern of the limestone.

Because of the slow rate of residuum accumulation, it is difficult to give these soils a precise age because material is constantly being added from below. For example, in a 3 m thick soil using the accumulation rate of Ciolkosz et al. (1986a), the bottom 30 cm would be 100 ka and the top of the soil would be 1 Ma.

Since the thickness of residuum varies on these landscapes, processes have been active to give different ages to different parts of the landscape. The study of Cronce (1988) on gentle slopes (1 to 5%) indicates that the properties of the upper 50 cm of these soils is the same over residuum and adjacent late Wisconsinan age ice wedge casts. This indicates a plug or rug-like mass movement down slope of the upper 50 cm of material. Cronce (1988) concluded that the movement occurred while permafrost was in the soil and the 50 cm depth marked the seasonal thaw or active zone. For a 2 m thick soil, this would indicate an in situ late Wisconsinan age for the upper 50 cm and at 50 cm a 500 ka age soil. Cronce also indicates that the soil forming processes of the last 15 ka have impressed themselves through the upper material into the truncated paleosol welding the two together in the manner described by Ruhe and Olson (1980). For these soils, this process probably has been repeated up to four times in the last 0.6 Ma, associated with glacial stages.

Genetic pathway

The texture of the limestone valley soils is silt loam at the surface and clay to silty clay loam in the subsoil (Table 1). The strong contrast in texture between the surface and subsurface horizons has in the past been attributed to eluviation of clay from the surface horizons and its deposition in the B horizon forming an argillic horizon. Cronce (1988) indicates that the soil material in the lower part of these profiles is directly inherited from the insoluble material that remains after the $CaCO_3$ of the limestone was dissolved and leached away during soil formation. Cronce also estimates that 398,000 to 716,400 kg ha^{-1} of eolian material has been mixed into the upper 60 to 100 cm of these soils. The larger quantity when mixed with 5 cm of clayey material would give about 10 cm of silt loam material. The B horizons of these soils have extensive clay films indicating clay illuviation from the upper horizons. Thus, the silty surface horizons appear to be a combination of some clay eluviation as well as the addition of eolian material.

The argillic horizon in these soils appears to have too much clay and free iron oxides than can be accounted for by weathering of the residuum or recent eolian additions (Cronce, 1988). Cronce proposes that the excess clay was eluviated into the B horizons from soil material that subsequently has been eroded from the soil surface. This process has also been proposed by Ballaugh and Runge (1970) to explain the genesis of limestone soils in Illinois. Cronce also proposes that a base saturation of 40% is a lower threshold for argillic horizon formation in these soils. This proposal is based on the association of 40% base saturation with the start of the argillic horizon in the soils he studied.

The color of most of the limestone soils is red; thus, the dominant iron oxide mineral is hematite. In these soils the red color extends throughout the B horizon to the bedrock with only a thin (2 to 6 cm) weathering rind (usually gray in color) on the limestone. In various places yellow limestone soils occur in juxtaposition to the red soils, and in some places red and yellow layers associated with certain beds in the bedrock occur in the same soil. Therefore, the red color cannot be just a reflection of a warmer and possibly drier climate. The yellow soils also tend to have less clay than the red soils. It is not known if the texture of the residuum is significant in determining the pathway of iron oxide mineral genesis. Assuming it is not important, it appears that the concentration of the iron in the weathering environment may be the major factor in directing the pathway of hematite synthesis. Support for this contention is the higher free iron oxide contents (about twice as much) found in the red than in the yellow soils (Ciolkosz et al., 1986a).

Although the thick limestone residuums date back many thousands of years, their clay mineralogy like that of the ridge top soils and the buried red colluvial and residual paleosols does not show advanced weathering. The clay minerals of the limestone soils below a 50 cm depth are dominantly illite with minor amounts of kaolinite and vermiculite.

Conclusions

The soils and landscapes of the Ridge and Valley of central Pennsylvania show a distinctive imprint of the climatically controlled processes that have acted with varying intensity over time in this area. The soils on the ridge tops and shoulder slopes have developed in sandstone colluvium or cryoturbated material and are of late Wisconsinan age. They are sandy, have a high rock fragment content, show weak development and are Dystrochrepts and Haplorthods. The lower parts of these soils are developed in residuum that is much older than the upper material and represent the lower part of a pre-Wisconsinan soil that was truncated during late Wisconsinan time by periglacial processes.

Material that was eroded from the ridge tops moved down the backslope and accumulated on the footslope as colluvium. The backslope area is the steepest part of the landscape and appears to have been truncated to a greater degree than the ridge tops. The truncation of these areas probably occurred as solifluction flow. The colluvium from the ridge tops and sideslopes moved down across a red (rubified) pre-Wisconsinan soil developed in colluvium and limestone residuum on the footslope, and covered the pre-Wisconsinan soil with a varying thickness of brown colluvium. In some areas the pre-Wisconsinan soil was significantly truncated and in other areas it was not altered greatly. The soils developed in the brown colluvium are medium to fine textured, have argillic horizons, most have fragipans, and are Fragiudults, Hapludults, Fragiaquults, and Fragiudalfs. When the brown colluvium is relatively thin (<1 to 2 m) the present-day soil-forming processes have welded the upper soil to the underlying paleosol. The bright red color of the paleosol appears to have developed in the soil under conditions that were warmer and possibly drier than today and may be associated with isotope stage 5e.

Adjacent to the footslope area in limestone valleys, soils occur that have developed from residuum that accumulated when the calcium carbonate was leached from the limestone. These soils have silty surface and usually red clayey subsurface horizons and are classified as Hapludalfs. If the soils are thick, they can be very old (mid-Pleistocene). Regardless of the age, the upper 50 cm has been affected by downslope movement of material which has given these soils a very uniform upper sequence of horizons. The downslope movement appears to have been coincident with the downslope movement on the ridge tops and footslope areas. The silty texture of the surface horizons of these soils is the result of eolian additions to the soil and the eluviation of clay from the surface to the subsurface horizons. Eolian material also has been added to the footslope colluvial and ridge top soils. In the ridge top soils, because of their coarse texture, very little of the eolian material has accumulated on the surface. It has moved downward into the lower horizons of the soil.

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Soils of Nittany Valley*

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Introduction

Nittany Valley is the first western valley of the Ridge and Valley physiographic province of Pennsylvania. A discussion of the geology of the valley is presented by Butts and More (1936), and its geomorphic evolution is discussed by Gardner (1980), Parizek and White (1985), and Ciolkosz and Dobos (1990). The soils of the valley are diverse (Fig. 1) as are the soils of the state (Fig. 2 & Table 1). Additional soils information for Pennsylvania is given by Ciolkosz et al. (1988, 1990a, 1990b) and for the Northeast by Cunningham and Ciolkosz (1984), Ciolkosz and Dobos (1989), and Ciolkosz et al. (1989).

The diversity of soils in Nittany Valley as well as in Pennsylvania is a reflection of the influence of the soil forming factors (parent material, organisms, climate, topography, and time) on the development of the soils. The effect of any one of these factors on soil development can vary from very little to very great. Of particular significance is the strong association these soils have to landform and parent material (Fig. 3). Because of these relationships, the soils can be arranged in parent material-drainage sequences (Table 2). This sequential arrangement is a natural association of the soils on the landscape and will be used along with the soil forming factors as a basis for discussing the characteristics, classification, and genesis of the soils of Nittany Valley.

Climate

On the broad scale, the climate of Nittany Valley and central Pennsylvania is classified as warm summer, humid continental (Trewartha, 1957). This classification only approximates the climate of Pennsylvania, and in particular, central Pennsylvania. The physiographic features found in the central part of the state have a marked affect on its weather and climate. The central part of the state, in particular the Ridge and Valley area, is not rugged enough for a true mountain type of climate, but it does have many of the characteristics of such a climate. The ridges and valleys influence air movement and cause somewhat greater temperature extremes than are experienced in the southeastern part of the state where the modifying coastal influence holds the temperature more constant. For example, the mean annual air temperature (MAAT) in State College at an elevation of 1170 feet is 50^{o}F while at Midstate Airport 14 air miles Northwest and at an elevation of 1918 feet, it is 45^{o}F. Midstate Airport is located on the backside of the Allegheny front and is somewhat representative of the ridge tops of central Pennsylvania. Precipitation also varies between the valley bottoms and the ridge tops. State College receives on the average 39 inches of precipitation per year while Midstate gets 45 inches. There is a slight seasonality of the precipitation

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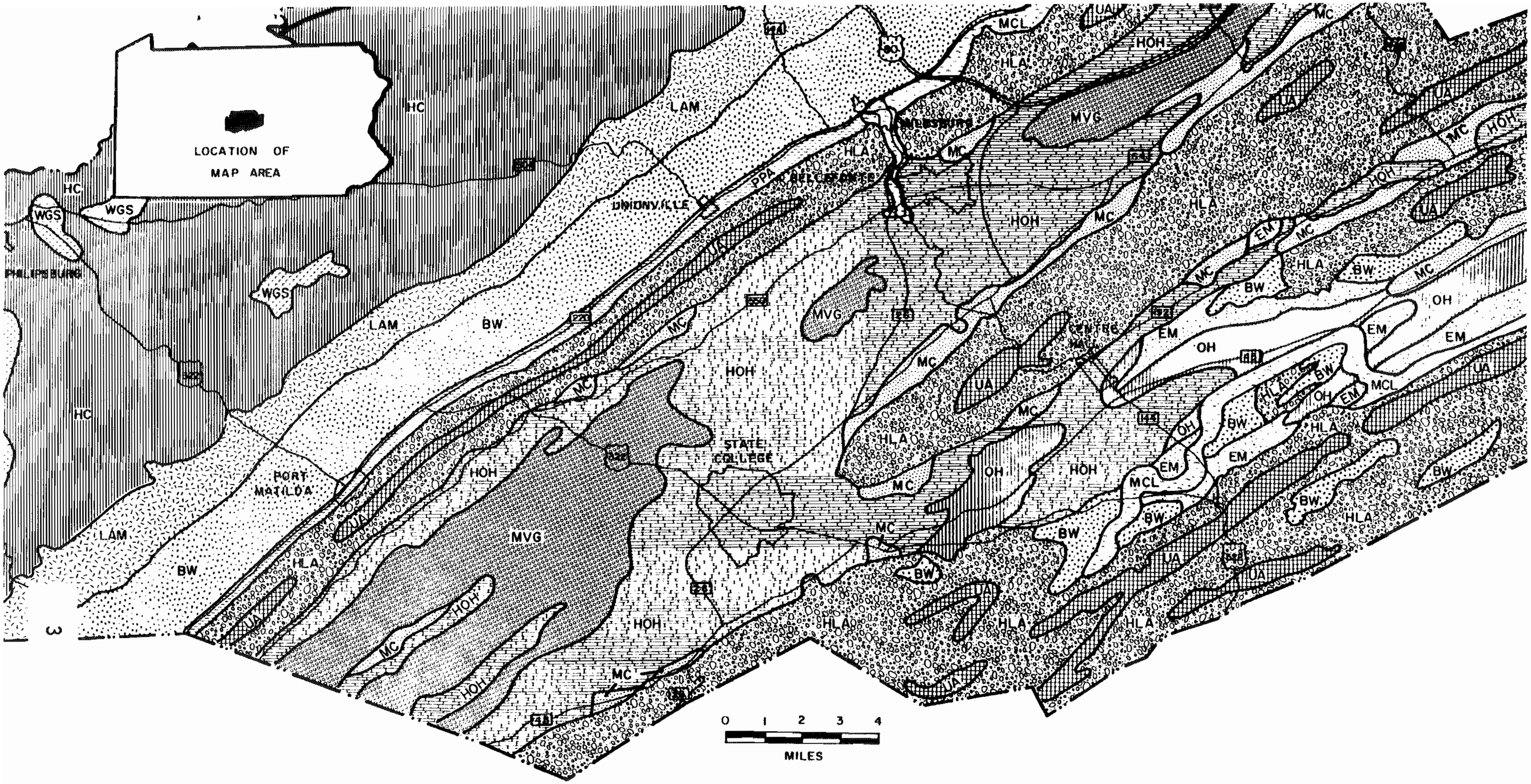
both on the ridges and in the valleys with the summer months receiving about 1 inch more per month than winter months.

In the winter season, central Pennsylvania receives about 1/3 of the available sunshine while in the summer the reverse is the case (2/3 of available sunshine). Thus summers (from the sunshine viewpoint) are the more pleasant season. When conditions are right, (primarily in the summer), the clouds show a banded appearance parallel to the ridges. This banded appearance is very evident on satellite imagery and from high flying aircraft. These clouds form as air rises over the ridges and the moisture condenses. The air then descends into the valleys and as it warms the condensed moisture is reabsorbed and the cloud dissipates. Thus the clouds are constantly forming and dissipating as the air moves over the ridges giving the banded cloud appearance. The differences in elevation also greatly affect the length of the frost free season on the ridges and in the valleys. In Nittany Valley, the frost free is about 50 days longer (about 160 vs. 110) than on the ridge tops. Even on the valley floor, the length of the growing season is somewhat variable. Hocevar and Martsolf (1971) report that on a clear, calm April night in Nittany Valley, the temperature can vary with the relief on the valley floor from 28^{o} to 39^{o}F. This variation is due to radiational heat loss and cold air drainage into the low areas. The cold air drainage, in addition to being a frost problem, also causes a significant amount of fog to form in the valleys in the spring and fall. The fall season in particular is foggy with October being the month in which the valleys of central Pennsylvania have the most days of fog. Another interesting elevational relationship is that if we use the rule of thumb that 1000 feet of elevation equals 300 miles of latitude, the ridge tops in this area would have a climate similar to that found in Ottawa, Canada. Additional climatic data are given in Table 3 and by Braker (1981).

These data indicate that the present day microclimate is quite variable in the Ridge and Valley area. These climate variations are small in comparison to the changes this area underwent during the Pleistocene and post-Pleistocene or Holocene time. During the Pre-Wisconsinan time (> 75,000 yrs), glacial ice came within 15 miles of Nittany Valley (Leverett, 1934; Marchand, 1978) and in the late Wisconsinan (Woodfordian) time, it came within 25 miles (Crowl and Sevon, 1979). During these ice advances, the climate was much different than today. It has been proposed that during these times the Ridge and Valley area had a tundra type of climate (Guilday et al., 1964; Martin, 1958; Watts, 1980; Maxwell and Davis, 1972). The presence of pingo scars (Marsh, 1987), ice wedge casts (Cronce, 1988), and other periglacial features (Ciolkosz et al., 1986) also indicates that permafrost was present. These data have lead Ciolkosz et al. (1989) to conclude that the MAAT in Nittany Valley varied from about 60^{o}F during the Sagononian interglacial (75,000-128,000 years ago) to about 15^{o}F during the Woodfordian glacial (20,000 years ago) to 50^{o}F today. This area's vegetation also has changed greatly particularly since the retreat of the Woodfordian glacial ice 18,500 years ago (Cotter et al., 1985).



SOILS OF THE VALLEYS FORMED IN RESIDUAL AND COLLUVIAL MATERIAL WEATHERED DOMINANTLY FROM LIMESTONE

- HOH Hagerstown-Opequon-Hublersburg association: Dominantly nearly level to sloping, deep and shallow, well drained soils underlain by limestone bedrock
- MVG Morrison-Vanderlip-Gatesburg association: Dominantly gently sloping to moderately steep, deep, well drained soils underlain by limey sandstone
- OH Opequon-Hagerstown association: Dominantly gently sloping and sloping, shallow and deep, well drained soils underlain by limestone bedrock
- MC Murrill-Clarksburg association: Dominantly nearly level to sloping, deep, well to somewhat poorly drained soils underlain by limestone bedrock
- EM Edom-Millheim association: Dominantly gently sloping and sloping deep, well drained soils underlain by calcareous shale bedrock

SOILS OF THE RIDGES THAT FORMED IN RESIDUAL AND COLLUVIAL MATERIAL WEATHERED FROM SANDSTONE AND SHALE

- HLA Hazleton-Laidig-Andover association: Dominantly gently sloping to very steep, deep, well drained and poorly drained soils underlain by brown acid sandstone and shale bedrock
- UA Ungers-Albrights association: Dominantly gently sloping to moderately steep, deep, well to somewhat poorly drained, soils underlain by red acid sandstone bedrock

SOILS OF THE VALLEY FLOODPLAINS

- PPA Pope-Philo-Atkins association: Dominantly gently sloping, deep, well to poorly drained soils developed in alluvium from acid sandstone and shale bedrock
- MCL Melvin-Chagrin-Lindside association: Dominantly gently sloping, deep, poorly to moderately well drained soils developed in alluvium from limestone bedrock

SOILS OF THE ALLEGHENY PLATEAU FORMED IN RESIDUAL MATERIAL WEATHERED FROM SANDSTONE AND SHALE

- HC Hazleton-Clymer association: Dominantly gently sloping to very steep, deep, well drained soils underlain by acid sandstone bedrock
- WGS Wharton-Gilpin-Strip Mines association. Dominantly gently sloping deep and moderately deep, moderately well drained and well drained soils and strip mines, underlain by acid shale bedrock

SOILS OF THE RIDGE AND VALLEY PROVINCE FORMED IN RESIDUAL MATERIAL WEATHERED FROM SHALE

- BW Berks-Weikert association: Dominantly sloping to very steep moderately deep and shallow, well drained soils underlain by brown acid shale bedrock
- LAM Leck Kill-Albrights-Meckesville association. Dominantly sloping to very steep, deep, well drained and moderately well drained soils underlain by red acid shale bedrock

Fig. 1. Soil association map of Southcentral Centre County.

Table 1 and Fig. 2. Soil associations of Pennsylvania.

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Map Symbol	Soil Series	Depth Class[†]	Drainage Class	Surface Texture	Subsoil Texture	Color	Parent Material	Classification
AR	Abbottstown	Deep**	Somewhat Poorly	Silt Loam	Silt Loam[+]	Grayish Red	Acid Red Shale	Aeric Fragiaqualf
	Readington	Deep**	Moderately Well	Silt Loam	Silt Loam[+]	Reddish Brown	Acid Red Shale	Typic Fragiudalf
BL	Berks	Mod. Deep	Well	Loam[+]	Loam[++]	Yellowish Brown	Acid Brown Shale	Typic Dystrochrept
	Leck Kill	Deep	Well	Silt Loam	Silt Loam[+]	Reddish Brown	Acid Red Shale	Typic Hapludult
BW	Berks	Mod. Deep	Well	Loam[+]	Loam[++]	Yellowish Brown	Acid Brown Shale	Typic Dystrochrept
	Weikert	Shallow	Well	Loam[+]	Loam[++]	Yellowish Brown	Acid Brown Shale	Lithic Dystrochrept
CB	Conotton	Deep	Well	Sandy Loam[+]	Sandy Loam[++]	Brown	Sand and Gravel	Typic Hapludult
	Birdsall	Deep	Very Poorly	Silt Loam	Silt Loam	Gray	Glacial Silts	Typic Humaquept
CC	Cavode	Deep	Somewhat Poorly	Silt Loam	Silty Clay	Grayish Brown	Acid Clay Shale	Aeric Ochraquult
	Cookport	Deep**	Moderately Well	Loam	Clay Loam[+]	Yellowish Brown	Acid Brown Shale	Aquic Fragiudult
CG	Chester	Deep	Well	Silt Loam	Silty Clay Loam	Brown	Gneiss and Schist	Typic Hapludult
	Glenelg	Deep	Well	Loam[+]	Silt Loam[+]	Brown	Gneiss and Schist	Typic Hapludult
DH	Duffield	Deep	Well	Silt Loam	Silty Clay Loam	Yellowish Brown	Shaly Limestone	Ultic Hapludalf
	Hagerstown	Deep	Well	Silt Loam	Clay	Red	Limestone	Typic Hapludalf
EH	Edgemont	Deep	Well	Sandy Loam[+]	Loam[+]	Yellowish Brown	Quartzite	Typic Hapludult
	Highfield	Deep	Well	Silt Loam[+]	Silt Loam[+]	Yellowish Brown	Metarhyolite	Ultic Hapludalf
EL	Erie	Deep*	Somewhat Poorly	Silt Loam[+]	Loam[+]	Grayish Brown	Calcareous Till	Aeric Fragiaqualf
	Langford	Deep**	Moderately Well	Silt Loam[+]	Loam[+]	Yellowish Brown	Calcareous Till	Aqueptic Fragiudalf
DC	Dormont	Deep	Moderately Well	Silt Loam	Silty Clay Loam[+]	Yellowish Brown	Limestone and Shale	Ultic Hapludalf
	Culleoka	Mod. Deep	Well	Silt Loam	Silt Loam[+]	Brown	Limestone and Shale	Ultic Hapludalf
GW	Gilpin	Mod. Deep	Well	Silt Loam[+]	Silt Loam[+]	Yellowish Brown	Shale and Sandstone	Typic Hapludult
	Wharton	Deep	Moderately Well	Silt Loam	Silty Clay Loam[+]	Brown	Shale and Siltstone	Aquic Hapludult
HA	Hanover	Deep**	Well-Mod. Well	Silt Loam[+]	Silt Loam[+]	Yellowish Brown	Leached Till	Typic Fragiudult
	Alvira	Deep**	Somewhat Poorly	Silt Loam[+]	Silt Loam[+]	Grayish Brown	Leached Till	Aeric Fragiaquult
HC	Hazleton	Deep	Well	Sandy Loam[+]	Sandy Loam[++]	Yellowish Brown	Acid Sandstone	Typic Dystrochrept
	Cookport	Deep**	Moderately Well	Loam	Clay Loam[+]	Yellowish Brown	Acid Sandstone	Aquic Fragiudult
HD	Hagerstown	Deep	Well	Silt Loam	Clay	Red	Limestone	Typic Hapludalf
	Duffield	Deep	Well	Silt Loam	Silty Clay Loam	Yellowish Brown	Shaly Limestone	Ultic Hapludalf
HE	Hagerstown	Deep	Well	Silt Loam	Clay	Red	Limestone	Typic Hapludalf
	Edom	Deep	Well	Silty Clay Loam	Clay[+]	Yellowish Brown	Shaly Limestone	Typic Hapludalf
HL	Hazleton	Deep	Well	Sandy Loam[+]	Sandy Loam[++]	Yellowish Brown	Acid Sandstone	Typic Dystrochrept
	Laidig	Deep**	Well	Loam[+]	Loam[+]	Brown	Sandstone Colluvium	Typic Fragiudult
HP	Howell	Deep	Well	Sandy Loam	Clay	Brown	Sand, Silt and Clay	Typic Hapludult
	Pope	Deep	Well	Loam	Loam	Yellowish Brown	Silty Alluvium	Fluventic Dystrochrept
LM	Leck Kill	Deep	Well	Silt Loam	Silt Loam[+]	Reddish Brown	Acid Red Shale	Typic Hapludult
	Meckesville	Deep**	Well	Loam	Clay Loam	Reddish Brown	Red Shale Colluvium	Typic Fragiudult
LO	Lordstown	Mod. Deep	Well	Silt Loam[+]	Silt Loam[+]	Yellowish Brown	Acid Brown Till	Typic Dystrochrept
	Oquaqa	Mod. Deep	Well	Loam[+]	Loam[+]	Reddish Brown	Acid Brown Till	Typic Dystrochrept
MV	Morrison	Deep	Well	Sandy Loam	Sandy Clay Loam[+]	Brown	Sandy Limestone	Ultic Hapludalf
	Vanderlip	Deep	Well	Loamy Sand	Loamy Sand[+]	Yellowish Brown	Sandy Limestone	Typic Quartzipsamment
NL	Neshaminy	Deep	Well	Silt Loam	Clay Loam[+]	Yellowish Red	Diabase	Ultic Hapludalf
	Lehigh	Deep	Mod. Well-S.W. Poorly	Silt Loam	Silt Loam[+]	Gray	Metamorphosed Shale	Aquic Hapludalf
PL	Penn	Mod. Deep	Well	Silt Loam[+]	Silt Loam[+]	Reddish Brown	Red Shale	Ultic Hapludalf
	Lewisberry	Deep	Well	Sandy Loam[+]	Sandy Loam[+]	Reddish Brown	Red Sandstone	Ultic Hapludalf
RC	Ravenna	Deep**	Somewhat Poorly	Silt Loam	Loam[+]	Grayish Brown	Neutral Till	Aeric Fragiaqualf
	Canfield	Deep**	Moderately Well	Silt Loam	Loam[+]	Yellowish Brown	Neutral Till	Aquic Fragiudalf
SP	Sheffield	Deep**	Poorly	Silt Loam	Silty Clay Loam	Brownish Gray	Fine Textured Till	Typic Fragiaqualf
	Platea	Deep**	Somewhat Poorly	Silt Loam	Silt Loam	Grayish Brown	Fine Textured Till	Aeric Fragiaqualf
VC	Venango	Deep**	Somewhat Poorly	Silt Loam	Loam[+]	Grayish Brown	Calcareous Till	Aeric Fragiaqualf
	Cambridge	Deep**	Moderately Well	Silt Loam	Loam[+]	Yellowish Brown	Calcareous Till	Aqueptic Fragiudalf
VM	Volusia	Deep*	Somewhat Poorly	Silt Loam[+]	Silt Loam[+]	Grayish Brown	Acid Brown Till	Aeric Fragiaquept
	Morris	Deep*	Somewhat Poorly	Loam[+]	Loam[+]	Grayish Red	Acid Red Till	Aeric Fragiaquept

*Fragipan at 10-16 inches from the soil surface; **Fragipan at 16-36 inches from the soil surface

[+]Some (15-35%) rock fragments; [++]Many (>35%) rock fragments

[†]Depth to bedrock: Shallow <20"; Mod. Deep 20-40"; Deep >40"

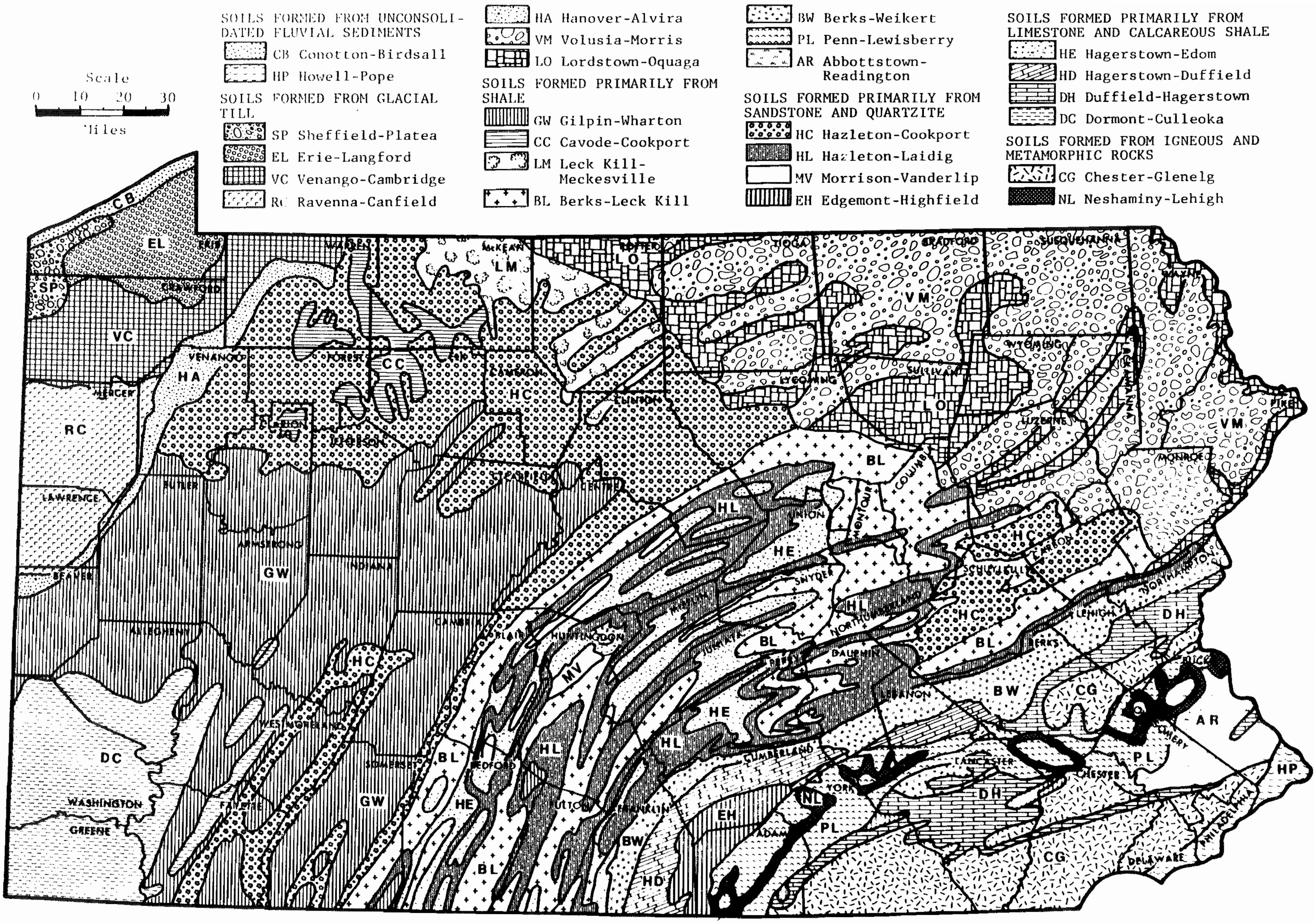

Scale
0 10 20 30
Miles
SOILS FORMED FROM UNCONSOLIDATED FLUVIAL SEDIMENTS
CB Conotton-Birdsall
HP Howell-Pope
SOILS FORMED FROM GLACIAL TILL
SP Sheffield-Platea
EL Erie-Langford
VC Venango-Cambridge
RC Ravenna-Canfield
HA Hanover-Alvira
VM Volusia-Morris
LO Lordstown-Oquaga
SOILS FORMED PRIMARILY FROM SHALE
GW Gilpin-Wharton
CC Cavode-Cookport
LM Leck Kill-Meckesville
BL Berks-Leck Kill
BW Berks-Weikert
PL Penn-Lewisberry
AR Abbottstown-Readington
SOILS FORMED PRIMARILY FROM SANDSTONE AND QUARTZITE
HC Hazleton-Cookport
HL Hazleton-Laidig
MV Morrison-Vanderlip
EH Edgemont-Highfield
SOILS FORMED PRIMARILY FROM LIMESTONE AND CALCAREOUS SHALE
HE Hagerstown-Edom
HD Hagerstown-Duffield
DH Duffield-Hagerstown
DC Dormont-Culleoka
SOILS FORMED FROM IGNEOUS AND METAMORPHIC ROCKS
CG Chester-Glenelg
NL Neshaminy-Lehigh

Table 2. Major Soils of Central Pennsylvania Arranged According to Parent Material and Drainage.*

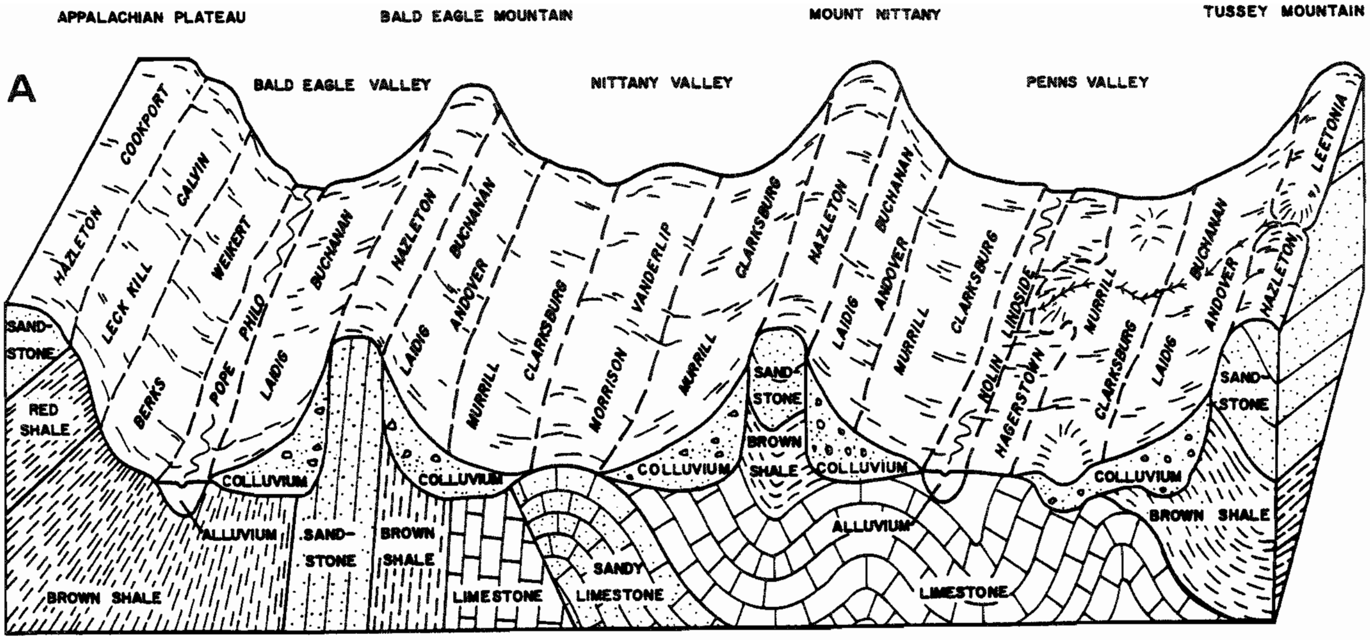
Parent Material	(Shallow) <20" to bedrock	(Moderately Deep) 20-40" to bedrock	(------ Deep >40" to consolidated bedrock ------)				
	Drainage Class and Depth to Mottling						
	(------ Well Drained ------) (------ >40" ------)			Moderately Well Drained (20-40")	Somewhat Poorly Drained (10-20")	Poorly Drained 0-10"; some gleying)	Very Poorly Drained (0-10"; strong gleying)
Residual							
Gray and brown acid shale and siltstone	Weikert Lithic Dystrochrept; loamy-skeletal	Berks Typic Dystrochrept; loamy-skeletal	Bedington Typic Hapludult; fine-loamy Hartleton Typic Hapludult; loamy-skeletal	Blairton [Mod. deep------] Aquic Hapludult; fine-loamy Comly------) Typic Fragiudalf; fine-loamy		Markes [Mod. deep] Typic Ochraqualf; loamy-skeletal Brinkerton Typic Fragiaqualf; fine-silty	
Gray and brown acid shale siltstone and some clay shale	Weikert Lithic Dystrochrept; loamy-skeletal	Gilpin Typic Hapludult fine-loamy	Rayne Typic Hapludult; fine-loamy	Wharton Aquic Hapludult; clayey	Cavode Aeric Ochraquult; clayey	Armagh Typic Ochraquult; clayey	
Red acid shale and siltstone; dull red 4 chroma or less	Klinesville Lithic Dystrochrept; loamy-skeletal	Calvin Typic Dystrochrept; loamy-skeletal	Leck Kill Typic Hapludult; fine-loamy	Albrights------) Aquic Fragiudalf; fine-loamy		Conyngham Unclassified	
Gray and brown acid sandstone	Ramsey Lithic Dystrochrept; loamy-skeletal †	Dekalb Typic Dystrochrept; loamy-skeletal	Hazleton Typic Dystrochrept; loamy-skeletal Clymer Typic Hapludult; fine-loamy	Cookport------) Aquic Fragiudult; fine-loamy		Nolo Typic Fragiaquult; fine-loamy	Lickdale Humic Haplaquept; fine-loamy
Red acid sandstone; dull red 4 chroma or less	Ramsey Lithic Dystrochrept; loamy-skeletal †	Lehew Typic Dystrochrept; loamy-skeletal	Ungers Typic Hapludult; fine-loamy	Albrights------) Aquic Fragiudalf; fine-loamy		Conyngham Unclassified	
Grayish brown sandstone (in some places a very sandy limestone)			Vanderlip Typic Quartzipsamment; coated Morrison Ultic Hapludalf; fine-loamy Leetonia Entic Haplorthod; sandy-skeletal, siliceous Gatesburg Entic Haplorthod; sandy, siliceous				
Very cherty limestone			Elliber Typic Hapludult; loamy-skeletal	Kreamer Aquic Hapludult; clayey	Evendale------) Aeric Ochraquult; clayey		
Cherty limestone			Hublersburg Typic Hapludult; clayey				
Relatively pure limestone	Oqequon Lithic Hapludalf; clayey		Hagerstown Typic Hapludalf; clayey**	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	

Parent Material	(Shallow) <20" to bedrock	(Moderately Deep) 20-40" to bedrock	(------- Deep >40" to consolidated bedrock -------) Drainage Class and Depth to Mottling (------- Well Drained -------) (------- >40" -------)	Moderately Well Drained (20-40")	Somewhat Poorly Drained (10-20")	Poorly Drained (0-10"; some gleying)	Very Poorly Drained (0-10"; strong gleying)
Residual (cont'd.)							
Thin bedded limestone and calcareous shale		Ryder Ultic Hapludalf; fine-loamy	Duffield Ultic Hapludalf; fine-loamy Edom Typic Hapludalf; clayey**, illitic Frankstown Typic Hapludult; fine-loamy	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	
Colluvium							
Brown and gray acid shale, siltstone and fine grain sandstone			Shelocta Typic Hapludult; fine-loamy	Ernest---------) Aquic Fragiudult; fine-loamy		Brinkerton Typic Fragiaqualf; fine-silty	
Red acid shale, siltstone and fine grain sandstone; dull red chroma 4 or less			Meckesville Typic Fragiudult; fine-loamy	Albrights---------) Aquic Fragiudalf; fine-loamy		Conyngham Unclassified	
Gray and brown acid sandstone and shale			Laidig Typic Fragiudult; fine-loamy	Buchanan---------) Aquic Fragiudult; fine-loamy		Andover Typic Fragiaquult; fine-loamy	
Brown and gray limestone, shale and sandstone			Murrill Typic Hapludult; fine-loamy	Clarksburg Typic Fragiudalf; fine-loamy	Penlaw Aquic Fragiudalf; fine-silty	Thorndale Typic Fragiaqualf; fine-silty	
Very cherty limestone and shale			Mertz Typic Hapludult; loamy-skeletal	Kreamer Aquic Hapludult; clayey	Evendale---------) Aeric Ochraquult; clayey		
Recent Alluvium (Floodplains)							
Alluvium from acid gray and brown shale, siltstone and sandstone uplands.			Pope Fluventic Dystrochrept; coarse-loamy	Philo Fluvaquentic Dystrochrept; coarse-loamy	Stendal Aeric Fluvaquent; fine-silty	Atkins Typic Fluvaquent; fine-loamy	Elkins Humaqueptic Fluvaquent; fine-silty
Alluvium from acid red shale, siltstone and sandstone uplands; dull red chroma 4 or less			Barbour; Linden Fluventic Dystrochrept; coarse-loamy	Basher---------) Fluvaquentic Dystrochrept; coarse-loamy		Holly Typic Fluvaquent; fine-loamy	Papakating Mollic Fluvaquent; fine-silty
Alluvium from limestone, shale and siltstone upland			Nolin Dystric Fluventic Eutrochrept; fine-silty	Lindside Fluvaquentic Eutrochrept; fine-silty	Newark Aeric Fluvaquent; fine-silty	Melvin Typic Fluvaquent; fine-silty	Dunning Fluvaquentic Haplaquoll; clayey**
Old Alluvium (Terraces) and Lacustrine							
Gray and brown acid shale siltstone and sandstone			Allegheny Typic Hapludult; fine-loamy	Monongahela Typic Fragiudult; fine-loamy	Tyler-Aeric Fragiaquult; fine-silty Tygart - Aeric Ochraquult; clayey	Purdy Typic Ochraquult; clayey	

* Almost all soils are also mixed, mesic.

** These soils are classified in the fine family but for the purpose of this table clayey will be used.

† In Pennsylvania most pedons are skeletal.



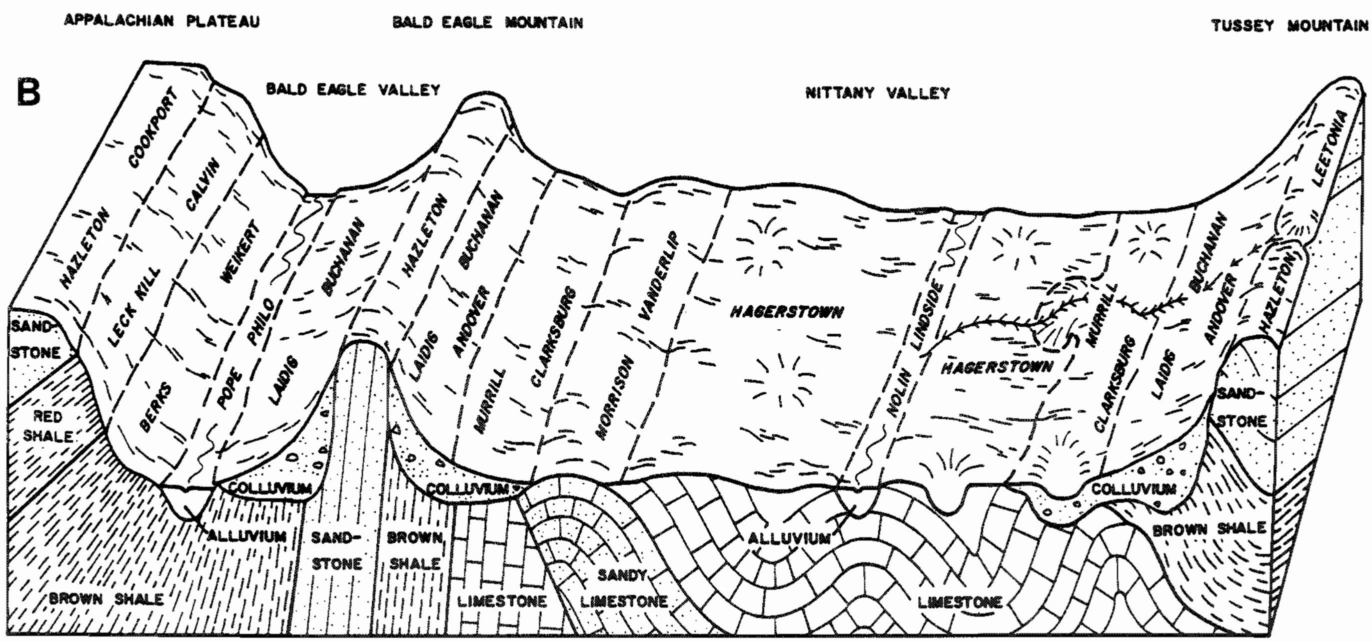


Fig. 3. General soil-landscape relations of Nittany Valley (A is northeast and B is southwest of State College).

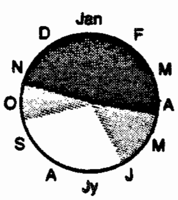
Table 3. Climatic data for Fort Collins, CO (just north of Denver); State College, PA; Phoenix, AZ; and Orlando, FL. In the pie diagrams, winter equals the mean daily low temperature below 32°F (dark shade), spring and fall equal the mean daily temperature between 32° and 60°F (light shade), and summer equals the daily mean > 60°F

FORT COLLINS, CO

Elevation: 5,004 feet

Relative Humidity: 60%
Wind Speed: 9 mph

Seasonal Change



Annual Rainfall 15 in

Annual Snowfall 41 in

Clear 118 days | Partly Cloudy 128 days | Cloudy 119 days

Precipitation Days: 37 Storm Days: 50

Average Temperatures

	Daily High	Daily Low	Monthly Mean
January	40.3	11.9	26.1
February	42.5	14.6	28.6
March	49.7	22.2	36.0
April	60.1	32.1	46.1
May	68.0	40.8	54.4
June	78.4	48.9	63.7
July	84.4	54.4	69.4
August	83.2	52.7	68.0
September	75.6	43.8	59.7
October	64.3	32.8	48.6
November	51.1	21.6	36.4
December	42.3	14.3	28.3

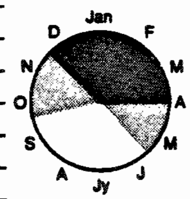
Zero-Degree Days: 15
Freezing Days: 175
90-Degree Days: 17
Heating- and Cooling-Degree Days: 7,052

STATE COLLEGE, PA

Elevation: 1,200 feet

Relative Humidity: 67%
Wind Speed: 7.8 mph

Seasonal Change



Annual Rainfall 37 in

Annual Snowfall 48 in

Clear 66 days | Partly Cloudy 114 days | Cloudy 185 days

Precipitation Days: 122 Storm Days: 35

Average Temperatures

	Daily High	Daily Low	Monthly Mean
January	34.2	19.8	27.0
February	36.1	20.2	28.2
March	45.4	27.7	36.5
April	59.2	38.9	49.1
May	70.2	48.8	59.3
June	78.7	57.3	68.0
July	82.6	61.1	71.9
August	80.7	59.1	69.9
September	73.5	52.0	62.8
October	62.9	42.5	52.7
November	48.7	33.2	41.0
December	36.3	22.9	29.6

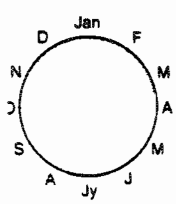
Zero-Degree Days: 4
Freezing Days: 132
90-Degree Days: 8
Heating- and Cooling-Degree Days: 6,797

PHOENIX, AZ

Elevation: 1,107 feet

Relative Humidity: 36%
Wind Speed: 6.2 mph

Seasonal Change



Annual Rainfall 7 in

Annual Snowfall 0 in

Clear 214 days | Partly Cloudy 81 days | Cloudy 70 days

Precipitation Days: 34 Storm Days: 23

Average Temperatures

	Daily High	Daily Low	Monthly Mean
January	64.8	37.6	51.2
February	69.3	40.8	55.1
March	74.5	44.8	59.7
April	83.6	51.8	67.7
May	92.9	59.6	76.3
June	101.5	67.7	84.6
July	104.8	77.5	91.2
August	102.2	76.0	89.1
September	98.4	69.1	83.8
October	87.6	56.8	72.2
November	74.7	44.8	59.8
December	66.4	38.5	52.5

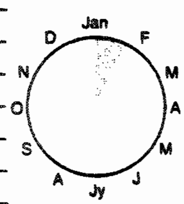
Zero-Degree Days: 0
Freezing Days: 32
90-Degree Days: 164
Heating- and Cooling-Degree Days: 5,060

ORLANDO, FL

Elevation: 106 feet

Relative Humidity: 74%
Wind Speed: 8.7 mph

Seasonal Change



Annual Rainfall 51 in

Annual Snowfall 0 in

Clear 94 days | Partly Cloudy 148 days | Cloudy 123 days

Precipitation Days: 116 Storm Days: 81

Average Temperatures

	Daily High	Daily Low	Monthly Mean
January	70.5	50.0	60.3
February	71.8	51.2	61.5
March	76.0	55.7	65.9
April	81.5	61.1	71.3
May	86.7	66.1	76.4
June	89.3	71.1	80.2
July	89.8	72.9	81.4
August	90.0	73.5	81.8
September	87.9	72.3	80.1
October	82.5	66.0	74.3
November	76.2	56.9	66.6
December	71.5	51.5	61.5

Zero-Degree Days: 0
Freezing Days: 2
90-Degree Days: 104
Heating- and Cooling-Degree Days: 3,959

Vegetation

Braun (1950) indicates that the present day natural vegetation of Nittany Valley falls within the oak-chestnut association of the Ridge and Valley Province. The oak-chestnut association occurs in very close proximity to the hemlock-white pine northern hardwoods association which begins on the Allegheny Plateau some ten miles to the northwest of State College. Braun (1950) also indicates that approximately ninety miles to the southwest, the mixed mesophytic association of the south reaches its northern limits in Bedford County, Pennsylvania, and an extension of the Blue Ridge section of the oak-chestnut association extends northward into the lower tier counties of south central Pennsylvania.

Although chestnut was a major (20-30%) part of the Appalachian forest at the time of settlement, today it is only found as stump sprouts which grow to a height of 10 to 20 feet and then die. The reason for this is that in 1904 chestnut blight (a fungus) was introduced into the United States via New York City from Asia, and in 40 years, all chestnut trees from Maine to Georgia and westward to Ohio and Tennessee were killed (Cochran, 1990). It has been estimated that 3.5 billion trees were killed. Chestnut was a very valuable tree for both its nuts and wood, and efforts are underway in crossbreeding with resistant Chinese and Japanese chestnuts and with the introduction of a fungus attacking virus to restore the chestnut to eastern forests. More detailed information on the vegetation of Nittany Valley is given by Shipman (1980) and Baldwin (1961).

One unique form of vegetation in the Nittany Valley area is that of tall grass prairie. At the time of settlement, an area near Centre Hall in Penns Valley of about 4,000-5,000 acres was vegetated with native prairie grasses (Losensky, 1961). This prairie area as well as others noted in Pennsylvania (Losensky, 1961) may well be remnants of more extensive prairie areas that were established during the warmer, drier Hypsithermal period which occurred 4,000 to 7,000 years ago (Schmidt, 1938; Guilday et al., 1964). According to Flint (1971) during the Hypsithermal, the mean annual temperature was about $4^{o}F$ higher and the mean annual precipitation was about 5 inches less than today. This would give the valleys, during the Hypsithermal time, a climate similar to today's climate in eastern Kansas, which is an area that has tall grass prairie as its native vegetation. The presence of prairie vegetation poses some interesting possibilities. Do these areas have Mollisol soils (soils of the tall grass prairies of the Midwest)? The soil survey of Centre County (Braker, 1981) does not identify any Mollisols in the Centre Hall area. In addition, field investigations by Waltman (1988) in the summer of 1982 indicated no soils in this area with Mollisol morphology. Although this is the case, a more detailed study may indicate a significant effect of prairie vegetation on the soils of the Centre Hall area and even possibly that these soils should have been classified as Mollisols. This may well be the case, for Waltman (1988) found that prairie vegetation did have a significant impact on the development of an area of soils near Meadville, Pennsylvania (Northwestern PA).

Soils Developed in Residuum

Generally, in the Ridge and Valley area of Pennsylvania, soils formed in residuum occupy 67% of the area while soils formed in colluvium occupy 27%, and soils formed in fluvial deposits (floodplain and terrace) occupy the remaining 6% of the area (Table 4). Soils developed in residuum are found on the ridge tops and on the valley floors in the Nittany Valley area. The ridge tops are usually relatively narrow and have hard sandstone as the underlying bedrock. On the valley floors, the soils are developed primarily in limestones and dolomites and interbedded sandstones and dolomite. In a few places in the valley and on some side slopes where the colluvial mantle is absent, soils developed in acid gray shales are found.

Table 4. The relative proportion of colluvial, fluvial (floodplain and terrace), and residual soils in four counties in Pennsylvania (from Ciolkosz and Dobos, 1990).

Physiographic Area and County	Percent		
	Colluvial	Fluvial	Residual
Ridge and Valley			
Fulton	27.2	6.2	66.6
Huntingdon	27.3	6.4	66.3
Plateau			
Fayette	14.3	4.6	81.1
Westmoreland	12.5	8.5	79.0

Soils from Sandstone

The major soils found on the ridge tops are Hazleton and Cookport. Some Leetonia is also found but its distribution is very irregular. Hazleton is by far the most extensive soil found on the ridge tops, and it is also the most extensive soil found in Pennsylvania (Table 5). Hazleton is a deep soil (> 40" to bedrock). In past soil surveys (Mooney et al., 1910), much of the ridge tops and large areas on the Appalachian Plateau to the west were thought to be moderately deep to bedrock (20-40") and the soils in these areas were classified as Dekalb. Many observations of excavations made with power equipment has indicated that these soils are deep to bedrock and they have a high content of rock fragments. When making observations with hand tools, the high content of rock fragments give the impression that the soil is shallow or moderately deep. The cracked and fractured nature of the sandstone bedrock of these soils apparently aids in the development of deep soils. These cracks and fractures act as zones of weakness to the weathering processes and allow the weathering to follow the zones creating deep soils with a high percentage of rock fragments. Although the soils are deep (> 40"), the bedrock is usually found at depths of 6 to 8 feet although in some areas it may be 10 to 20 feet deep (Carter and Ciolkosz, 1986). Thus, these very hard parent materials greatly resist deep weathering when compared to other parent materials in this area.

Table 5. Ranking, acreage, and percent on a state basis of the common soils in in the Nittany Valley area. Pennsylvania has a total of 28.9 million acres. Data from Cunningham and Day (1986).

Soil	Rank According to Acreage	Acres in PA	Percent of PA Soils	Soil	Rank According to Acreage	Acres in PA	Percent of PA Soils
Hazleton	1	2,945,000	10.39	Leetonia	55	123,000	0.43
Gilpin	2	1,347,000	4.75	Murrill	57	120,000	0.43
Weikert	4	856,000	3.02	Andover	60	112,000	0.40
Cookport	6	854,000	3.01	Clarksburg	67	90,000	0.32
Berks	7	831,000	2.93	Morrison	69	87,000	0.31
Laidig	11	605,000	2.14	Pope	72	84,000	0.30
Buchanan	14	509,000	1.80	Opequon	75	81,000	0.29
Hagerstown	18	425,000	1.50	Melvin	113	40,000	0.14
Leck Kill	20	384,000	1.36	Lindside	116	39,000	0.14
Calvin	31	249,000	0.88	Nolin	121	35,000	0.12
Duffield	37	206,000	0.73	Vanderlip	171	15,000	0.05
Philo	48	144,000	0.51				

The Hazleton soil is classified as an Inceptisol (Dystrochrept) which means it shows weak soil development (Cambic B horizon). The weak development is more an indication of a parent material that is very resistant to weathering and soil formation than of age because these soils are believed to be about 15,000 years old (Ciolkosz et al., 1990b). Further evidence supporting this conclusion is the juxtaposition of Cookport soils with Hazleton soils. Cookport soils are classified as Ultisols (Fragiudults--fragipan and argillic horizon) which indicates moderate development. Although found associated with Hazleton soils, Cookport soils usually are located in low lying or depressional areas that may have had some finer material washed in or they are on large flat areas which have shale material interbedded with the sandstone. Thus, the argillic horizon and fragipan of the Cookport are a reflection of parent material more than of any other soil forming factor.

Leetonia soils are found on the ridge tops and are classified as Spodosols. Their sparce occurrence (only about 3% of the ridge top soil area) is apparently associated with scattered occurrences of coniferous vegetation on the ridge tops (Ciolkosz et al., 1990b).

Soils from Limestone

Two groups of soils dominate the residual soils of the valley floor. These are the soils developed from limestone and those developed from interbedded sandstone and limestone.

The major soils developed from limestone are Opequon and Hagerstown. The Opequon is shallow (< 20") to bedrock while Hagerstown is deep (> 40") to bedrock. These soils are well drained, red in color, and have a clayey, argillic B horizon. They are found in karst topography areas with many sink holes and a limited integrated drainage network on the landscape. A good part of the runoff of these soils does not drain directly from the land but it drains into sink holes, and then into the ground water. It is then discharged via springs into streams (see Parizek et al., 1971; Wood, 1980; and White, 1988 for a discussion of karst hydrology).

Hagerstown and Opequon soils are developed from limestone and it is assumed that the soil material found at the surface is residuum from the limestone. Table 6 gives soil and rock data from Hagerstown and Duffield soils from Nittany Valley and southeastern Pennsylvania. Duffield is a siltier, browner soil than Hagerstown, but it is also developed from limestone.

These data and other Pennsylvania data (total of 15 pedons of Hagerstown and 10 of Duffield--not all pedons were analyzed for free iron oxides) indicate a range of free iron oxides in B horizons of 4 to 7% for Hagerstown and 2 to 4% for Duffield soils (Table 6). From these data a logical conclusion might be that the higher the iron oxide content, the redder the soil. Observations of other soils and data in the literature does not consistently support this conclusion. Iron oxides are the main coloring agents in soils below the surface horizon, and Schwertmann (1988) lists in addition to quantity of iron oxide, the type of iron mineral, and its crystal size as factors influencing soil color. Hematite and goethite are the major iron oxide minerals in soils with hematite imparting red colors and goethite imparting brown colors (Schwertmann and Taylor, 1977; Schwertmann, 1988) to the soil. Schwertmann (1988) also indicates that hematite has greater pigmenting power than goethite. No data are available to indicate the major iron oxide mineral in Hagerstown or Duffield soils. Although no data are available for these soils, a review of the literature (see Dobos, 1986) indicates that soils with yellow hues (7.5 YR or 10 YR) have 10% or less of hematite. Thus, the color of Duffield soils is apparently due to goethite, and the color of Hagerstown is due primarily to hematite (hues of 2.5 YR probably indicate about 50% hematite and 50% goethite). This suggests an interesting question. If the iron oxide mineralogy between these soils is different, why is it different? According to Schwertmann and Taylor (1977), in humid temperate zones that have cool, wet, low pH, high organic matter soil conditions goethite formation is favored over hematite. The application of these factors to the Hagerstown-Duffield soil color question is not clear-cut but the work of Hsu and Wang (1980) which indicates a high iron solution concentration favors the formation of hematite over goethite may explain the differences noted. A recent report by Kampf and Schwertmann (1985) also supports the contention that a high release rate of iron during rock weathering favors the formation of hematite over goethite. The higher content of iron oxides in the Hagerstown than in the Duffield may indicate a higher content of iron in the weathering solution which would favor hematite formation over goethite, and explain the color difference between these two soils. Although it may be a factor, limited data do not indicate that the type of limestone (calcic vs. dolomitic) is a factor in the development of Hagerstown and Duffield soils.

Another interesting question about the Hagerstown soil concerns the origin of the soil material. If all the material is of residual origin, then it would take about 1,000 feet of bedrock to give about 10 feet of soil (2-8% acid insoluble residue--Table 6). The time required to accumulate these residual materials is also an interesting question. Studies of limestone tombstone weathering give limestone dissolution rates of 10 to 100 mm/1,000 yrs. (Colman, 1981; Mejerding, 1981). Trudgill (1976, 1985) gives limestone dissolution rates of 1,000 to 5,000 mm/1,000 yrs. for limestone under calcareous brown earth soils. These rates are a little higher than those given by Saunders and Young (1983) of 20 to 100 mm/1,000 yrs. or by Jennings (1983, 1985) of 5 to 30 mm/1,000 yrs. Additional data are given in Ollier (1984); and for Nittany Valley, Parizek and White (1985) give a rate of 30 mm/1,000 yrs. It is interesting to note that over 100 years ago, Ewing (1885) published a limestone dissolution rate of 27 mm/1,000 yrs. for Nittany Valley. Insoluble residue data can be used to compute some general accumulation rates for the residual soils of the valley (Table 7). Residual material on the stable landscape surfaces in the valley varies from 2 to 5 meters in thickness. This would give an age of about 1 to 2 million years for these soils using Parizek and White's (1985) dissolution rate of 30 mm/1,000 yrs. An estimate of 2.5 to 5 million yrs. for 5 meters of limestone residual soil accumulation in the great valley near Harrisburg has been given by Sevon (1985). Thus an age of 1 to 2 million years for the deep limestone soils of Nittany Valley seems reasonable.

The insoluble residue of the Hagerstown soil is also apparently mainly clay size material. Duffield, in addition to contrasting with Hagerstown in color, also contrasts in subsoil texture (silty vs. clayey). Limited laboratory data do not indicate if Duffield has a higher concentration of acid insoluble residue (field observations do support this contention) but apparently most of it is silt and sand size material. The high clay content of the Hagerstown B also contrasts strongly with the low clay content of its A horizon (Table 6). There can be a 40-70% increase in clay content from the A to the B, particularly the lower B. This increase has in the past been attributed to eluviation of clay from the A into the B during the formation of an argillic horizon. Observations of clay coatings in the B of the Hagerstown indicate that some clay has been illuviated in this soil. Another possibility is that aeolian additions of silt size material may have accumulated on the surface of the Hagerstown soil. The silt size material would dilute the clay content and exaggerate the difference in clay content between the A and B horizons. Aeolian loess (Carey, Cunningham, and Williams, 1976; Millette and Higbee, 1958) and sand deposits (Marchand et al., 1978) are known in various parts of Pennsylvania. In addition, the work of Jackson et al. (1971) and Smith et al. (1970) indicate that significant amounts of dust have been added to soils in the Eastern United States, and the work of Cronce (1988) in Nittany Valley indicates that a significant amount of dust has been added to soils in central Pennsylvania. Thus, the silty surface of the Hagerstown is due both to aeolian additions and some clay eluviation.

Table 6. Soil and rock data for selected soils developed from limestone.

Horizon	Depth Inches	Color	Rock Mineral	Percent Fe_2O_3	Acid Insoluble residue	Sand	Silt	Clay
				Hagerstown				
Ap*	0-7	10YR 3/2		2.5		14.1	64.5	21.4
Bt	7-38	5YR 4/6		4.6		19.7	41.5	38.8
C	38-40	5YR 3/3		6.0		20.9	24.3	54.8
R	40+	N 4/0	dolomite	0.4	3.5			
Ap*	0-7	10YR 2/2		2.1		11.4	64.2	24.5
Bt	7-36	5YR 4/6		3.9		7.8	39.1	53.1
R	36+	N 5/0	dolomite	0.5	8.7			
Ap*	0-8	10YR 3/2		2.6		15.0	65.6	19.4
Bt	8-30	5YR 4/6		5.2		6.0	27.0	67.0
R	40+	N 3/0	calcite	0.2	2.6			
R*		N 5/0	dolomite	1.2	6.4			
R**			dolomite		5.8			
Bt3*	33-46	2.5YR 4/6		5.8		8.2	31.7	60.1
Bt2*	23-33	5YR 5/6		6.9		9.7	31.2	59.1
Bt3*	28-34	5YR 5/4		4.4		7.6	43.6	48.7
				Duffield				
Ap*	0-13	10YR 3/3		2.0		5.7	73.8	20.5
Bt2	22-30	7.5YR 5/6		3.6		11.0	53.6	35.4
C1	52-67	10YR 5/6		2.4		15.8	56.1	28.2
R	80+	N 6/0	dolomite	1.9	11.8			
Bt3*	31-39	7.5YR 5/6		4.4		9.3	64.2	26.5
Bt2*	22-33	7.5YR 5/8				13.0	52.7	34.3
Bt3*	34-46	7.5YR 5/8				16.6	52.2	31.2
Bt4*	34-48	10YR 4/4		2.0		13.0	51.2	35.8
Bt2*	29-40	10YR 5/6		2.3		29.6	44.8	25.6

*Soil Characterization Laboratory (1986)
**Jeffries and White (1940)

Table 7. Accumulation of residual soil from a limestone with 6% acid insoluble residues, assuming varying rates of dissolution over time. A bulk density of 2.85 (g/cc) for the rock and 1.65 for the soil is used in the calculations.

Dissolution rate mm/1,000 yrs.	Time (years)			
	10,000	100,000	1,000,000	2,000,000
	----------cm of residuum----------			
10	1	10	100	200
30	3	30	300	600
100	10	100	1,000	2,000

Another interesting aspect of the genesis of these soils is a landscape overturning model of the genesis of limestone derived soils proposed by Simpson (1979). This model was developed during a research project in Nittany Valley at the waste-water irrigation site 2 miles north of the Penn State Campus. In this model (Fig. 4), a sink hole forms (stage 2) which fills with material (stages 3 and 4) and then other sink holes form in the area that was the source area of the sediment for the filling in stage 1 (stage 6). Next the sediment from the filling of stages 3, 4, and 5 is eroded into the new sink holes. This model could be viewed as a group of pistons that are being lowered into the earth's crust.

These data and observations have interesting significance in the soil-geomorphology of the valley floor. The valley floor has significant relief. The main limestone uplands are at an elevation of about 1,200 feet, and the stream bottom are at 950 to 1,000 feet. Thus, there is 200-250 feet of relief on the limestone areas of the valley floor. Observations of excavations give varying results of depth to bedrock. This is probably a result of varying lithogies of the limestone, but in general, one trend does seem to be evident. This is that the relatively flat upland area has deeper soils (deeper to bedrock) than are found on the slopes. In addition, although this is a karst region, there are in some areas an integrated drainage network with drainage ways working headward into surrounding upland areas. These headward working drainage ways are most evident close to Spring Creek (the major stream draining of the valley). Thus, Nittany Valley is a fluviokarst area (karst landforms superimposed on a fluvial landscape; White and White, 1979). The karst areas of the valley underlain by dolomitic limestone show more subdued landforms compared to karst areas underlain by calcitic limestone. Although this is the typical situation, according to White (1984) the denudational rates for both types of carbonate rock is about the same.

Soils from Dolomite and Sandstone

The last group of residual soils to be discussed are those formed from interbedded dolomite and sandstone. These soils are the Vanderlip, Morrison, and Gatesburg. All three of these soils are sandy and have developed in a deep residual accumulation of sandy material that has weathered from the parent

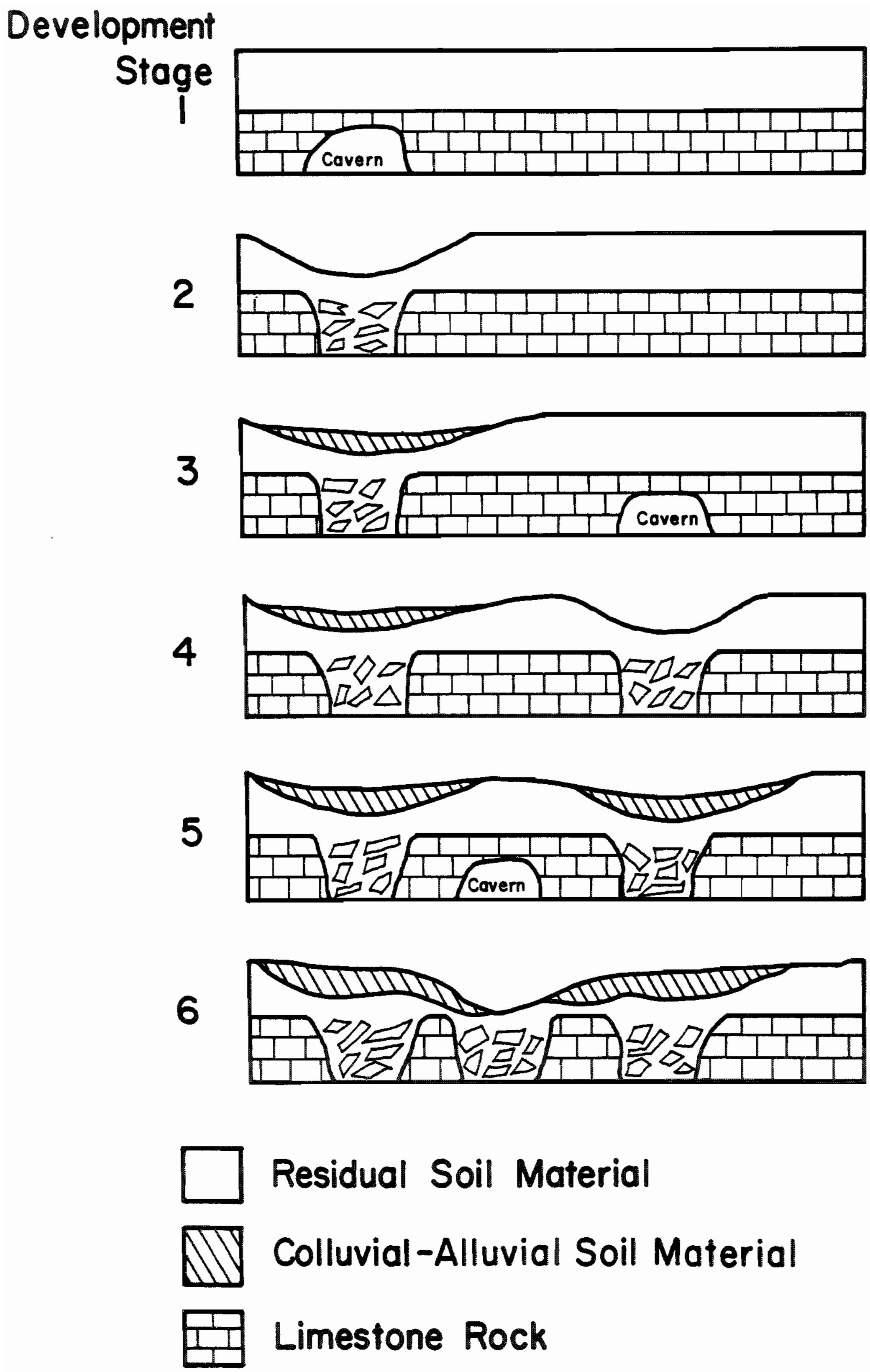


Fig. 4. Soil developmental sequence for soils developed from limestone in Nittany Valley (from Simpson, 1979).

rock. The Vanderlip is classified as a Psamment while Morrison is classified as a Udalf. The parent rock of these soils is an interbedded dolomite and sandstone (Gatesburg formation). As the bedrock weathered, the dolomite beds dissolved leaving the insoluble materials as layers of fine textured material in a very thick, sandy regolith. Although the sandy material is most evident, Butts and More (1936) state that the sandstone probably makes up only about 20% of the bedrock formation. Butts and More also state that some of the sandstone beds are as much as 10 feet thick. Where the thicker sandstone residuum intercepts the surface, the Vanderlip soils are found. Apparently the sandstone residuum did not have enough clay in it (Vanderlip soils have 4-8% clay) to be illuviated to form an argillic horizon, but where the sandstone residuum was thinner, some of the finer material has been eluviated from the dolomite residuum and has accumulated as an argillic horizon. In these places, Morrison soils are found. In the Vanderlip and below the argillic B horizon in the Morrison, lamellae (thin bands of slightly higher clay and iron oxide content) are frequently found. They have not been studied in these soils but their origin in other sandy soils has been attributed to (1) clay flocculation by free Fe oxides, (2) clay flocculation by carbonates, (3) periodic chemical precipitation of Fe with subsequent flocculation of the clay, (4) evapotranspiration at the wetting front, or (5) by sieving by a layer having fine pores (Bond, 1986). In Centre county, there are about 300 acres of Vanderlip and 28,000 acres of Morrison soils. On a state wide basis, there are about 87,000 acres of Morrison and 15,000 acres of Vanderlip (Table 5). The Gatesburg soil is found in the same area as the Morrison and Vanderlip soils. The Gatesburg is also sandy, but it has the morphology of a Spodosol. This means it has a B horizon of Fe, Al, and humus accumulation. The Gatesburg soil is unique in that it is the only soil with Spodosol morphology that is found on valley floors at low elevations in Pennsylvania. It is of limited extent (400 acres) and all of it is located just outside of State College. Other soils with Spodosol morphology (Leetonia) are found only at higher elevations in Pennsylvania, and like the Gatesburg, they are developed in sandy parent materials.

Well drained Spodosols form in coarse textured material at high latitudes and are usually associated with coniferous vegetation (McKeague et al., 1978; Soil Survey Staff, 1975). Stanley and Ciolkosz (1981) as a part of a study of Spodosols in West Virginia and Pennsylvania studied the Gatesburg soil. At their study area in the Barrens north of State College just off U.S. 322 two pedons were sampled; one with Spodosol morphology (E, Bhs, Bs), and one without Spodosol morphology (E, Bs). The data in Table 8 indicate that neither of these pedons (14-2, 14-3) met the chemical criteria of Soil Taxonomy (Soil Survey Staff, 1975) for a Spodosol nor the criteria for the Podzol of the Canadian Classification System (Canada Soil Survey Committee, 1978). Apparently, the Gatesburg soil is at a critical threshold in which the soil forming processes which form the Spodic horizon (B horizon of accumulation of humus, Al, and Fe) do not function well. Although McKeague et al. (1978) indicate that vegetation is only a partial factor in Spodosol formation, the very small acreage of Gatesburg soil in the Barrens area indicates that the white and pitch pine which were part of the vegetation of the Barrens were probably the natural vegetation of the Gatesburg soil. Although the pine formed an acid litter necessary for Spodosol formation (McKeague, 1978), the temperature was apparently high enough that most of the humus in the surface horizon decomposed in place and was not eluviated or the humus that was eluviated had a tendency to decompose fairly rapidly. The trend of decreasing humus and iron oxides content with increasing soil temperature noted by Stanley and Ciolkosz (1981) tends to support this conclusion (Fig. 5).

Table 8. Chemical criteria data for spodic horizon identification of Gatesburg pedons 14-2 and 14-3 (from Stanley, 1979).

Horizon	Depth cm	Percent clay	Pyro Fe + Al / Clay	Pyro Fe + Al / D-C Fe + Al	Index of Accumulation
Pedon 14-2					
A	0-5		0.01	0.47	---
E	5-36	4.0	----	0.10	---
Bhs	36-38	7.7	0.03	0.26	7
Bs	38-51	7.0	0.02	0.39	9
Bw1	51-71	6.2	0.02	0.21	52
Bw2	71-91	4.9	0.01	0.18	41
Bw3	91-122	3.6	0.01	0.36	37
Bt	122-132	17.8	0.01	0.04	83
Pedon 14-3					
E	0-20	6.6	0.01	0.42	---
Bs	20-30	8.5	0.02	0.06	15
Bw1	30-46	7.3	0.02	0.51	6
Bw2	46-63	10.9	0.01	0.33	13
Bt1	63-89	14.8	0.01	0.21	120
Bt2	89-109	19.3	0.01	0.04	120
C	109+	7.9	0.01	0.06	---

Another interesting aspect of Spodosol genesis is the undulating spodic horizon. It is believed that the undulating surface of the spodic horizon is a reflection of the vegetation of an area and its affect on the movement of water through the soil. These undulations were studied in three dimensions by Stanley (1979) at the State College Gatesburg site. Stanley concluded that there were two types of patterns (high centers and low centers), and that these patterns were a reflection of water draining down into the soil from trees with tap root (low centers) and fibrous root (high centers) systems (Fig. 6). Stanley's conclusion is partially supported by the work of Crampton (1982) and Ryan and McGarity (1983), but in conflict with the conclusion of Lag (1951). Lag concluded the undulations were a reflection of the surface topography which concentrated water in low spots, which in turn promoted deeper leaching and thicker E horizons development under the low spots.

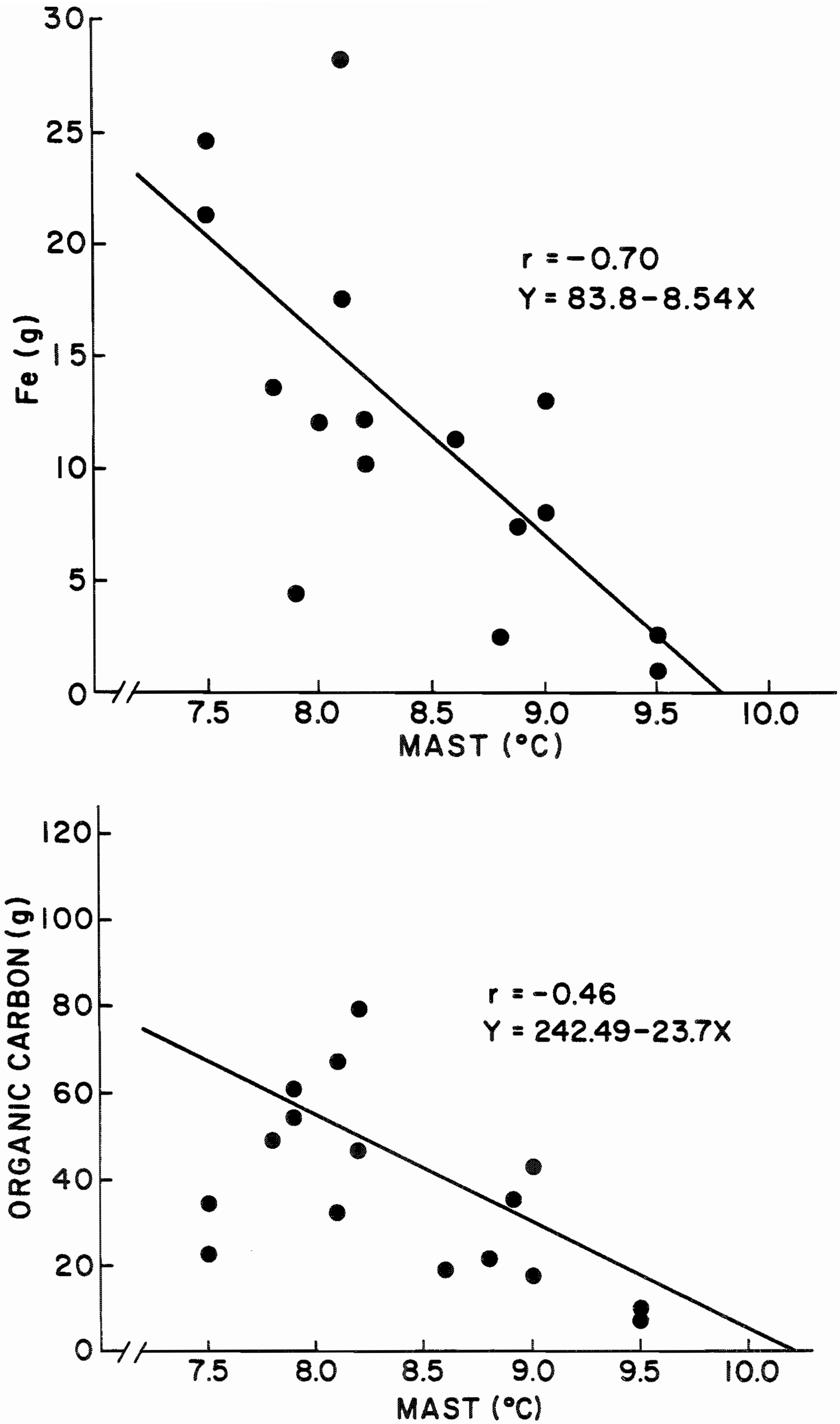


Fig. 5. Total accumulated (includes all subhorizons of the B) of organic carbon (humus) and iron (Fe) in Pennsylvania Spodosols (Stanley and Ciolkosz, 1981).

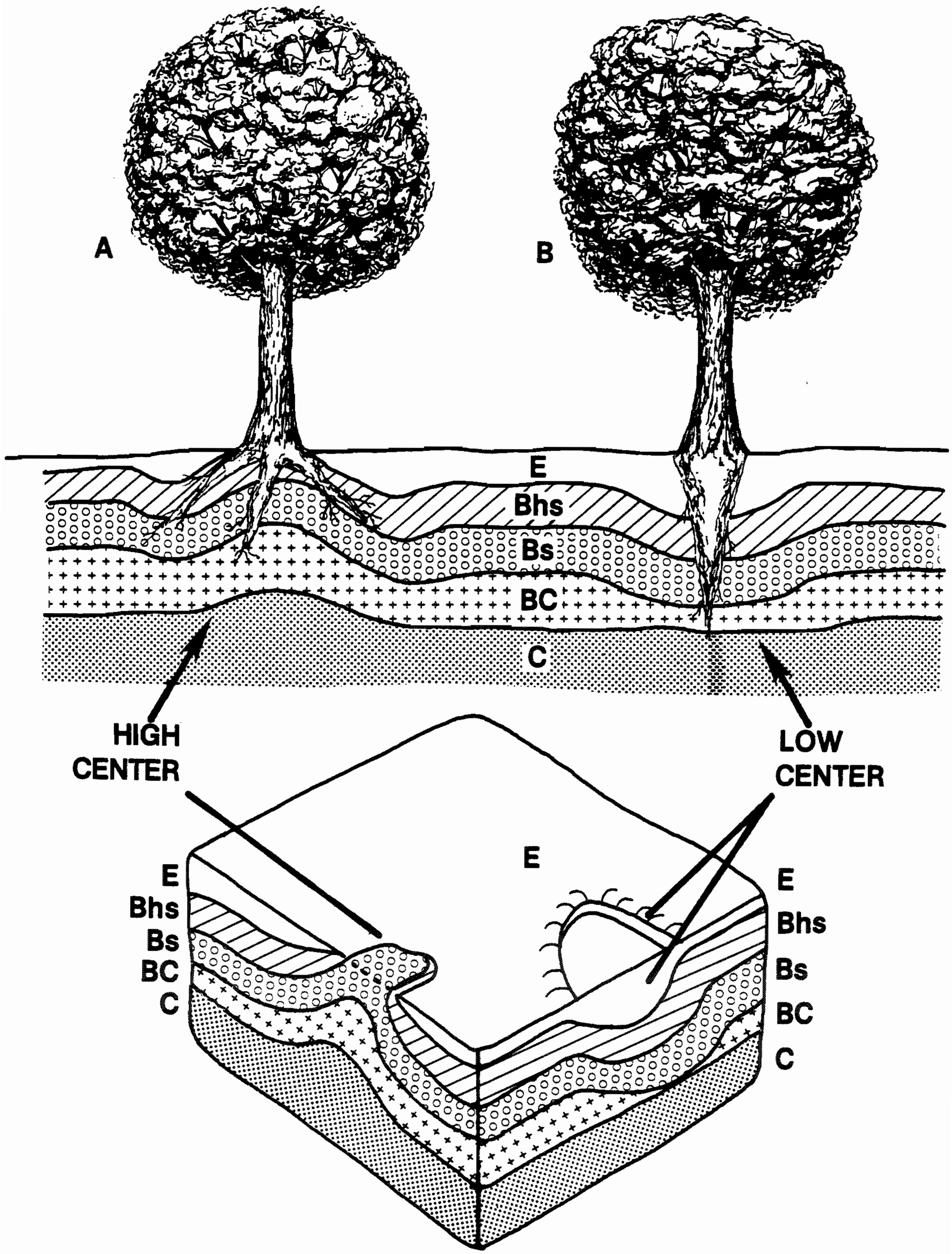


Fig. 6. Diagramic presentation of high centers (A) and low centers (B) in Spodosols (modified from Stanley, 1979).

The sandy soils of the Barrens contributed greatly to the initial settlement of central Pennsylvania. Although it did not produce good crops; hence the name, Barrens, it did have deposits of iron ore which was used by the iron furnaces in this region to make pig iron. According to Butts and Moore (1936), the ore is a hydrous oxide of iron which is a weathering product of the interbedded dolomite and sandstone. The ore occurs as concretionary or irregular masses mixed with clay, chert, and sand. The ore is relatively rich in iron, having about 40-50% iron (Butts and More, 1936). The last ore was mined in this area in 1909 and many of the buildings that made up the main mining town of Scotia were sold in 1911 and carted away. The railroad which served Scotia was taken up in 1924 and today the only thing that remains of Scotia or the mining era are many holes in the ground from which the iron ore was taken. These holes are found both near Scotia and throughout the Barrens.

Soils Developed in Alluvium

The only other soils of any acreage on the valley floor in this area are the Nolin, Lindside, and Melvin. These soils are developed in recent alluvium and are found on floodplains. They have formed in the silty alluvium derived primarily from the limestone upland areas. These soils are relatively young probably 200-300 years old (Bilzi and Ciolkosz, 1977). The age of the soils on the floodplain areas in Nittany Valley are probably related in part to man's activity in the surrounding watersheds. For example, in Nittany Valley, the extensive logging for charcoal during the early 1800's probably accelerated erosion in the Spring Creek watershed. The sediment from this erosion in turn buried the soil forming on the floodplain with 2 to 3 feet of material. Subsequent better management of the land in this area has apparently decreased the erosion and sedimentation. This sequence of events is apparent on the floodplain of Spring Creek. On this floodplain, there is a buried soil 2 to 3 feet below the surface. Above the buried soil, there is the Nolin soil with an ochric epipedon and a cambic B horizon (an Inceptisol soil). The presence of the Nolin in this relatively young material also indicates that a cambic horizon and an ochric epipedon can form relatively rapidly in floodplain sediments (Bilzi and Ciolkosz, 1977). Data for the Nolin pedon studied by Bilzi and Ciolkosz are given in Table 9.

The streams of Nittany Valley, as well as the large rivers in Pennsylvania such as the Allegheny, Juniata, Susquehanna, and Delaware, have relatively narrow floodplains. This feature plus the large amount of sloping land (Table 10) which promotes rapid runoff contributes to the frequent (deep) flooding experienced in Pennsylvania.

Soils Developed in Colluvium

The valley side slopes of Nittany Valley, as well as all of the major ridges in the Ridge and Valley area, have the lower one-half to three-fourths of their slopes mantled with colluvium. The colluvium ranges from less than one foot to more than 100 feet in thickness, and it forms simple side slope as well as more complex fan deposits. The simple side slope deposits extend on the average one-half mile from the ridge crests while the fan deposits (adjacent to gaps in the ridges) commonly extend one-quarter to one-half mile beyond the simple slope deposits. The colluvial material extends downslope until the slope become very gentle or until a secondary ridge or stream is

Table 9. Selected soil characterization data for Nolin silt loam (from Bilzi and Ciolkosz, 1977).

Horizon	Depth cm	Sand (2.0-5 mm)	Silt (0.5-0.002 mm)	Total Clay, (<0.002 mm)	Fine Clay, (<0.0002 mm)	pH	Base Saturation	Free Iron Oxides (Fe_2O_3)	Organic Carbon
		%					%		
A	0-8	22.3	58.3	19.4	10.5	7.1	72.9	2.0	2.76
Bw1	8-25	17.7	59.9	22.4	13.3	7.4	75.4	2.3	1.23
Bw2	25-40	28.8	53.6	18.6	9.9	7.6	82.3	2.2	1.59
BC	40-59	44.1	34.6	21.3	13.2	7.7	83.3	1.9	1.82
Ab	59-78	36.1	47.1	16.8		7.7	82.9	1.6	2.18
Bw1b	78-90	50.6	31.1	18.3		7.8	81.0	1.4	1.71
Bw2b	90-99	54.4	28.8	16.8		7.8	80.5	1.5	1.91
2C	99-120	63.1	21.9	15.0		7.8	83.1	2.2	1.34

Table 10. Percentage of land in various slope classes in Pennsylvania (Cunningham and Day, 1986).

Area and County	Slope Classes (Percent) 0-3	3-8	8-15	15-25	> 25
	%				
Southeast					
Bucks	38	44	10	6	2
Lancaster	24	39	23	7	5
Ridge and Valley					
Fulton	5	19	22	26	28
Huntingdon	6	17	17	29	31
Centre	7	35	13	21	25
Unglaciated Plateau					
Armstrong	6	22	23	16	33
Indiana	13	20	22	24	21
Greene	7	8	9	26	50
Clinton	7	24	5	23	41
Glaciated Plateau					
Crawford	42	41	10	5	2
Mercer	35	50	10	3	2
Pennsylvania	12	33	17	20	18

encountered. In general, there are two kinds of colluvium. The first kind was probably deposited only by mass movement (probably solifluction). This type of colluvium has many rock fragments in it. The second type might be better classed as alluvium. It is usually silty and has only a few pea size rounded pebbles in it. This second type of material is found in close proximity to the stony colluvium in drainage ways and low areas which extend beyond the stony colluvium into the valley. Limited data indicate this material can be at least 10 feet thick. Apparently, as the stony colluvium was moving down slope, a considerable amount of material was carried by very fluid mud flows or by running water and was deposited in the lower areas adjacent to the footslopes. The deposition of these materials had the effect of reducing the relief adjacent to the footslopes creating a more gently rolling landscape than was there prior to the movement and deposition of these materials.

The occurrence of colluvium is not restricted to the Ridge and Valley area; it is also extensive in the Appalachian Plateau area. The data in Table 4 indicate that in the Ridge and Valley area a typical county has about 27% of its area covered by colluvium; while on the Plateau, colluvium occupies an area of about 13% of a typical county. These data indicate the extensive nature of this material in Pennsylvania.

The genesis of the soils developed in these colluvial deposits is discussed by Ciolkosz et al. (1979). Data from a Buchanan and Andover soil, from their publication, is presented in Table 11. In this publication, the authors indicate that these soils may have textural changes with depth. These changes are a reflection of both textural variation in the parent material and argillic horizon development. They also indicate that these soils have fragipans in them if they are not too clayey or if they do not have much limestone influence (Murrill soils) in the parent material. The authors concluded from an evaluation of the properties of these soils and the degree of weathering of their clay minerals that they show only a moderate degree of soil development. The colluvial material on these slopes does not appear to be moving down slope today. Little, if any deformation of trees, is noted and the argillic horizons and fragipans in these soils indicate landscape stability. The colluvium may well be in a "super stable" condition. Under periglacial conditions, the angle of repose of the material was lower than under today's conditions. This would indicate that these slopes are at a lower angle of repose for current conditions making the slopes "super stable" to natural downslope movement.

Thus, if these slopes are currently stable, when did the colluviation take place? Early workers attributed the movement and deposition to periglacial activity associated with Wisconsinan glaciation (Denny, 1956; Peltier, 1949). To the authors' knowledge, the only direct evidence of the age of the colluvium comes from some radiocarbon dates in West Virginia (Ciolkosz et al., 1990), which indicate a late Wisconsinan age (15,000-20,000 yrs B.P.). Another line of evidence pointing to the late Wisconsinan is that during this time the area had a periglacial climate with permafrost. Under this type of climate, mass movement of material downslope and its accumulation on the lower side slopes occurs. Support for a periglacial climate comes from pollen analyses (Watts, 1979; Maxwell and Davis, 1972; Martin, 1958) and the presence of other periglacial features such as patterned ground (Walters, 1978; Clark, 1968; Clark and Ciolkosz, 1988); and boulder fields, grezes litees, and involutions (Ciolkosz, 1978; Ciolkosz et al., 1986); and pingos (Marsh, 1987) in New Jersey and Pennsylvania. The similarity in soil development between the soils of the colluvial deposits and Wisconsinan glacial till deposits also indicates Wisconsinan periglacial movement and deposition (Ciolkosz, 1978).

The chronology of deposition of the colluvium is not known absolutely. The West Virginia radiocarbon dates indicate late Wisconsinan deposition of the colluvium. The colluvium through Pennsylvania seems to be of a similar age. This is indicated by similar soil development in the same kind (same lithology) of colluvium throughout the state. The soils developed in the colluvium are similar in some aspects to late Wisconsinan (Woodfordian) age soils developed in glacial till in that they have fragipans, but in addition to fragipans, they have argillic horizons. This probably indicates the colluvium was more weathered than the till, and it had more fine clay which could be moved to form an argillic horizon. A speculative chronology would start with the early Woodfordian, and as the ice moved forward, a periglacial climate with tundra vegetation triggered solifluction movement down slope. As the climate warmed and the ice retreated, soil formation progressed on the stabilized slopes.

Table 11. Selected soil characterization data for a Buchanan and Andover soils (from Ciolkosz et al., 1979).

Soil and Horizons	Depth cm	Bulk density g/cc	IWP* ------ illite	pH	Base Saturation	Rock Fragments >2 mm	Sand 2-.05 mm	Silt .05-.002 mm	Clay <.002 mm	Textural Class
					%					
Buchanan (39-41)**										
Ap	0-28	1.40	2.84	5.6	22	17	36.0	49.1	14.9	l
Bt1	28-40	1.53	1.89	5.9	33	32	36.8	44.8	18.4	l
Bt2	40-50	1.60	0.71	5.7	34	34	32.2	39.7	28.1	cl
Bx1	50-75	1.56	0.46	5.2	17	13	32.7	37.9	29.4	cl
Bx2	75-95	1.64	----	5.2	10	13	33.8	37.0	29.2	cl
Bx3	95-113	1.67	0.62	5.2	12	14	31.3	37.2	31.5	cl
Bx4	113-135	1.67	----	5.2	9	15	31.2	38.6	30.2	cl
Bx5	135-160	1.68	0.60	5.2	9	14	31.6	38.3	30.1	cl
Bx6	160-188	1.67	0.54	5.2	8	29	35.7	36.3	28.0	cl
Andover (18-13)**										
Ap	0-23	1.24	1.59	6.0	56	8	29.6	46.6	23.8	l
E	23-35	1.48	----	6.2	57	36	34.4	46.4	19.2	l
Bt1	35-45	1.61	0.31	6.1	58	26	33.5	42.4	24.1	l
Bt2	45-68	----	----	5.4	53	34	31.7	40.4	27.9	cl
Bx1	68-80	----	0.30	5.1	37	28	28.6	46.6	24.8	l
Bx2	80-95	---	----	5.1	39	26	41.0	39.5	19.5	l

*IWP = Illite Weathering Products to Illite ratio (see Ciolkosz et al., 1979).
**Pennsylvania State University Soil Characterization Lab. Number.

Recent work by Hoover (1983) and Hoover and Ciolkosz (1988) indicates that the colluvial deposits are more complex than previously reported. These studies indicate that the slopes in the Ridge and Valley area have older red colluvium buried by a brown Wisconsinan age surface colluvium. Limited data indicates that the older colluvium is thicker than the more recent material. The red color of the buried colluvium is due to oxidation associated with soil formation when the buried colluvium was at the surface. Thus, the red material is a buried paleosol. The age of the buried paleosol is not known, although it is thought to be pre-Wisconsinan (about 125,000 yrs B.P.) (Ciolkosz et al., 1990b). Where the Wisconsinan colluvium is thin, present soil forming processes extend into the older material welding the present surface soil to the buried paleosol.

Although Denny (1956) states that periglacial activity decreased rapidly with distance from the Wisconsinan ice margin, the presence of colluvium on all the major ridge slopes in Pennsylvania indicates significant periglacial mass movement a considerable distance south of the Wisconsinan terminal moraine. This has particular significance when related to the work of Clark (1968), who has found patterned ground sites, presumably associated with periglacial activities, at high elevations in the Appalachians as far south as southern Virginia and Walters (1978) who has found patterned ground in southern New Jersey. Thus, soils developed from colluvium similar to those in Pennsylvania extend as far south in the Appalachians as Virginia and Tennessee. Although not studied in detail, soil scientists in Virginia indicate that the present mesic-thermic soil temperature boundary (Washington, DC west to the Blue Ridge and then south along the Blue Ridge) may separate the zone of periglacial erosion to the north from less altered areas to the south. This conclusion is based on presence of thin soils north of the line and thick, well developed soils south of the line on the same landforms and lithologies.

Concluding Remarks

The discussion given above on the soils of the Nittany Valley area indicates that parent material, time, and topography (erosion) are locally the most important soil forming factors. Most of the soil surfaces we see are Wisconsinan in age (< 75,000 years old). Soils older than Wisconsinan age are found but their genesis is complex. For example, the buried colluvium on the ridge sideslopes has a pre-Wisconsinan soil developed in it. This soil may be buried so deeply that present day soil forming processes do not affect it greatly, or it may be near enough to the surface that the present processes are affecting its development and the two soils are being welded together. Only on the stablest landscape surfaces have pre-Wisconsinan soils been preserved, and even on these surfaces, the upper part of many of the soils have been modified by cryoturbation and the addition of aeolian dust. The major erosion and turbation that has affected the soils of the area is associated with periglacial climates of the Wisconsinan glacial time and to a certain extent the pre-Wisconsinan. Because these soils are relatively young, the effect of parent material is greater than if these soils were very old. For example, a Berks soil is brownish-gray in color and has a loamy texture with many shale chip rock fragments, while a Murrill is yellowish-brown, silty with numerous sandstone rock fragments. With time, both of these soils will develop towards a soil with a red color, no rock fragments, and a clayey texture.

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CHAPTER 2

Historic Habitation of Nittany Valley*

by

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Introduction

The first permanent white settlement in Pennsylvania was established in 1643 by Johan Printz, the Swedish governor of New Sweden. The settlement was called New Gottenburg and was located on Tinicum Island just below the present site of Philadelphia. The settlement prospered and the Dutch grew jealous of the colony and seized it in 1655. But in turn the Dutch were ousted in 1664 and their holdings in Pennsylvania were claimed by the English.

Thus, much of the early history of Pennsylvania is closely associated with the English, in particular the Englishman William Penn. Penn was born in 1644 in England the son of Admiral Penn, a favorite of the reigning monarch of England, King Charles II. Young William Penn did not endear himself to his father or the King when he espoused the cause of the prosecuted and unpopular Quaker religious sect. For his ardent support of the Quakers William was expelled from Oxford, and later imprisoned and tried by the courts for suspected treason. In spite of this the outspoken Penn was forgiven by his father, and ultimately became a powerful leader of the Quaker movement in England. Upon the death of Admiral Penn, King Charles II settled a debt of 16,000 pounds which he owed him by granting to William, in 1681, a 35,361,000 acre (55,251 sq. miles) tract in America (present area of Pennsylvania is 29,013,120 acres; 45,333 sq. miles). This tract was named Pennsylvania (Penn's Woods) in honor of William Penn's father, and as soon as the charter was signed by Charles II, Penn organized his fellow Quakers for a mass movement to Pennsylvania. The word of Penn's democratic form of self government spread in Europe and other oppressed peoples and religious sects also began moving to Pennsylvania, including many groups from Germany, who later were called the Pennsylvania Dutch.

Most of the early settlers who came to Pennsylvania came to Philadelphia and then moved westward. Thus the settlement pattern for Pennsylvania was from the southeast to the west and northwest. The settlement of Pennsylvania was slow at first but by 1730 the population began to explode and by 1750 all of southeastern (SE of the first ridge of the Ridge and Valley Area) Pennsylvania

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was settled. The first and following ridges of the Ridge and Valley Area formed formidable obstacles to the settlers and much of the westward migration was deflected southwestward down the Great Valley (Cumberland or Shenandoah Valley as it is known further south) to Virginia and the Carolinas. The ridges were such an obstacle that settlement was slow but by 1800 the Ridge and Valley Area was settled but sparcely populated. After 1800 the Ridge and Valley filled rapidly and by 1840 the valleys had been cleared of woodland and the Ridge and Valley area took on its characteristics banded appearance of agricultural valleys and wooded ridges.

After 1800 the true settlement frontier passed westward out of Pennsylvania. Those remaining unsettled areas were marked by poor soil, rugged terrain and remote location. These areas were gradually filled by farmers lacking the desire or funds to move west to more fertile, unsettled areas, by mineral developers, and finally by those hoping to develop the summer recreational aspects of the areas. Much of this land has, in fact, never been permanently settled, and is part of the 62% of Pennsylvania that is wood land today.

Nittany Valley

James Potter was the first white man to explore Nittany Valley. Potter was a captain in the colonial militia stationed at Fort Augusta (Sunbury), and in 1759 he traveled up the west branch of the Susquehanna River to the mouth of Bald Eagle Creek and up Bald Eagle Creek to the mouth of Spring Creek. From there he traveled overland across Nittany Valley to the top of Nittany Mountain. Potter seeing the prairies and noble forest beneath him in Penns Valley cried to his attendant "By heavens Thompson, I have discovered an Empire."

Although first explored in 1759 Nittany and Penns Valleys were not settled rapidly. One of the main reasons was Indian problems associated with the Revolutionary War. During the war the British enlisted the help of the Indians to raid the Frontier settlements. The Indian raids were so common that by 1778 most of the settlers had left and returned eastward in what has been called the "Great Run-Away." Cornwallis surrendered in Yorktown in 1781 ending the Revolutionary War, but it was not until 1783 that the settlers began in earnest to return.

Much of the early development of Nittany Valley and Central Pennsylvania is associated with iron making. In 1784 high quality iron ore was discovered by surveyers in Nittany Valley in an area called the Barrens. This discovery lead Col. Samuel Miles and Col. John Patten to buy 8,000 acres of land (later called

Centre Furnace lands) in the valley. In 1789 Patten moved to the valley, erected Centre Furnace and put it into blast in 1791. Centre Furnace was not the first operating blast furnace in the area, the first was the Bedford Furnace in Huntingdon County near the town of Orbisonia. It was put into blast in 1785. Centre Furnace became a well known place in Central Pennsylvania and may have suggested the name "Centre" for the new county that was formed in this area in 1800 with a population of 2,075 (1980 population est. 113,494). Centre Furnace operated from 1791 to 1809 and from 1826 to 1858. In the early 1800's there was about 50 iron furnaces in operation in the Juniata Iron District (Blair, Centre, Mifflin and Huntingdon Counties). In Centre county there were 10-15 in operation and almost all of them were in a 15 mile radius of Centre Furnace. These furnaces produced about 12,000 tons of pig iron annually in the 1830's. Most of these furnaces produced about 1,000 tons annually. By 1858 competition from anthracite blast furnaces and the exhaustion of charcoal producing timber forced the final close of Centre Furnace. Most of the other furnaces in the district also closed about this time and by the late 1870's the era of Juniata iron was over in central Pennsylvania.

The effect of making charcoal for the iron furnaces was very significant in the settlement history of Nittany Valley and much of Central Pennsylvania, the reason being the large amount of wood needed for the charcoal. If we assume that an average white oak forest stand would yield 70 cords of wood (7 cords wood = 1 ton of metal), that would indicate that a furnace like Centre Furnace would use the wood from about 100 acres/year. Thus during its 50 year history Centre Furnace (about 1,000 tons of metal per year) used the wood from about 5,000 acres. This value extended to the 10-15 furnaces in this area would equal about 50,000 acres of forest land that was used for charcoal during the Juniata iron era in Nittany Valley. These estimates are less than others which indicate that a furnace like Centre Furnace would use the charcoal from one acre of forest for each day's production. This would indicate a consumption of about 300 acres/year, and 15,000 acres over its 50 year life. Regardless of which estimates are the best they do indicate the very large consumption of forest for charcoal in Central Pennsylvania during the Juniata iron era.

Although the iron industry greatly shaped the initial history of Nittany Valley it is a part of the past history. Although a part of past history Centre Furnace greatly influenced future development of the valley. In 1855 the Pennsylvania legislature created the Farmers High School; however there was no designed location for the school. Different areas in the state vied for the school but it was an offer of 200 acres of land and an option to buy 200 more acres from Moses Thompson and James Irvin (owners of Centre Furnace) and a guaranteed $10,000 pledge from Irvin and other citizens of Centre county that was accepted and responsible for the location of the school on Centre Furnace lands. Although established in 1855 the school first opened on February 16, 1859 with 69 students present. The school was renamed the Agricultural College of Pennsylvania in 1862 and the Pennsylvania State College in 1874. The last name change occurred in 1953 when the Pennsylvania State College was renamed The Pennsylvania State University.

In addition to iron making and the location of the Pennsylvania State University Nittany Valley's history and appearance has been shaped appreciably by agriculture. The Nittany Valley today with the exception of the urbanization around State College (75% of the population of Centre County is located within a 10 miles radius of State College) is much like is was in the mid-1800's. The Valley is cleared for agriculture while the barrens and the ridges are wooded. The barrens and ridges have never been cleared for farming although they have been logged and burned. The amount of agricultural land in Nittany Valley has been somewhat constant since the late 1800's. This stability is somewhat different than the State wide trend. In 1840 there were 14 million acres (about half of the total State's area) in farms in Pennsylvania. In 1900 there were 19 million acres, in 1940, 14 million and in 1980, 9 million acres. Most of the post 1900 decrease was due to the abandonment of rough or infertile marginal land primarily on the Appalachian plateau. More recent losses are the result of urbanization. Today Pennsylvania has about 3.5 million acres (about 12% of State's area) urbanized, with a good share of the urbanization in the southeastern and southwestern parts of the state. The southeastern part of the state and Nittany and a few other valleys in the central part of the state have almost all of the class I and II agricultural land in Pennsylvania. The bulk of the abandoned lands in Pennsylvania have reverted to forest land. In the late 1800's when indescriminate cutting and burning and the settling of land peaked, Pennsylvania had about 30% of its land in forests. Today this figure has increased to about 62% (26% farming and 12% urban; Fig. 2.1). The large area of forest land has fostered outstanding hunting conditions. Today Pennsylvania sells more hunting licences (about a million) than any other state and in 1979 114,000 deer were taken by hunters. The abundance of deer also causes problems. In 1979, 25,000 deer were killed on highways by cars and trucks. This is more deer than taken by hunters in 35 other states.

The agriculture in Nittany Valley during the early 1800's was largely wheat and livestock. With time wheat was replaced by corn and the livestock enterprise has centered on dairy cattle. Thus Nittany Valley today is primarily a corn and diary farming area.

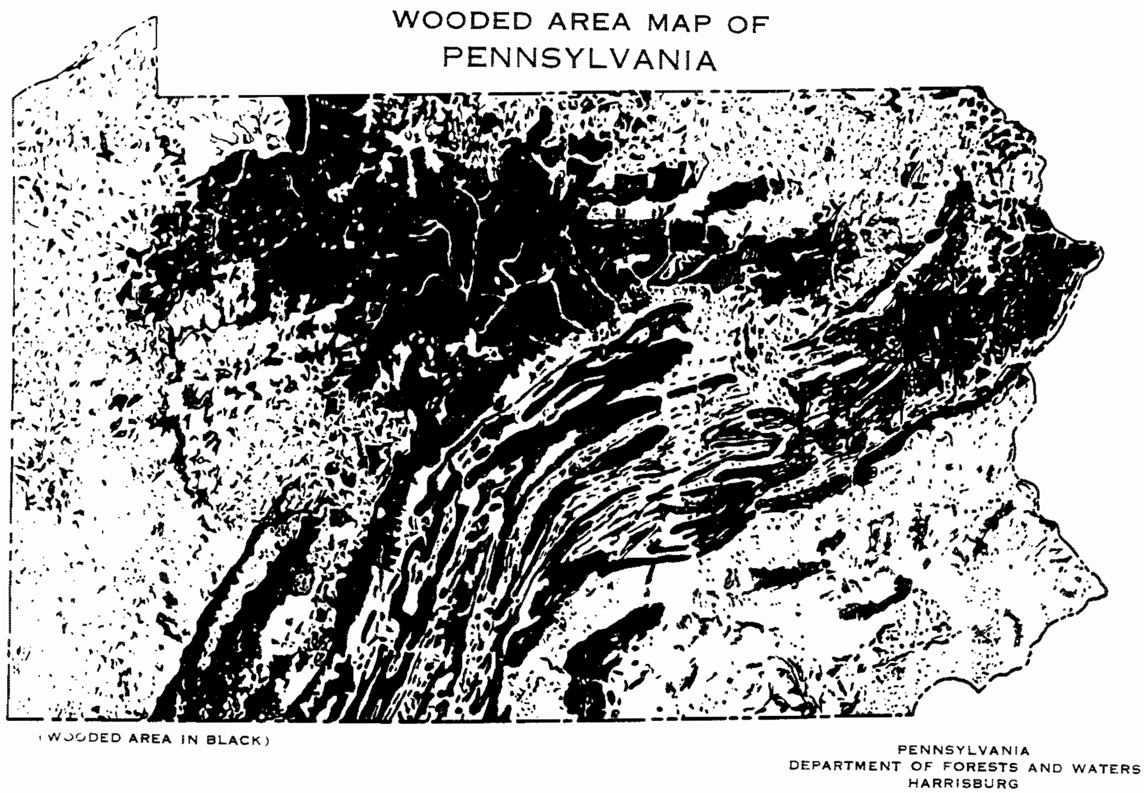

Fig. 2.1 Wooded area (in black) map of Pennsylvania.

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